DYNAMIC LABORATORY
SIMULATIONS OF FLUID-ROCK
COUPLING WITH APPLICATION
TO VOLCANO SEISMICITY AND
UNREST.

Marco Fazio

School of Earth and Environmental Sciences, University of Portsmouth

The thesis is submitted in partial fulfilment of the requirements for the award of the degree of
Doctor of Philosophy of the University of Portsmouth

January, 2017
DECLARATION

Whilst registered as a candidate for the above degree, I have not been registered for any other research award. The results and conclusions embodied in this thesis are the work of the named candidate and have not been submitted for any other academic award.

Words count: 63040
ACKNOWLEDGEMENTS

First of all, I would like to thank my first supervisor, Dr. Philip Benson, who gave me the opportunity to do this PhD project. He believed in me from the first time we met, always supporting and mentoring me. He knew that I would have done something different from my previous studies, therefore he helped me in any way possible. “From your mistakes, you are learning as well” he said to me. Instead of letting me down, he cheered me up. That was very helpful.

I would like to express my gratitude to Dr. Sergio Vinciguerra and Prof. Philip Meredith, second and third supervisor respectively. Through their guidance and their experience I learnt a lot and took benefit from their knowledge.

I thank my colleagues at the School of Earth and Environmental Sciences, Kostas and Orla, and, particularly my labmates at Rock Mechanics Laboratory (RML), Stephan, Pete and Emily who constantly helped me in progressing with my experiments.

A special mention for the external visitors at RML, Richard and Angela. With their projects, they increased my curiosity in Rock Mechanics. With their friendships, I found the perfect mates to enjoy talking and cycling.

In Portsmouth I found a family made of a lot of friends, who I thank for these 4 years. To start with, my housemates, Enzo and Massi, who saw me every day, both struggling and enjoying my PhD life. Without them, leaving abroad would have been more difficult. Several other people need a mention here. These are Serena (first person met) and Tony, Eduardo, Giulia, Maria and Luigi, Valentina, Gina, Brad, Maria and Fabio, Giuseppe, Lisa, Chiara and Daniele, Andrea, Giulia and David, Anna Maria. In Portsmouth I also found two amazing cycling mates: Francesco and Valerio. Their company made all those kilometres cycled together funnier and more enjoyable. It was a pleasure to meet all these people and they were paramount for my life in Portsmouth.

Finally, I owe this thesis to my family, my father, my mother, my siblings and my nephews. They always believed in me, supporting when it was hard and never letting me done.
DISSEMINATION

PEER REVIEWED JOURNAL PUBLICATIONS

CONFERENCE PRESENTATIONS
From Lab to Field: eruption forecasting using volcano-tectonic and low frequency seismicity (Talk). *EURO conference on Rock Physics and Geomechanics 2015*, Ambleside (UK), 6-11 September 2015

The role of pore fluids on deforming volcanic rocks: an experimental study (Poster). *EGU 2015*, Vienna (Austria), 12-17 April 2015

Laboratory simulations of fluid-induced seismicity in shallow volcanic faults (Poster). *EGU 2015*, Vienna (Austria), 12-17 April 2015

Laboratory simulations of fluid-induced seismicity in shallow volcanic faults (Talk). *AGU 2014*, San Francisco (USA), 14-15 December 2014
ABSTRACT

Pore fluids play a key role in how crustal rocks deform, particularly in a volcanic environment where fluids span a wide range of types, and exist across a wide spectrum of temperature, pressure, and phase, influenced by the presence of the magmatic system at depth. Not only do pressurized pore fluids affect the mechanical properties and the elastic velocities of the host rock mass (volcanic edifice), but they are also responsible in the generation of a range of seismic signals, characterized by Low Frequency and long coda as compared to the seismicity generated by simple shear, resulting in Volcano-Tectonic events.

While great progress has been made in understanding Volcano-Tectonic events, fluid-induced signals resulting in Low Frequency seismicity are not fully understood, and how these signals evolve from other signal types in time and space. To investigate, this study presents a series of rock triaxial deformation (in both wet and dry conditions) and fluid depressurization experiments, using a servo-controlled triaxial testing machine and state-of-the-art acoustic emission (AE) instrumentation. AE signals are the laboratory analogue of field-scale earthquakes, representing the key to understand the physics of the macro-scale events.

Considering shallow volcanic conditions (up to 1.6 km deep), this thesis shows that the presence of pore fluid delays the fracturing and the onset of microseismic activity, likely explaining sudden increase of precursory seismic activity before volcanic eruptions. Fluids also homogenize the rock material, decreasing the elastic wave anisotropy as they flow inside the newly formed cracks. In addition, the depressurization of fluids reveals how different fluid phases contributes to form different spectral peaks, characterizing the fluid-induced signals. Finally a fundamental microseismic event, (which presents a remarkable similarity with a natural volcanic earthquake, Tornillo), has been generated during gas depressurization, representing a new key link between earthquake features (such as amplitude modulation) and a physical properties (such as pressure drop).
# TABLE OF CONTENTS

DECLARATION..................................................................................................................2  
ACKNOWLEDGEMENTS..................................................................................................3  
DISSEMINATION..............................................................................................................4  
  PEER REVIEWED JOURNAL PUBLICATIONS...............................................................4  
  CONFERENCE PRESENTATIONS....................................................................................4  
ABSTRACT.........................................................................................................................5  
TABLE OF CONTENTS.....................................................................................................6  
LIST OF FIGURES ...........................................................................................................9  
LIST OF TABLES .............................................................................................................15  
LIST OF ACRONYMS AND SYMBOLS............................................................................16  

1. INTRODUCTION.........................................................................................................17  
  1.1. VOLCANO: A THREATENING NEIGHBOUR .......................................................17  
  1.2. VOLCANO TECTONICS AND SEISMOLOGY......................................................18  
  1.3. AIMS AND OBJECTIVES.....................................................................................19  
  1.4. THESIS OUTLINE ..............................................................................................19  
2. VOLCANO SEISMOLOGY..........................................................................................21  
  2.1. TERMINOLOGY....................................................................................................21  
  2.2. CHARACTERISTIC FEATURES..........................................................................24  
  2.3. POSSIBLE SOURCE MECHANISMS.................................................................30  
  2.4. MOMENT TENSOR SOLUTION...........................................................................36  
  2.5. ERUPTION FORECASTING..................................................................................38  
  2.6. WHAT IS MISSING FROM THIS PLETHORA OF STUDIES?..............................42  
3. LABORATORY EXPERIMENTS.................................................................................43  
  3.1. ROCK MECHANICS TESTING............................................................................43  
  3.2. ACOUSTIC EMISSIONS.....................................................................................44  
  3.3. EARLY RESEARCH USING AE AS A LABORATORY TOOL...............................45  
  3.4. SIMULATION OF VOLCANIC CONDITIONS IN LABORATORY.......................50  
    3.4.1. HYDROSTATIC TESTS ..................................................................................51  
    3.4.2. UNIAXIAL TESTS .......................................................................................51  
    3.4.3. TRIAXIAL TESTS .......................................................................................53  
  3.5. SCALING LAWS FOR AEs....................................................................................56  
4. EQUIPMENT AND METHODS..................................................................................59  
  4.1. TEST MATERIAL AND SAMPLES.......................................................................59  
  4.2. PRELIMINARY TESTS .......................................................................................61
6.3. ROCK-FIUID COUPLING DURING PORE PRESSURE RELEASE
6.3.1. REQUIREMENTS FOR THE ONSET OF THE LONG-DURATION AE ACTIVITY

5. RESULTS

5.1. DENSITY AND POROSITY

5.2. OVERVIEW OF THE EXPERIMENTS

5.3. DEFORMATION UNDER DRY CONDITIONS

5.3.1. ACTIVE EVENTS (ELASTIC-WAVE VELOCITY)

5.3.2. PASSIVE EVENTS (ACOUSTIC EMISSION)

5.4. DEFORMATION UNDER SATURATED CONDITIONS (p = 5 MPa)

5.4.1. ACTIVE EVENTS (ELASTIC-WAVE VELOCITY)

5.4.2. PASSIVE EVENTS (ACOUSTIC EMISSION)

5.5. DEFORMATION STAGE UNDER SATURATED CONDITIONS (p = 16 MPa)

5.5.1. ACTIVE EVENTS (ELASTIC-WAVE VELOCITY)

5.5.2. PASSIVE EVENTS (ACOUSTIC EMISSION)

5.6. RELEASE STAGE IN LOW TEMPERATURE CONDITIONS (WATER)

5.7. RELEASE STAGE IN HIGH TEMPERATURE CONDITIONS (WATER)

5.8. RELEASE STAGE IN LOW TEMPERATURE CONDITIONS (NITROGEN)

6. DISCUSSION

6.1. THE NATURE OF ROCK COUPLING DURING THE DEFORMATION STAGE

6.1.1. ON THE MECHANICAL PROPERTIES

6.1.2. ON THE ELASTIC WAVE PROPERTIES

6.1.3. ON THE AE ACTIVITY

6.2. EFFECT OF THE CONDUIT DURING THE DEFORMATION STAGE

6.2.1. ON THE MECHANICAL PROPERTIES

6.2.2. ON THE ELASTIC WAVE PROPERTIES

6.2.3. ON THE AE ACTIVITY

6.3. ROCK-FIUID COUPLING DURING PORE PRESSURE RELEASE

6.3.1. REQUIREMENTS FOR THE ONSET OF THE LONG-DURATION AE ACTIVITY
6.3.2. PARAMETERS AFFECTING THE LONG-DURATION AE ACTIVITY ........................................209
6.3.3. RELATIONSHIPS BETWEEN LONG- AND SHORT-DURATION AE SIGNALS ..............221
6.3.4. WAVEFORM SIMILARITY: SHORT-DURATION AE EVENTS .....................................229
6.3.5. SOURCE COMPONENTS OF THE SHORT-DURATION AE EVENTS ............................230
6.4. FROM THE LABORATORY TO THE FIELD .................................................................233
6.5. UNCERTAINTIES ..................................................................................................238

7. CONCLUSIONS ........................................................................................................241
SOLVING MACRO-SCALE PROBLEMS FROM MICRO-SCALE CASES ..........................243
FUTURE DIRECTIONS ..................................................................................................244

8. BIBLIOGRAPHY ........................................................................................................246

APPENDIX ..................................................................................................................258
APPENDIX 1: ETHIC REVIEW FORM ...........................................................................258
APPENDIX 2: MATLAB SCRIPTS .................................................................................263
MATLAB CODE 1 ........................................................................................................263
MATLAB CODE 2 ........................................................................................................265
MATLAB CODE 3 ........................................................................................................269
MATLAB CODE 4 ........................................................................................................273
MATLAB CODE 5 ........................................................................................................274
### LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1</td>
<td>VT signals recorded at Redoubt Volcano (from Lahr et al., 1994)</td>
<td>25</td>
</tr>
<tr>
<td>2.2</td>
<td>LP signal recorded at Redoubt Volcano (from Lahr et al., 1994)</td>
<td>26</td>
</tr>
<tr>
<td>2.3</td>
<td>tremor recorded at El Hierro Island (from Tarraga et al., 2014)</td>
<td>27</td>
</tr>
<tr>
<td>2.4</td>
<td>HB signal recorded at Redoubt Volcano (from Lahr et al., 1994)</td>
<td>28</td>
</tr>
<tr>
<td>2.5</td>
<td>Tornillo recorded at Vulcano Island (from Milluzzo et al., 2010)</td>
<td>29</td>
</tr>
<tr>
<td>2.6</td>
<td>VLP pulses recorded at Kilauea Volcano (from Ohminato et al., 1998)</td>
<td>30</td>
</tr>
<tr>
<td>2.7</td>
<td>Schematic model of a fluid-filled crack (from Chouet et al., 2003)</td>
<td>32</td>
</tr>
<tr>
<td>2.8</td>
<td>Schematic model of a rising plug of magma (from Neuberg et al., 2006)</td>
<td>34</td>
</tr>
<tr>
<td>2.9</td>
<td>Schematic model of a separated gas-liquid flow in a crack in the proximity of a nozzle (from Ohminato et al., 1998)</td>
<td>36</td>
</tr>
<tr>
<td>2.10</td>
<td>Representation of the source components (from Julian, 1998)</td>
<td>38</td>
</tr>
<tr>
<td>2.11</td>
<td>Rate and inverse rate against time for the change in length of the dome at Mount St Helens (from Voight, 1988)</td>
<td>39</td>
</tr>
<tr>
<td>2.12</td>
<td>Inverse-event rate vs. time diagram shows the change in recorded seismicity before 1991 eruption on Mt. Pinatubo (from Kilburn, 2003)</td>
<td>40</td>
</tr>
<tr>
<td>3.1</td>
<td>Schematic diagram of a hydrostatic, uniaxial and triaxial test</td>
<td>44</td>
</tr>
<tr>
<td>3.2</td>
<td>Clustering of AE events during rock fracturing (from Scholz et al., 1968b)</td>
<td>46</td>
</tr>
<tr>
<td>3.3</td>
<td>Evolution of b-value for triaxial experiments at different conditions (from Sammonds et al., 1992)</td>
<td>49</td>
</tr>
<tr>
<td>3.4</td>
<td>Fractures in the magma after the uniaxial test (from Lavallée et al., 2012)</td>
<td>52</td>
</tr>
<tr>
<td>3.5</td>
<td>Volumetric strain-stress curve and cumulative AE events (from Aker et al., 2014)</td>
<td>53</td>
</tr>
<tr>
<td>3.6</td>
<td>Transition from DC to non DC components during pore pressure release (from Benson et al., 2008)</td>
<td>55</td>
</tr>
<tr>
<td>3.7</td>
<td>AE signals generated during pore pressure release of water and nitrogen (from Benson et al., 2014)</td>
<td>56</td>
</tr>
<tr>
<td>3.8</td>
<td>Different types of AE compared with volcanic earthquakes recorded at Mt. Etna (from Burlini et al., 2007)</td>
<td>58</td>
</tr>
<tr>
<td>4.1</td>
<td>Optical microscope pictures taken from a pre-test sample of EB</td>
<td>59</td>
</tr>
<tr>
<td>4.2</td>
<td>Location of Mt Etna and the quarry supplying EB blocks, cored to produce cylindrical samples</td>
<td>61</td>
</tr>
<tr>
<td>4.3</td>
<td>The Sanchez Technologies triaxial apparatus</td>
<td>64</td>
</tr>
<tr>
<td>4.4</td>
<td>Schematic overview of the Sanchez triaxial apparatus</td>
<td>64</td>
</tr>
<tr>
<td>4.5</td>
<td>Stress – strain curve for the theoretical and measured values on an aluminium-alloy cylinder</td>
<td>66</td>
</tr>
<tr>
<td>4.6</td>
<td>Comparison between the stress-strain curve using the raw values of strain and adjusted strain.</td>
<td>67</td>
</tr>
</tbody>
</table>
Figure 4.7 Sketch of the PZT sensors
Figure 4.8 Frequency response of HF and LF sensors
Figure 4.9 Overview of the AE recording system
Figure 4.10 Schematic figure illustrating a generic PZT transducer and its active element
Figure 4.11 Screenshot of Insite-Lab, showing a continuous 230-ms-long waveform
Figure 4.12 Location of seismic stations at Mt Etna and at Vulcano Island
Figure 5.1 Stress – strain plot of all 7 experiments run at dry conditions
Figure 5.2 Temporal evolution of P-wave velocities for experiment EB21
Figure 5.3 Temporal evolution of P-wave velocities for experiment EB23
Figure 5.4 Temporal evolution of P-wave velocities for experiment EB26
Figure 5.5 Temporal evolution of P-wave velocities for experiment EB32
Figure 5.6 Temporal evolution of P-wave velocities for experiment EB33
Figure 5.7 Temporal evolution of P-wave velocities for experiment EB35
Figure 5.8 Temporal evolution of P-wave anisotropy for the dry experiments
Figure 5.9 Temporal evolution of the hit rate for the dry experiments
Figure 5.10 Temporal evolution of the b-value for the dry experiments
Figure 5.11 Locations of the events recorded around the time of the sample’s failure for the dry experiments
Figure 5.12 Temporal evolution of the hypocentres of the experiments EB21, B23, EB26 and EB31
Figure 5.13 Temporal evolution of the hypocentres of the experiments EB32, EB33, EB35
Figure 5.14 Temporal evolution of the average inter-event distance for the dry experiments
Figure 5.15 Temporal evolution of the average location magnitude for the dry experiments
Figure 5.16 Stress – strain plot of the 7 experiments run at saturated conditions ($p_c = 35$ MPa, $p_p = 5$ MPa)
Figure 5.17 Temporal evolution of P-wave velocities for experiment EB14
Figure 5.18 Temporal evolution of P-wave velocities for experiment EB15
Figure 5.19 Temporal evolution of P-wave velocities for experiment EB16
Figure 5.20 Temporal evolution of P-wave velocities for experiment EB17
Figure 5.21 Temporal evolution of P-wave velocities for experiment EB18
Figure 5.22 Temporal evolution of P-wave velocities for experiment EB19
Figure 5.23 Temporal evolution of P-wave velocities for experiment EB36
Figure 5.24 Temporal evolution of P-wave anisotropy for the saturated experiments ($p_p = 5$ MPa)
Figure 5.25 Temporal evolution of the hit rate for the saturated experiments ($p_p = 5$ MPa)
Figure 5.26 Temporal evolution of the b-value for the saturated experiments ($p_p = 5$ MPa)
Figure 5.27 Locations of the events recorded around the time of the sample’s failure for the saturated experiments (p_p = 5 MPa)

Figure 5.28 Temporal evolution of the hypocentres of the experiments EB14, EB15, EB16, EB17

Figure 5.29 Temporal evolution of the hypocentres of the experiments EB19, EB36

Figure 5.30 Temporal evolution of the average inter-event distance for the saturated experiments (p_p = 5 MPa)

Figure 5.31 Temporal evolution of the average location magnitude for the saturated experiments (p_p = 5 MPa)

Figure 5.32 Stress – strain plot of the 8 experiments run at saturated conditions (p_c = 46 MPa, p_p = 16 MPa)

Figure 5.33 Temporal evolution of P-wave velocities for experiment EB20

Figure 5.34 Temporal evolution of P-wave velocities for experiment EB24

Figure 5.35 Temporal evolution of P-wave velocities for experiment EB25

Figure 5.36 Temporal evolution of P-wave velocities for experiment EB27

Figure 5.37 Temporal evolution of P-wave velocities for experiment EB28

Figure 5.38 Temporal evolution of P-wave velocities for experiment EB29

Figure 5.39 Temporal evolution of P-wave velocities for experiment EB30

Figure 5.40 Temporal evolution of P-wave velocities for experiment EB34

Figure 5.41 Temporal evolution of P-wave anisotropy for the saturated experiments (p_p = 16 MPa)

Figure 5.42 Temporal evolution of the hit rate for the saturated experiments (p_p = 16 MPa)

Figure 5.43 Temporal evolution of the b-value for the saturated experiments (p_p = 16 MPa)

Figure 5.44 Locations of the events recorded around the time of the sample’s failure for the saturated experiments (p_p = 16 MPa)

Figure 5.45 Temporal evolution of the hypocentres of the experiments EB20, EB24, EB25, EB27

Figure 5.46 Temporal evolution of the hypocentres of the experiments EB28, EB29, EB30, EB34

Figure 5.47 Temporal evolution of the average inter-event distance for the saturated experiments (p_p = 16 MPa)

Figure 5.48 Temporal evolution of the average location magnitude for the saturated experiments (p_p = 16 MPa)

Figure 5.49 Cumulative hits superimposed on the pore pressure – time plot for the 3 experiments with liquid water as pore fluid (T = 25°C).

Figure 5.50 Long-duration AE activity recorded during experiment EB16

Figure 5.51 Long-duration AE activity recorded during experiment EB29
Figure 5.52  Long-duration AE activity recorded during experiment EB34
Figure 5.53  Similarity plots for the experiments releasing liquid water
Figure 5.54  Correlated and master events for the experiments releasing liquid water
Figure 5.55  Temporal evolution of the dominant frequency and variation of the FWHM for the experiments releasing liquid water
Figure 5.56  Cumulative hits superimposed on the pore pressure – time plot for 6 experiments with water and steam as pore fluid (T = 175°C).
Figure 5.57  Long-duration AE activity recorded during experiment EB18
Figure 5.58  Long-duration AE activity recorded during experiment EB19
Figure 5.59  Long-duration AE activity recorded during experiment EB27
Figure 5.60  Long-duration AE activity recorded during experiment EB28
Figure 5.61  Long-duration AE activity recorded during experiment EB15
Figure 5.62  Long-duration AE activity recorded during experiment EB36
Figure 5.63  Similarity plots for the experiments releasing superheated water and steam (1)
Figure 5.64  Similarity plots for the experiments releasing superheated water and steam (2)
Figure 5.65  Correlated and master events for the experiments releasing superheated water and steam (1)
Figure 5.66  Correlated and master events for the experiments releasing superheated water and steam (2)
Figure 5.67  Temporal evolution of the dominant frequency and variation of the FWHM for the experiments releasing superheated water and steam (1)
Figure 5.68  Temporal evolution of the dominant frequency and variation of the FWHM for the experiments releasing superheated water and steam (2)
Figure 5.69  Bimodality plot for the experiments releasing superheated water and steam
Figure 5.70  Cumulative hits superimposed on the pore pressure – time plot for the 4 experiments with nitrogen gas as pore fluid (T = 25°C).
Figure 5.71  Long-duration AE activity recorded during experiment EB31
Figure 5.72  Long-duration AE activity recorded during experiment EB32
Figure 5.73  Long-duration AE activity recorded during experiment EB33
Figure 5.74  Long-duration AE activity recorded during experiment EB35
Figure 5.75  Similarity plots for the experiments releasing nitrogen gas
Figure 5.76  Correlated and master events for the experiments releasing nitrogen gas
Figure 5.77  Temporal evolution of the dominant frequency and variation of the FWHM for the experiments releasing nitrogen gas
Figure 6.1  Stereonets of the survey of intact rock in dry conditions and with pressurized water at 5 MPa and 16 MPa
Figure 6.2  P-wave anisotropy versus Strain percentage for experiments run in dry conditions and with pressurized water at 5 MPa and 16 MPa

Figure 6.3  Hit rate versus strain percentage relative for experiments run in dry conditions and with pressurized water at 5 MPa and 16 MPa

Figure 6.4  Pore volume change curve for the water saturated experiments with $p_p = 5$ MPa and $p_p = 16$ MPa

Figure 6.5  Sketch of the precursory seismic rate at different volcanoes

Figure 6.6  Temporal evolution of the b-value for experiments run in dry conditions and with pressurized water at 5 MPa and 16 MPa

Figure 6.7  Pictures of the failed samples used for experiments run in dry conditions and with pressurized water at 5 MPa and 16 MPa

Figure 6.8  Temporal evolution of the average location magnitude for experiments run in dry conditions and with pressurized water at 5 MPa and 16 MPa

Figure 6.9  Temporal evolution of the average inter-event distance for experiments run in dry conditions and with pressurized water at 5 MPa and 16 MPa

Figure 6.10  Stress-strain curve for experiments using solid and pre-drilled samples

Figure 6.11  Stereonets of the survey of intact rock solid and pre-drilled samples

Figure 6.12  P-wave anisotropy versus Strain percentage for experiments using solid and pre-drilled samples

Figure 6.13  Hit rate versus strain percentage relative for experiments using solid and pre-drilled samples

Figure 6.14  Temporal evolution of the b-value for experiments using solid and pre-drilled samples

Figure 6.15  Temporal evolution of the average location magnitude for experiments using solid and pre-drilled samples

Figure 6.16  Temporal evolution of the average inter-event distance for experiments using solid and pre-drilled samples

Figure 6.17  Requirements for the onset of long-duration AE events

Figure 6.18  Comparison between the waveform and the envelope of a long-duration AE signal

Figure 6.19  Sketch illustrating the parameters used in Table 6.2 and 6.3

Figure 6.20  Pressure drop peak versus envelope peak

Figure 6.21  Normalized signal envelope and normalized absolute pressure drop during the release of liquid water, water & steam and nitrogen gas

Figure 6.22  Waveform and frequency spectrum of the long-duration AE during the release of liquid water, water & steam and nitrogen gas

Figure 6.23  Temporal evolution of the cumulative hits during the release of liquid water, water & steam and nitrogen gas
Figure 6.24  Variation of the Quality factor and in time and frequency during the release of liquid water, water & steam and nitrogen gas 224

Figure 6.25  Temporal evolution of dominant frequency, characteristic FWHM and bimodality evaluation for experiment B28 225

Figure 6.26  Continuous signal and associated spectrogram for experiment EB31 and EB33 227

Figure 6.27  Temporal evolution of the dominant frequency and variation of the FWHM for the experiments releasing nitrogen gas 228

Figure 6.28  Similarity of the events during the release of liquid water, water & steam and nitrogen gas (1) 229

Figure 6.29  Similarity of the events during the release of liquid water, water & steam and nitrogen gas (2) 230

Figure 6.30  Continuous signal recorded during the deformation stage and during the venting stage, plotted using the same vertical scale 231

Figure 6.31  Average source components of the events during the release of liquid water, water & steam and nitrogen gas 232

Figure 6.32  Comparison between a long-duration AE (gas release) and a Tornillo 234

Figure 6.33  Comparison between a long-duration AE (water & steam release) and tremor 236

Figure 6.34  Comparison between a short-duration AE (water & steam release) and a LP event 237

Figure 6.35  Comparison between a short-duration AE (water & steam release) and a VLP 238
<table>
<thead>
<tr>
<th>Table</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1</td>
<td>Volcanic earthquakes terminology</td>
<td>22</td>
</tr>
<tr>
<td>4.1</td>
<td>Physical properties of EB</td>
<td>60</td>
</tr>
<tr>
<td>5.1</td>
<td>Overview of the 22 experiments</td>
<td>80</td>
</tr>
<tr>
<td>5.2</td>
<td>Physical properties and results of the 7 experiments run at dry conditions (p_c = 30 MPa, p_p = 0)</td>
<td>82</td>
</tr>
<tr>
<td>5.3</td>
<td>P-wave velocities statistics of the 6 over 7 experiments run at dry conditions (p_c = 30 MPa, p_p = 0)</td>
<td>84</td>
</tr>
<tr>
<td>5.4</td>
<td>Physical properties and results of the 7 experiments run at saturated conditions (p_c = 35 MPa, p_p = 5 MPa)</td>
<td>105</td>
</tr>
<tr>
<td>5.5</td>
<td>P-wave velocities statistics of the 7 experiments run at saturated conditions (p_c = 35 MPa, p_p = 5 MPa)</td>
<td>107</td>
</tr>
<tr>
<td>5.6</td>
<td>Physical properties and results of the 8 experiments run at saturated conditions (p_c = 46 MPa, p_p = 16 MPa)</td>
<td>127</td>
</tr>
<tr>
<td>5.7</td>
<td>P-wave velocities statistics of the 8 experiments run at saturated conditions (p_c = 46 MPa, p_p = 16 MPa)</td>
<td>129</td>
</tr>
<tr>
<td>5.8</td>
<td>Conditions and results of the 3 venting experiments run at room temperature (T = 25°C), using water as pore fluid</td>
<td>151</td>
</tr>
<tr>
<td>5.9</td>
<td>Conditions and results of the 7 venting experiments run at 175°C, using water as pore fluid</td>
<td>160</td>
</tr>
<tr>
<td>5.10</td>
<td>Conditions and results of the 4 venting experiments run at room temperature (T = 25°C), using nitrogen as pore fluid</td>
<td>178</td>
</tr>
<tr>
<td>6.1</td>
<td>Summary of the experimental data analysed in this study</td>
<td>187</td>
</tr>
<tr>
<td>6.2</td>
<td>Variables and features of the long-duration AE activity</td>
<td>213</td>
</tr>
<tr>
<td>6.3</td>
<td>Reduced list of variables and features of the long-duration AE activity</td>
<td>214</td>
</tr>
<tr>
<td>6.4</td>
<td>Summary of the venting stage characteristics and associated field analogue</td>
<td>233</td>
</tr>
<tr>
<td>Acronym</td>
<td>Description</td>
<td></td>
</tr>
<tr>
<td>---------</td>
<td>--------------------------------------------------</td>
<td></td>
</tr>
<tr>
<td>AE</td>
<td>Acoustic Emission</td>
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</tr>
<tr>
<td>CLVD</td>
<td>Compensated Linear Vector Dipole</td>
<td></td>
</tr>
<tr>
<td>D'</td>
<td>Onset of dilatancy</td>
<td></td>
</tr>
<tr>
<td>DC</td>
<td>Double-couple</td>
<td></td>
</tr>
<tr>
<td>E</td>
<td>Young’s Modulus</td>
<td></td>
</tr>
<tr>
<td>EB</td>
<td>Etna Basalt</td>
<td></td>
</tr>
<tr>
<td>EDF</td>
<td>Eddy Displacement system</td>
<td></td>
</tr>
<tr>
<td>ERR_{Loc}</td>
<td>Location Error</td>
<td></td>
</tr>
<tr>
<td>FFM</td>
<td>Failure Forecasting Model</td>
<td></td>
</tr>
<tr>
<td>FFT</td>
<td>Fast Fourier Transform</td>
<td></td>
</tr>
<tr>
<td>FWHM</td>
<td>Full Width Half Maximum</td>
<td></td>
</tr>
<tr>
<td>GLM</td>
<td>Generalized Linear Model</td>
<td></td>
</tr>
<tr>
<td>HB</td>
<td>Hybrid</td>
<td></td>
</tr>
<tr>
<td>HCD</td>
<td>Hit Count Data</td>
<td></td>
</tr>
<tr>
<td>HF</td>
<td>High Frequency</td>
<td></td>
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<tr>
<td>HT</td>
<td>High Temperature</td>
<td></td>
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<tr>
<td>ISO</td>
<td>Volumetric component</td>
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<tr>
<td>LF</td>
<td>Low Frequency</td>
<td></td>
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<tr>
<td>LP</td>
<td>Long Period</td>
<td></td>
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<tr>
<td>LT</td>
<td>Low Temperature</td>
<td></td>
</tr>
<tr>
<td>LVDT</td>
<td>Linear Variable Differential Transformers</td>
<td></td>
</tr>
<tr>
<td>Mg</td>
<td>Mach number</td>
<td></td>
</tr>
<tr>
<td>ML</td>
<td>Maximum Likelihood</td>
<td></td>
</tr>
<tr>
<td>M_{L}</td>
<td>Location magnitude</td>
<td></td>
</tr>
<tr>
<td>MT</td>
<td>Moment Tensor</td>
<td></td>
</tr>
<tr>
<td>M_{w}</td>
<td>Moment Magnitude</td>
<td></td>
</tr>
<tr>
<td>PAD</td>
<td>Pulser Amplifier Desktop</td>
<td></td>
</tr>
<tr>
<td>p_{c}</td>
<td>Confining Pressure</td>
<td></td>
</tr>
<tr>
<td>p_{eff}</td>
<td>Effective Pressure</td>
<td></td>
</tr>
<tr>
<td>p_{p}</td>
<td>Pore pressure</td>
<td></td>
</tr>
<tr>
<td>PZT</td>
<td>Piezo electric</td>
<td></td>
</tr>
<tr>
<td>Q</td>
<td>Quality factor</td>
<td></td>
</tr>
<tr>
<td>RMS</td>
<td>Root Mean Square</td>
<td></td>
</tr>
<tr>
<td>R^2</td>
<td>Coefficient of determination</td>
<td></td>
</tr>
<tr>
<td>SNR</td>
<td>Signal to Noise Ratio</td>
<td></td>
</tr>
<tr>
<td>T</td>
<td>Temperature</td>
<td></td>
</tr>
<tr>
<td>UCS</td>
<td>Unconfined Compressive Strength</td>
<td></td>
</tr>
<tr>
<td>VLP</td>
<td>Very-Long Period</td>
<td></td>
</tr>
<tr>
<td>VT</td>
<td>Volcano Tectonic</td>
<td></td>
</tr>
<tr>
<td>\varepsilon</td>
<td>Strain</td>
<td></td>
</tr>
<tr>
<td>\rho</td>
<td>Density</td>
<td></td>
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<tr>
<td>\sigma</td>
<td>Stress</td>
<td></td>
</tr>
</tbody>
</table>
1. INTRODUCTION

1.1. VOLCANO: A THREATENING NEIGHBOUR

Hundreds of millions of people sleep with a dangerous threat just outside their door: a volcano. Some of the largest cities in the world (e.g. Mexico City, Seattle, Quito, Manila, Tokyo and Naples) can be hit by a nearby volcanic eruption. Although volcanic soil constitutes a fertile substratum for harvesting and provides unique materials, the very process creating such grounds can destroy everything in a matter of seconds.

In ancient history, it is believed that the Minoan civilization, living the island of Crete (Mediterranean Sea), was wiped out by a caldera-forming eruption at Santorini, a volcanic-island some 100 km north. What previously was a cone-shaped edifice is now a caldera, with a new cone building inside.

Perhaps the most known historical eruption occurred in 79 AD at Mt. Vesuvius, south of Naples (Italy). Its present shape, with a central cone and the arcuate shaped Monte Somma, is the result of the long eruptive history following the plinian eruption. The Roman civilization built several cities around the volcano (e.g. Herculaneum and Pompeii) which were destroyed by this major event. The remains of those cities are now worldwide heritage sites. What makes this eruption so famous is the uniquely detailed description made by Pliny the Younger, which constituted the first recorded volcanological observation.

Volcanic eruptions do not only have local effects. Depending on their magnitude and location, they may have global consequences. The eruption of Mt. Tambora in 1815, largely considered as the most powerful eruption in recorded history, is thought to be the cause of the so-called “year without a summer” in 1816, characterized by a global lowering of the average temperature. More recently, in 2010, a relative small eruption occurred at Eyjafjallajökull volcano (Iceland). The large amount of ashes generated and dispersed in the atmosphere was carried by the winds into the commercial routes of transatlantic aviation, causing almost 2 months of disruption in Europe.

Therefore understanding the physics of volcanoes, particularly the mechanisms preceding major eruptions is without doubt the great challenge for all people fascinated by such a powerful, yet dangerous geologic feature.
1.2. VOLCANO TECTONICS AND SEISMOLOGY

Seismicity and ground deformation represent two of the most common short-term phenomena detected before volcanic unrest, whether this results in an eruption or otherwise. For cases when an eruption does result, the final approach is commonly preceded by an accelerating occurrence rate of different geophysical indicators. Seismologically, key data are high-frequency volcano-tectonic (VT) earthquakes and Low-Frequency (LF) and harmonic tremor events, both of which are generated in the volcano-tectonic system but with different dominant frequencies. These characteristic seismic signals are unique to volcanoes and have been linked to the interplay of magma ascent and volcanic edifice response over the final hours before eruption. Because of this, for a time there was hope that a better understanding of their generation might give way to more accurate forecasting methods, but this has since been proven overly optimistic (Neuberg et al., 2006). This is partly due to our lack of understanding of the rock-fluid coupling that generates these signals and under relevant pressure, stress, and trigger conditions. In this thesis, new insights into the details of the rock-fluid coupling are presented under conditions relevant to volcanic conditions using laboratory simulations to reproduce scaled-down volcanic earthquakes.

The laboratory analogue of macro-scale earthquakes are microseismic signals, known as Acoustic Emission (AE), and have become a well-used proxy for laboratory studies of rock deformation for many years (e.g. Scholz 1968a, 1968b; Ohnaka & Mogi, 1982). These signals are characterized by very low magnitude (e.g. Magnitude -4), high frequency content (tens of kHz to MHz), limited propagation (mm to 10’s of cm) and are recorded by piezoelectric transducers that convert the generated mechanical stress into a voltage. For the purposes of this research, the source of the AE signals is either the fracturing process at the micro-scale (microcracking), or subsequent movement of small quantities of fluid through the generated damage zone.

Importantly, the physics of the generation process is the same as the field-scale earthquakes, except for scale: with field sources being $10^4$ – $10^5$ times larger. On the other hand, the frequency content of AEs is $10^4$ – $10^5$ times higher, which contributes to apply the size-frequency scaling law to relate these two scenarios (e.g. Burlini et al., 2007). However, even if AEs and earthquakes do have different magnitude, frequency and source size, they share similar frequency-magnitude distribution (i.e. $b$-value), temporal evolution of number of events, locations and spectral pattern. For these reasons, the study of laboratory generated events under controlled conditions of stress
and pressure can provide important new information about field-generated signals that cannot be directly accessed.

1.3. AIMS AND OBJECTIVES

The aim of this research project is to produce AE activity due to specific set of relevant volcanic conditions (state of stress, temperature, and pore pressure) during the processes of rock fracturing and fluids movements. In doing so, this study explores how differing types of signals are generated, similar to those occurring at volcanoes, linking them to the known generation conditions and hence to the physical process. In this way, the analysis of volcanic earthquakes could provide us precious new information regarding the state of the volcano. The great advantage of laboratory simulations is that several different parameters, such as stress and strain, porosity and permeability, temperature and fluid composition, are directly measured and controlled. On a volcano, most of these parameters cannot be measured, leading to interpretations and assumptions about the origin of volcanic earthquakes.

In the case of volcano seismicity the family of low-frequency activity can be further split into Tremor, Long Period (LP), Very-Long Period (VLP) and Tornillo events. Over the years, numerous models (discussed in chapter 2) have been developed to explain their origin. The generally accepted theory is that these signals are generated as a direct result of fluid-movement through fractures and cracks. Following this, the thesis explores the role played by high pressure pore fluids in a damaged rock mass undergoing deformation (which leads to failure and, at the macro-scale, eruption). And, when the same fluids move through the fractured rock, providing a mechanism for the low-frequency activity.

1.4. THESIS OUTLINE

In the Chapters 2 and 3 a detailed literature review is presented: Chapter 2 deals with the concept of Volcano-seismology and the classification of the different type of signals recorded on volcanoes, together with the several theories regarding the source mechanism. In addition, a brief overview of the methods used for eruption forecasting is discussed. Chapter 3 develops this concept further and introduces the integration of AE data with rock deformation experiments, with particular emphasis on the AEs and their use to characterize the level of deformation and to
investigate the fluid-induced seismicity. Scaling laws between AEs and field-scale earthquakes are also discussed.

Chapter 4 introduces the methods used to obtain data, including rock characterization and sample preparation, the equipment used to collect mechanical and acoustic data, and experimental protocol, while Chapter 4 also describes the types of data collected in laboratory and gathered by external institution (i.e. field-scale volcanic earthquakes). Chapter 5 first describes the physical properties of the rock material, further dividing these aspects into two sections. The first section lists the results obtained during the triaxial deformation stage of the experiments, which include stress-strain data, P-wave velocities and AEs features (counts, location, magnitude, frequency). The second section details the results gathered during the pore pressure release stage, mainly focusing on the frequency content and amplitude of both continuous and transient signals.

Chapter 6 is the discussion of the results. First, the effect of pore fluid on the mechanical and acoustic properties of the sample, during deformation, is discussed. Second, the relationships between the physical properties of the pore fluids used and the AE activity generated by the movement of that fluid is explored.

Finally, Chapter 7 presents the conclusions of this study, including implications for the data in interpreting field-scale events and eruption forecasting.
2. VOLCANO SEISMOLOGY

2.1. TERMINOLOGY

Volcano seismology is “the study of earthquakes as well as of velocity structure, attenuation, and other physical properties of earth materials that affect the passage of seismic waves at volcanoes” (McNutt, 1996, p.100). Earthquakes occurring at volcanoes, or caused by volcanic processes, are therefore called volcanic earthquakes and can be traditionally classified in terms of spectral features of their seismograms (McNutt, 1996), although different local terminologies have been widely used in the past (Table 2.1). Malone et al. (1983) recognized the spectral similarity between tectonic and VT activity on Mount St. Helens (USA), with the tectonic events located at some distance from the volcano and unrelated to the volcanic activity. Although characterized by different arrival times, the shallow volcanic earthquakes, medium frequency (m-type) and low frequency (l-type) events all share similar frequency content.

A number of key studies, dating back to the 1970’s, have greatly informed scientific opinions on the interpretation, understanding, and observation in volcano seismology. The first tomography images of the volcano plumbing system were derived from earthquake travel-times as developed in the pioneering work of Aki et al. (1977a) together with Aki & Lee (1976), Aki et al. (1976), Husebye et al. (1976), and Aki (1977). For the first time, scientists were able to study volcanic structures and associated magma reservoirs based on the diagnostic seismicity generated due to the interaction between the stressed volcanic edifice and movement of magmatic fluids (Aki et al., 1977b). Early observations of these signals were given the name B-Type and volcanic tremor, where their source was associated to fluid-driven cracks. The first estimations of magma transport budget beneath Kilauea Volcano (Hawaii, USA) was possible via seismic observation by Aki & Koyanagi (1981) with the energetics of volcanic tremor beneath Long Valley Caldera (USA) quantified by Aki (1984).
Table 2.1: Volcanic earthquakes terminology

<table>
<thead>
<tr>
<th>PAPER</th>
<th>VOLCANO</th>
<th>BASED ON</th>
<th>TYPES OF VOLCANIC EARTHQUAKES</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minakami (1974)</td>
<td>Japan</td>
<td>Frequency</td>
<td>A-type</td>
</tr>
<tr>
<td>Latter (1979)</td>
<td>Ruapehu – Mt. Ngauruhoe (New Zealand)</td>
<td>Frequency</td>
<td>Volcano-tectonic (VT)</td>
</tr>
<tr>
<td>Malone et al. (1983)</td>
<td>Mt. St. Helens (USA)</td>
<td>Location and frequency</td>
<td>Tectonic like (t- and h-type)</td>
</tr>
<tr>
<td>Lahr et al. (1994)</td>
<td>Redoubt Volcano (Alaska, USA)</td>
<td>Source</td>
<td>VT</td>
</tr>
<tr>
<td>Julian (1994)</td>
<td>-</td>
<td>Frequency and shape</td>
<td>-</td>
</tr>
<tr>
<td>McNutt (1996)</td>
<td>-</td>
<td>Frequency</td>
<td>High-frequency (HF)</td>
</tr>
<tr>
<td>Ohminato et al. (1998)</td>
<td>Kilauea (USA)</td>
<td>Frequency</td>
<td>Short period</td>
</tr>
<tr>
<td>Neuberg (2000)</td>
<td>Soufriere Hills (Montserrat)</td>
<td>Source</td>
<td>VT</td>
</tr>
<tr>
<td>McNutt (2005)</td>
<td>-</td>
<td>Frequency</td>
<td>HF</td>
</tr>
<tr>
<td>Alparone et al. (2010)</td>
<td>Vulcano (Italy)</td>
<td>Frequency</td>
<td>HF</td>
</tr>
<tr>
<td>Jousset et al. (2013)</td>
<td>Mt. Merapi (Indonesia)</td>
<td>Source</td>
<td>VT</td>
</tr>
<tr>
<td>Bean et al. (2014)</td>
<td>Mt. Etna (Italy), Turrialba Volcano (Costa Rica), Ubinas (Peru)</td>
<td>Source</td>
<td>Stress driven mechanical class (VT, LP)</td>
</tr>
</tbody>
</table>
While Minakami (1974) postulated that both high-frequency (HF) A-type and low-frequency (LF) B-type earthquakes were related to rock fracture processes, but differing in terms of locations (the latter being deeper, resulting in the high frequencies being attenuated), Latter (1979) suggested that LF events and volcanic tremor were essentially the same, as evidenced by similar spectral features, and that the swarms of LF signals may simply merge into quasi-continuous tremor. 

Further waveform categories called L-type and m-type, became better known as LP (Long Period) and HB (hybrid) earthquakes respectively, with the term “hybrid” highlighting the mixed characteristics of the high energy VT-like onset and the LP-like resonant coda (Lahr et al., 1994). In line with Malone et al. (1983), Neuberg (2000) then grouped the LP and HB events in the same LF (Low-Frequency) class, as they represent the two end-members of a continuum of signals.

Tornillos (Spanish for “screws”) – which are long-coda events – have also been recorded on a wide range of volcanoes (e.g. Galeras, Colombia), and were included in the LP class by Julian (1994). Meanwhile, Alparone et al. (2010), when describing the seismicity occurring at Vulcano (Italy), considered them as a particularly long event pertaining to the monochromatic signals. Bean et al. (2014) also consider Tornillo as a separate class from LP signals, with the latter included in a broader stress-driven mechanical class together with the VT earthquakes. On the other hand, Jousset et al. (2013) grouped LP, Multiphase (Hybrid), tremor and VLP events in the Long Period Seismicity (LPS) class, and thus separating brittle failure processes (VT) from the fluid-induced seismicity (LPS) driven processes. Suffice to say, the wide range of different opinions and possibilities for classification makes the task a challenge.

Key to recording and subsequent understanding all of these unique events was greatly enhanced with the development and deployment of broadband seismometers (period up to 100 s), more recently extending to Very-Long-Period or VLP signals at periods up to 200s (Ohminato et al., 1998). However, the use of the term “long period” was criticized by McNutt (1996; 2005) for a number of reasons. Firstly because, while many authors (e.g. Lahr et al., 1994) described the events in terms of the sources, they tended to use this as a purely descriptive term. Secondly, in seismology long period represent both the instrumentation and the energy of earthquakes with period longer than the Earth noise peak, of around 10s (McNutt, 1996) making this subjective description less than ideal. For these reasons, the term “low frequency” was preferred for classification as it was based on the spectral features and using generic terms in place of source-
based terminology. Still, the term LP has been preferred by many authors (e.g. Waite et al., 2008; Alparone et al., 2010; Jousset et al., 2013; Zecevic et al., 2013; Bean et al., 2014; Cauchie et al., 2015) when referring to this particular class. Due to this, the terminology for the field data in this thesis is based on a combination of Lahr et al. (1994), Ohminato et al. (1998) and Alparone et al. (2010): VT, LP, Tremor, HB, Tornillo, and VLP events.

To date, the most used terminology is still that published in Lahr et al. (1994), with the addition of the VLP term (Cassisi et al., 2016; Curilem et al., 2016; De Arcangelis et al., 2016; Lara-Cueva et al., 2016; La Rocca & Galluzzo, 2016). While these source-based terms are becoming a standard, avoiding confusion among researchers, they also carry the author’s interpretation. As one can see in section 2.3, some events still lacks of a univocal interpretation leading to different seismo-genetic processes to generate seismic signals having the same name. To avoid this, a frequency-based terminology should be adapted, which incorporates both time- and frequency-domain features. This could be achieved by combining McNutt (1996, 2005) and Alparone et al. (2010) terms. In this way, for example, Tornillo may be called “LF slowly decaying” events.

2.2. CHARACTERISTIC FEATURES

VT events resemble tectonic earthquakes in terms of spectral components, but have a source located at or near a volcanic system (Chouet, 1996) and are generally of low magnitude (up to 4). They are characterized by an impulsive onset, clear P- and S- phase arrivals and a short coda (Fig. 2.1, Lahr et al., 1994). In terms of frequency, they possess a broadband signal with the majority of seismic energy between 5 and 25 Hz (Alparone et al., 2010), a peak energy of the P- and S-phases ranging between 6 and 8 Hz and coda with energy up to 15 Hz (Lahr et al., 1994). Shallow VT events show a slightly longer and narrower band coda than deep VT signals (Fig. 2.1a and 2.1d), with a more emergent P-and S-wave onset due to scattering effects (Fig. 2.1b and 2.1e, Wassermann, 2012). However, they are still distinguishable from HB or LP signals (Chouet, 1996). In terms of polarity, VT events show a mix of first motion polarities with hypocentres well-spaced, both in space and time (Chouet, 1996).
While tectonic earthquakes generally occur in sequences of foreshocks – mainshock–aftershocks, Volcano-Tectonic events are characterised by swarms of events with less than 0.5 magnitude unit between the largest and second largest events (McNutt, 1996). The absence of classic foreshock/mainshock/aftershock pattern for VT events is due to a high rate of seismic event generation during seismic crisis, which has the effect of making individual events undetectable in many cases (Traversa & Grasso, 2010). Another difference between tectonic and VT signals is the Gutenberg-Richter magnitude distribution: both show a typical behaviour overall (Lahr et al., 1994), but, while tectonic events have $b$-value close to 1.0 (Frohlich and Davis, 1993), VT events have a generally higher $b$-value that can vary significantly, both in space (e.g. zones of exsolation level of gases, McNutt, 2005) and in time (e.g. eruptive phases, Roberts et al., 2015).

LP signals differ from VT events in that they have an amplitude envelope similar to that of tectonic earthquakes, but with a spectrum more similar to that of tremor (Chouet, 1992). They are characterized by an emergent high frequency (up to 10 Hz) onset, followed by a harmonic, quasi monochromatic, 20 to 30 second-long coda (Chouet, 1992). Sharp peak frequencies are usually in the range 0.5-5 Hz (Fig. 2.2a, e.g. Lahr et al. 1994; Neuberg, 2000; Alparone et al., 2010), with the S-phase generally absent (Fig. 2.2b, McNutt, 1996) and sometimes with several spectral peaks equally spaced (Fig. 2.2c, e.g. Fehler & Chouet, 1982; Jousset et al. 2003). First motions of LP
signals are dilatational and consistent with crack collapse (Lahr et al., 1994; Waite et al., 2008).

When forming a swarm, LP signals have highly consistent waveforms and spectra (Jousset et al., 2003; Saccorotti et al., 2007; Waite et al., 2008; Alparone et al., 2010) in terms of frequency and duration.

Importantly, and unlike VT events or tectonic earthquakes, LP events do not follow a linear Gutenberg-Richter magnitude distribution (Lahr et al., 1994; Bean et al., 2014; Cauchie et al., 2015), and often possess a $b$-value up to $\sim 3$, indicating a very large number of small magnitude events originating from a small source (McNutt, 1996). The reason for this is straightforward, as they are not generated due to a fracture event, but instead form due to rock-fluid coupling, hence explaining the lack of a foreshock activity.

![Figure 2.2](image)

*Figure 2.2: (a) spectrogram, (b) waveform and (c) spectra of a LP event recorded at Redoubt Volcano. Modified from Lahr et al. (1994).*

Tremor is a limited-bandwidth continuous signal of sustained amplitude (Fig. 2.3, Chouet, 1992) lasting several minutes, days or even months (Chouet, 1996). It starts with an emergent onset (Hofstetter & Malone, 1986; Chouet, 2003) at high frequencies (5-10 Hz) and moderate amplitude. Signals then evolve to a lower frequency (1.5-3 Hz) and higher amplitude, with sporadic (but weak) higher frequencies towards the end of the signal (Aki & Koyanagi, 1981). These systematic variations in time of evenly spaced, spectral peaks produce one of the more peculiar features of tremor, known as gliding lines (McNutt, 1996; Neuberg, 2000), probably caused due to gas bubble oscillations (Unglert & Jellinek, 2015). Like LP signals, they possess a dominant spectral content in the 1-5 Hz frequency band (McNutt, 1996), similar spectral
peaks recorded across different stations (Aki & Koyanagi, 1981), and a spectrum consisting of regularly spaced, narrow peaks (Julian, 1994). Generally, however, tremor spectra are more chaotic with no clear dominant frequencies and irregularly spaced (Julian, 1994). In addition, tremor has been observed that appears to consist of a particular sequence of LP activity (e.g. Minakami, 1974) suggesting that swarms of LP events could merge into tremor under certain conditions (Latter, 1981; Neuberg, 2000). Negative and positive first motions occur equally in tremor signals, and locations usually lie in the same source area of the LP event (Aki & Koyanagi, 1981; Chouet, 2003).

![Figure 2.3: (a) 3-days-long waveforms and (b) spectrogram of the tremor recorded at El Hierro Island, Spain. Modified from Tarraga et al. (2014).](image)

HB events share features from both VT and LP signals: they have an impulsive, high-frequency onset (Fig. 2.4a) showing mixed polarities (like VT signals), but with a longer coda (Fig. 2.4b) and characterized by non-dispersive harmonic frequencies (Fig. 2.4c) typical of the LP signals (Lahr et al., 1994). When HB events occur in swarms, they often show a high degree of waveform similarity (Jousset et al., 2003; Alparone et al., 2010). Their hypocentres are found to lie above typical sources of VT earthquakes but below the source of LP events (Lahr et al., 1994), with the proportion of high frequency energy indicating the depth of the source (Neuberg et al., 1998). Because LP signals have highly emergent onsets they are difficult to locate, HB events are often used to infer the source dimension of LP events as a proxy. In terms of magnitude distribution they show a similar behaviour to that of LP signals, with a rectangular shape, rather than the linear signal typical of VT events (Lahr et al., 1994).
An unusual and distinct seismic signal, first recorded at Galeras, have been termed with the Spanish name Tornillos, due to the resemblance of the envelope to a screw (Julian, 1994; Torres et al., 1996; Gomez & Torres, 1997). These events, later recognized in several other volcanoes and distinguished from the LP signals, are characterized by a long, quasi-linear decaying codas, sometimes with amplitude modulation effect (Fig. 2.5a). They have a duration of up to several minutes and quasi-monochromatic waveforms at low frequency, sometimes with weak, high frequency onset (Fig. 2.5b, Torres et al., 1996). The onset itself is emergent, and commonly with a positive polarity or a weak negative impulse followed by a larger positive polarity (Gomez & Torres, 1997). Spectral analysis shows a small number (1-3) narrow peaks (Fig. 2.5c), with the same value recorded at all stations (Torres et al., 1996; Gomez & Torres, 1997), but with different characteristic frequencies measured at different volcanoes (in the range 0.9 – 8.0 Hz, Gomez & Torres, 1997). Cross-correlation analysis reveals high waveform similarity (Alparone et al., 2010) and relatively shallow hypocentres have been inferred by the attenuation pattern (Gomez & Torres, 1997).

To distinguish Tornillos from LP data, a Slenderness parameter was first introduced by Gomez & Torres (1997). This is defined as the ratio between duration and maximum amplitude, revealing that Tornillos have much larger values of slenderness than LP events: up to a factor of 10. Another parameter that distinguish Tornillos from the other classes of volcano seismicity is the damping coefficient for coda waves. This parameter, depending on the shape of the waveform, compares the amplitude at any point in the coda with the amplitude at the

Figure 2.4: (a) spectrogram, (b) waveform and (c) spectra of a HB event recorded at Redoubt Volcano. Modified from Lahr et al. (1994).
end and increases for sharply decaying coda. It was found that the damping coefficient varies from 0.002 to 0.02 for Tornillos, whereas this parameter ranges from 0.01 to 0.025 for the LP signals and from 0.01 to 0.04 for the VT events. (Gomez & Torres, 1997).

Figure 2.5: (a) waveforms, (b) spectrograms and (c) spectra of four Tornillos event recorded at Vulcano (Italy). Modified from Milluzzo et al. (2010).

Finally VLP events have been observed as a series of sawtooth displacement pulses with rise time of few minutes and drop time of 5-10 s at Kilauea (Fig. 2.6, Ohminato et al., 1998). These signals have been called Very Long Period events, because they have similar harmonic behaviour with LP events and tremor, but the energy of these signals is concentrated at lower frequencies, between 0.01 and 0.3 Hz (McNutt, 2005). All first motions are either compressional or dilatational (Ohminato et al., 1998; Jousset et al., 2013) and, like all other classes but VT activity, a high degree of waveform similarity has been observed for these signals (Ohminato et al., 1998; Waite et al., 2008; Cauchie et al., 2015).
Figure 2.6: Series of VLP pulses recorded at different seismic stations at Kilauea. The signals has been bandpass filtered between 0.02 – 0.125 Hz and amplitude normalized. Modified from Ohminato et al. (1998).

2.3. POSSIBLE SOURCE MECHANISMS

Shear failure or stick-slip processes on faults are widely accepted as the causes of VT activity (e.g. McNutt, 1996; 2005; Lahr et al., 1994; Chouet, 1996; Neuberg, 2000). This destructive source mechanics is also suggested by a low degree of waveform similarity (Alparone et al., 2010). The source of elastic strain energy, regardless of rock failure or stick-slip, is provided by magmatic processes where fluid movement within fracture networks affect the structural integrity of the volcano, but which are themselves not directly involved (Chouet 1996). Other sources have been suggested for the triggering of VT events, including regional tectonic forces, gravitational loading (Moran, 2003) and large remote earthquakes (e.g. Power et al., 2001).

The high degree of waveform similarity, combined with a small source region suggests a repetitive, non-destructive source mechanism for all other classes of events (e.g. Chouet, 1992; Gomez & Torres, 1997; Ohminato et al., 1998; Neuberg, 2000; Jousset et al., 2003; Waite et al., 2008; Cauchie et al., 2015). Originally LP events were thought to differ from the VT events in terms of source location, with the transition between HB and LP signals due to different rock types affected by the maximum deformation front (Minakami et al., 1974; Malone et al., 1983).

However, although focal mechanisms were not available, Malone et al. (1983) suggested another explanation for LP signals via the source mechanism itself. Further evidence for this idea comes from the relative consistency in terms of frequency recorded at all stations, implying that a source
effect rather than a path effect was responsible to explain the frequency content (e.g. Aki & Koyanagi, 1981). In such a scenario, fluids may be actively involved in the generation of such signals (Chouet, 1992).

The similarities between LP event and tremor suggest a common source process for both types of seismicity (Julian, 1994; Chouet, 1996; Jousset et al., 2003): LP events are viewed as a response of the tremor - generating system to a sudden transitory pressure, whilst tremor is generated by pressure fluctuations (e.g. Aki et al., 1977b; Aki & Koyanagi, 1981; Chouet, 1996). The fluid-filled crack model (Fig. 2.7) originally proposed by Aki et al. (1977b) is the most regarded source mechanism for both LP signals and tremor, where an excess of magmatic pressure induces a fluid-filled crack to resonate, and in turn, generate a far-field seismic response recorded as tremor. A competing idea was proposed by Aki & Koyanagi (1981) consisting of a stationary model comprising a pre-existing chain of cracks, filled by magma and connected by narrow channels; these will then resonate in response to a constant excess of pressure. In this way, the crack dimensions are likely to be critical in generating the characteristic frequency of tremor. This model also assumed a periodic excitation of the crack of different mode, as proposed by Aki et al. (1977b) and Chouet (1981), who assumed random jerky opening of the channel between cracks.

The abrupt impedance contrast at the rock-fluid interface generates interfaces waves at low frequencies (e.g. Aki et al., 1977b; Chouet, 1988; Neuberg, 2000), called crack waves, explaining the much longer than expected resonant period of a fluid-filled crack. Their existence, demonstrated analytically by Ferrazzini & Aki (1987) and experimentally by Tang & Cheng (1989), suggests that the source size is reasonably small if compared to the dimension of the magma chamber (Jousset et al., 2003).
Figure 2.7: Schematic model of a fluid-filled crack. The shaded area in the centre of the crack marks the zone where the fluid provide the pressure transient which is applied on both walls of the crack. Re-drawn from Chouet (2003).

The composition of the fluid involved can be determined by studying the dominant frequency and the quality factor ($Q$) of the damped coda of LP events (Chouet, 2003), as the coda describes the damping oscillations of specific modes (Aki & Richards, 1980). As $Q$ increases with the impedance contrast, the observed high values of $Q$ are generated if gas bubbles are present in the fluid which act to increase the impedance contrast between the fluid and solid. These gas pockets reduce the sound speed of the fluid (Kieffer, 1977), the speed of the crack wave, and consequently lower the dominant frequency (Chouet, 1992). Alternative theories on the meaning of the variation of $Q$ and dominant frequencies were postulated by Jousset et al. (2013) and Tary et al. (2014): the former associated them to the excitation of different resonators, the latter to the physical state and fluid flow respectively.

In the context of a fluid-filled crack, several phenomena were taken into account as the initial trigger of the excitation of the crack. These include depressurisation during eruptive activity (Chouet, 1996), sudden pressure decreases (Neuberg, 2000), repeated collapse of the crater floor that in turn generate transient pressurisations (Falsaperla et al., 2002), compound choking of the flow triggering acoustic oscillations of the liquid/gas mixture (Chouet, 2003), degassing processes (Waite et al., 2008; Cauchie et al., 2015), and the rapid discharge of hydrothermal fluids (Zecevic et al., 2013).
For the shallow tremor events, magma degassing that creates a sustained pressure oscillation is easily invoked. However, tremor occurring at depth of 30-60 km needs another mechanism as free gas pockets are unlikely to be present as such depth (Chouet, 1992). This is explained by Aki & Koyanagi (1981): assuming that magma transport occurs aseismically at great depth, deep tremor may be generated when magma passes through some stiff channel or barrier. The presence of such a barrier is also evident during eruption: the start of unrest often commences with a vigorous tremor signal due to stiff channel barriers, later exhibiting lower amplitude tremor/ LP events as these barriers became weaker (Chouet, 1992).

The role of the magmatic pressure is also paramount in the model of Julian (1994), who developed a tremor model where an increasing pressure in the steady flow regime, and initially with no oscillations, passes to simple and complex oscillations causing harmonic tremor, and finally to chaotic oscillations. Channel geometry and compliance strongly affects the efficiency of flow at exciting such oscillations. Above a critical threshold, flow speed may generate self-excited oscillations, which in turn cause tremor. Below this threshold, LP signals may form as a response of an external disturbance, which create decaying oscillation that returns to the steady state.

In addition to fluids, the brittle failure of magma moving through the glass transition has been proposed by Neuberg et al. (2006), where magma pressure transients are the result of high strain, high viscosity at the conduit wall (Fig. 2.8). At the wall of a conduit, loss of heat and gas occur resulting in viscosity gradients, in turn leading to the build-up of shear stress that eventually causes the magma to fail in a brittle manner.
In recent years another source mechanism, where fluids are not actively involved, has also been taken into account to explain the origin of LP events. McNutt (2005) suggested that some signals may be normal earthquakes, but featuring rupture over a much slower timescale. It was this concept that was taken by Bean et al. (2008; 2014). Here, the path effects have been found to exert a significant influence on the LP seismicity (Bean et al., 2008), and going further still, Bean et al. (2014) suggested that weak brittle signals may generate LP events if deforming in a largely ductile deformation field. In particular, they proposed that VT events morph to LP signals when weak low-stiffness materials promote slower rupture speeds. This implies that some LP events (but excluding Tornillo) are caused by brittle failure where no fluid-driven source model is required (as well as VT events), forming altogether a stress-driven mechanical class. They finally suggested that these swarms of pulse-like LP signals are due to failure in material close to brittle-ductility transition in shallow volcanic materials, primary controlled by low internal friction angle rather than high pressure/temperature conditions.

Hybrid events are mainly viewed as a result of the combination of brittle failure zone and excitation of a fluid-filled crack because of their features shared with both VT and LP events (Lahr et al., 1994; Chouet, 1996). In particular, the opening of a crack (high-frequency onset) generates a pressure gradient which drags the fluids into it (low-frequency coda). However,
White et al. (1998) argued that the periodicity and similarity of HB signals occurring in swarms may be caused by violent degassing into adjacent fractures.

The generation mechanism of Tornillos has been debated since the late 1990’s (Alparone et al., 2010; Lesage & Surono, 1995). Here, a relatively simple source model accounting for the monochromatic features was proposed, while Julian (1994) consider them an intermediate signal between LP events and tremor, stating that they do not require special explanation. In fact, source effects are invoked for the consistency of the frequencies at all stations, similarly to LP signals (Chouet, 1992; Gomez & Torres, 1997). More recent idea focus on the free vibration of a fluid-filled crack in response to a pressure transient (Chouet, 1992; Alparone et al., 2010), as already observed for LP events and tremor. However, to justify the higher quality factor observed for the Tornillos (up to 400, Milluzzo et al., 2010), a higher gas fraction (Chouet, 1992; Kumagai & Chouet, 2000) or a smaller aspect ratio of the crack (Kumagai & Chouet, 2001) has been postulated, allowing a larger velocity contrast between the fluid and the crack. The amplitude modulation could be explained in terms of either slow waves along the interface and the cracks, which is reflected back at the end of the crack (Sturton & Neuberg, 2006), or beating, which describes the alternating constructive and destructive interference of two or more waves at different frequencies, with the frequency of modulation being the difference between the two initials frequencies (Milluzzo et al., 2010).

The direction of the first motion observed on the VLP signals indicates the addition of mass, and upward migration of magma (upward first motion). Alternatively, a downward first motion points towards a mass loss and gas release (Jousset et al., 2013). In fact, Ohminato et al. (1998) assumed that in a volcanic context, volume changes associated with magma transport and degassing play fundamental roles in the source processes and analysing the recorded events, the authors interpreted the rising part of the signal as a slow accumulation phase of magma and a gas pressure build-up in crack-like source (Fig. 2.9a), while the second part was linked to a rapid deflation phase (Fig. 2.9b). Because of this and their long period feature, VLP events may be used to map the conduit structure and to resolve mass transport budget (Chouet, 2003). A close relationship has been found between VLP and LP signals where a simultaneous occurrence was observed by Waite et al. (2008) who suggested that the VLP events are a passive response of the magmatic system to the active LP mechanism, which is thought to be steady rates of degassing and
crystallization. Conversely, Saccorotti et al. (2007) proposed that an LP event is caused by the generation of a VLP signal. In particular, LP event forms in response to the intrusion of a gas-rich phase in a gas-poor phase, while VLP signal is caused by movement and decompression of the gas slug itself. Cannata et al. (2009) also suggested a causal relationship between them, but rather than a common source, LP and VLP events have different connected sources.

Figure 2.9: Schematic model of a separated gas-liquid flow in a crack in the proximity of a nozzle. (a) phase of magma accumulation and pressure build-up; (b) phase of rapid deflation. The gas slug is stationary in front of the nozzle (Mach number, $M_g < 1$), while the gas flow is choked inside it ($M_g = 1$). After the nozzle the flow is supersonic ($M_g > 1$). Re-drawn from Ohminato et al. (1998).

2.4. MOMENT TENSOR SOLUTION

To understand the genetic process behind an earthquake, it is useful to derive a source mechanism. The simplest representation of this is the classic view of a fault plane solution, which requires the polarity of the first P-wave arrivals recorded on the seismograms. The polarity of the arrival indicate whether the first motion is compressional (positive arrival) or dilatational (negative arrival). Once the earthquake is localized, azimuth and plunge of the ray-path between source and each seismometer are calculated. By plotting each polarity on a lower hemisphere stereonet, a graphical representation of the motion of the fault may be derived, often represented by the classic beach ball type diagram. For a shear-faulting process, the polarities can be divided by two
orthogonal planes, in which only one plane represents the fault plane, while the other is called the auxiliary plane. Without any other information (e.g. visual inspection) it is impossible to distinguish them (Pettitt, 1998). For such cases, the forces occurring at the source are modelled as a pair of complementary couples, or double couple (DC, Fig. 2.10, centre), which produce no net torque. The majority of earthquakes are modelled as DC (Shearer, 2009).

However for non-DC sources, the fault plane solution give ambiguous information and a more sophisticated method, called moment tensor (MT) solution, is needed. MT solution requires both amplitude and polarity of the first P-wave arrival and corresponds to a 3x3 matrix (Pettitt, 1998), where each component represents a force couple (Shearer, 2009). By decomposing the MT solution, an isotropic (ISO, Fig. 2.10, left) and a deviatoric component are extracted. The ISO component is a measure of volume change, which is zero for simple shear process (Shearer, 2009). The deviatoric part is in turn decomposed into a best-fitting DC and a second term called compensated linear vector dipole (CLVD, Fig. 2.10, right). Therefore the MT solution can be expressed in terms of its source components as:

\[
\text{MT} = \text{ISO} + \text{DC} + \text{CLVD} \quad \text{(Eq. 2.1)}
\]

A MT solution is useful for better describing the relative contribution of pure brittle (mechanical) movement (such as DC), compared to the processes that generate or destroy volume, as derived from ISO or CLVD components. Although difficult, this have been achieved both in the field (Pettitt, 1998) and the laboratory (Benson et al., 2008).
2.5. ERUPTION FORECASTING

As discussed earlier, the use of VT and other seismic earthquake swarms has been heralded as a new tool for the forecasting of volcanic unrest. This is possible as the final approach to eruption is commonly preceded by accelerating occurrence rates of both VT and LP events (e.g. Tokarev, 1971; Malone et al., 1983; Swanson et al., 1983; Voight, 1988; Kilburn, 2003).

The importance of the increase of number and energy of volcanic earthquakes in eruption forecasting dates back to 1970s (Tokarev, 1971; Malone et al., 1983), with a first quantitative method described by Voight in 1988. The Voight method (1988) for prediction of volcanic eruptions is based on the fundamental power-law for fracture in brittle materials, known as the Failure Forecast Method (FFM). The FFM can be applied with any observable quantity describing the behaviour that precede a volcanic eruption (e.g. seismic, tilt and displacement data) and relates the logarithmic of rate and the logarithmic of acceleration of the observable quantity through two empirical constants, $A$ and $a$. Under certain circumstances the time of eruption is simply the time of failure which can be graphically evaluated by using the reciprocal-rate curve, where the inverse
rate decreases continuously in time. For volcanic edifices $\alpha = 2$, making the inverse rate linear so that the time of failure lies at the intersection between the inverse rate and the time axis in the reciprocal-rate curve plot (Fig. 2.1).

![Rate and Inverse Rate vs Time](image)

*Figure 2.1: (left) rate and (right) inverse rate against time for the change in length of the dome at Mount St Helens. Different lines represent different values of $\alpha$. Note that for $\alpha = 2$, the rate tends to infinity and the inverse rate decreases linearly to intercept the time axis. From Voight (1988).*

The model of Kilburn (2003), following the study of Voight (1988), linked the change in peak event rate to the development of the major pathway, generated by progressive (but not necessarily forward) coalescence of existing fractures, allowing for forecasting the time of an eruption. This highlighted the importance of distinguishing between LP and VT data, as only the latter reveal information on the fracturing processes. In particular, Kilburn (2003) observed that, though the non-cumulative VT rate shows an accelerating overall pattern of seismicity, oscillations occur where the accelerating trend of the seismic rate were associated to fault extension and coalescence while the decelerating one to the energy dissipation at coalescence. In addition, while Voight (1988) assumed a constant value of $\alpha$ throughout the precursory trend (meaning a linear inverse rate), Kilburn (2003) argued that VT inverse rate follows a curvy trend due to the increasing $\alpha$, from 1 to 2, as the eruption approached. This method works well due to the scale invariance of rock fracturing, permitting a scale fracturing law at small scales to take the same form of those at large scales. However while this method is applicable to all cases when a new pathway leads to an
eruption (e.g. Mt. Pinatubo, Philippines, Fig. 2.12), it would be less useful in open-vent conditions (e.g. 1992 second eruption at Mt. Spurr, USA, McNutt, 1996) or in volcanoes with short repose intervals (e.g. Mt. Etna, Italy). On a following paper, Kilbun (2012) extended this model to the earlier stages in precursory sequences, showing that an exponential trend characterizes the deformation process at constant compression rate up to about 90% of the strain at failure.

![Figure 2.12: Inverse-event rate vs. time diagram shows the change in recorded seismicity before 1991 eruption on Mt. Pinatubo. Open triangles connected by small dashed lines represent sequences of energy dissipation, while filled triangles linked by large dashed lines represent fault extension and correspond to the peaks in the event rate (i.e. the minima in the inverse rate curve). The arrow E marks the onset of the eruption. In this occasion, this model could have provided a 48-h warning of eruption. From Kilburn (2003).](image)

Based on the FFM, several successful forecasts were issued (e.g. Pinatubo eruption in 1991, Newhall & Punongbayan, 1996). However, in recent years, the standard FFM has been supplemented by new theories. Bell et al. (2011) proposed the Generalized Linear Model (GLM), which takes into account a non-Gaussian distribution of earthquakes occurrence uncertainties, yielding greatly reduced error in the forecast as the sequence proceed, and the convergence of the forecast to the true value if compared to larger variances or an earlier forecast time of failure of the FFM. Both FFM and GLM were questioned by Bell et al. (2013), who stated that both methods requires regressions on bins of data, therefore leading biased forecasts. The authors then suggested to use the Maximum Likelihood (ML) estimation to analyse a point process like the earthquake
occurrence, which provides the most accurate, least biased forecasts compared to the FFM and the GLM.

All of the above methods, however, rely on the use of VT events in order to generate a forecast. As LP seismicity is thought to be a sign of pressurization in a magmatic/hydrothermal system, its rate can be related to the rate and magnitude of pressurisation and to the intensity of explosive activity (Chouet, 1996). For example, Chouet et al. (1994) observed that the eruptions at Redoubt Volcano in December 1989 were preceded by the onset of LP activity for 23 hours, after which graded into tremor, and no VT swarms. Hammer & Neuberg (2009) applied the FFM to the average seismic rate as applied to LP swarms preceding a dome collapse at Soufriere Hills (Montserrat) in June 1997, finding improving estimates of the time of failure as more swarms were added. Here the increase of event rate is likely as a result of accelerated magma ascent. Despite these cases, the use of LP data in forecasting appears to be difficult, with Saccorotti et al. (2007) and Cannata et al. (2009) finding a general lack of correlation between the LP-generating process and the renewal of effusive activity on Mt. Etna during the 2004-2005 eruption.

Because tremor is considered a short-term precursor and usually accompanies an eruption rather than precedes it (McNutt, 1996), long-term forecasts based solely on tremor are rare. For example, it was observed that the eruption at Izu Oshima (Japan) in 1986 was preceded by 7 months of increased seismicity and 4 of tremor, with a shift from banded tremor to continuous tremor with an increasing energy release rate some 3 weeks before the eruption (McNutt, 1996). However, a more recent study correctly forecasted (in hindsight) four of five eruptions at White Island volcano (New Zealand) between 2011 and 2014, modelling the amplitude of tremor with the FFM (Chardot et al., 2015). This illustrates the variability and difficulty in applying LP data to pre-eruptive conditions, further reinforcing the need to better understand the physics that generate these events.

Finally, the use of Tornillos did not have a great success in forecasting either. While Gomez & Torres (1993) and Stix et al. (1993) found that all but one of the eruptions occurring at Galeras during 1992-1993 were preceded by tornillos, showing a decrease in dominant frequency and increase in duration, at other volcanoes these events were recorded during, after or even in periods of quiscence (Gomez & Torres, 1997).
2.6. WHAT IS MISSING FROM THIS PLETHORA OF STUDIES?

What emerges from the several cited studies in the previous section is the lack of consensus about the origin of all volcanic earthquakes, but VT events. Volcanic tremor and LP signals have been recognized since the 70’s, yet their source mechanism is under investigation. Of particular note is the case of LP events, where instead of pointing towards a single univocal interpretation, multiple theories involving a diverse range of processes have been presented over the years. Whether these theories are all valid or not, it is unlikely that signals having the same characteristics (both in time and in frequency domain) comes from different sources. It is likely, however, that subtle differences exist requiring extra care when calling signals with the same source-based term.

VLP signals and Tornillos have the alibi that they have been recognized about 20 years ago: the former due to technical limitations, the latter because confused with LP events. Tornillos in particular have been rarely debated (likely due to their rare occurrence), even though they show unique characteristics, such as long duration, amplitude modulation effect, narrow spectral peaks, which make them a clear-standalone class of seismic signals.

The greatest limitation in studying such phenomena is the lack of knowledge about the underlying physical processes. However, small-scale laboratory experiments have been proven to provide hints, if not the solution, to geological problems. In fact, in a lab environment, parameters such as temperature and pressure (which for the macro-scale case are unknown) are measured and controlled and the origin of geological/seismological events better understood as a result. The use of laboratory experiments in improving our knowledge in volcano-seismology is presented in chapter 3.
3. LABORATORY EXPERIMENTS

3.1. ROCK MECHANICS TESTING

While some materials, such as metals, have highly standardised procedures for strength and other mechanical tests, and which are considered constant for all samples, the mechanical properties of rock depends on a large number of variables due to its inherent inhomogeneity (e.g. mineral composition, grain size) that can even vary between samples taken from the same block of rock. Therefore, laboratory experiments are fundamental in rock mechanics studies (Jaeger et al., 2007). “The complete ISRM suggested methods for rock characterization, testing and monitoring, 1974-2006”, edited by the International Society for Rock Mechanics, specifies a large number of standard procedures to run rock mechanics tests (Jaeger et al., 2007).

In a hydrostatic test, the rock specimen undergoes a uniform hydrostatic stress, such as the stresses at the three orthogonal direction ($\sigma_1$, $\sigma_2$, and $\sigma_3$) are all compressive (positive) and equal and no shear is applied:

$$\sigma_1 = \sigma_2 = \sigma_3 > 0$$  \hspace{1cm} (Eq. 3.1)

The specimen is placed in a pressure vessel and it is surrounded by a pressurised liquid or gas which builds up the confining pressure (Fig. 3.1a). The purpose of this type of test is to determine the bulk modulus of the rock and poroelastic parameters (e.g. bulk compressibility and pore compressibility) (Jaeger et al., 2007).

A uniaxial test consists of a compression of a rock specimen between two rigid platens (Fig. 3.1b), and is normally run until the failure of the specimen to calculate Young’s modulus (E), the unconfined compressive strength (UCS) and the Poisson ratio ($\nu$) of the specimen (Jaeger et al., 2007). As the name suggests, stress is only applied in one direction, such as:

$$\sigma_1 > \sigma_2 = \sigma_3 = 0.$$  \hspace{1cm} (Eq. 3.2)

Finally, in a triaxial test (the type used in this study), all stresses are compressive with one stress greater than the other two (which are equal in conventional triaxial cases to simplify the engineering requirements) as in:

$$\sigma_1 > \sigma_2 = \sigma_3 > 0,$$  \hspace{1cm} (Eq. 3.3)

This represents a state of stress that suits subsurface conditions for most cases. However, having $\sigma_2 = \sigma_3$, known simply as confining pressure, may be considered as an experimental
limitation rather than true conditions. As for the hydrostatic test, the specimen is placed inside a pressure vessel, surrounded by pressurised fluid and loaded by a piston (Fig. 3.1c), which generates the differential stress ($\sigma_1 - \sigma_3$) (Jaeger et al., 2007). In this study this method is used to impose a shallow, but not surface, volcanic condition of pressure.

Figure 3.1: Schematic diagram of a (a) hydrostatic, (b) uniaxial and (c) triaxial test.

3.2. ACOUSTIC EMISSIONS

An earthquake is the shaking of Earth’s surface associated with the energy release caused mostly, but not only, by the movement of a fault due to shear stress. Measuring this intensity is achieved via numerous scales such as the Mercalli intensity scale (Mercalli, 1902) and Richter magnitude scale (Richter, 1935). The most commonly used scale is currently the logarithmic Moment Magnitude scale, defined by Hanks & Kanamori (1979) as:

$$M_w = \frac{2}{3} \log_{10}(M_0) - 10.7$$  \hspace{1cm} (Eq. 3.4)

Where: $M_0$ is the seismic moment, defined by Aki (1966) as:

$$M_0 = \mu \bar{u} S$$  \hspace{1cm} (Eq. 3.5)

And where $\mu$ is the shear modulus, $\bar{u}$ is the average displacement along the fault and $S$ is fault surface.
Equations (3.4) and (3.5) show a correlation between fault dimensions and magnitude of the earthquake: the bigger the fault, the higher the magnitude. At smaller scales (cm), the energy radiated by micro-fracture has been recorded in the laboratory where it is known as Acoustic Emission, and so-named as they could be heard by the human ear in the early studies in the 1960’s. Acoustic emissions (AEs) are characterized by having a small amount of energy \((M_{w} < -4)\), small source dimension (< 10 cm), high frequency content (> 10 kHz) and propagation of few centimetres (Pettitt, 1998). More recent work, recognising this frequency content, sometimes refers to these events as nano-seismicity (Selvadurai and Glaser, 2015), and at the metre scale events are often referred to as micro-seismicity (Collins et al., 2002). In this work, the historical convention is used, referring to these signals as AE.

Due to their characteristic frequencies, AEs are generally recorded by ultrasonic piezoelectric transducers (PZTs) which convert the mechanical energy generated by the elastic wave into an electrical signal (voltage) that is pre-amplified by a factor ranging from 10-1000, and subsequently digitised by a dedicated PC-based recorder (e.g. Benson et al., 2007). The inverse process also occurs: PZT can be stimulated by high voltage pulse to generate a mechanical pulse. In this way, the PZT sensor can be used either to record the microcracking (so called passive AE) occurring in a sample undergoing deformation, or to generate P-wave and S-wave elastic waves (which here are called surveys and used for P-wave only).

### 3.3. Early Research Using AE as a Laboratory Tool

One of the earliest studies involving the recording of AEs generated during rock deformation experiment was carried out by Mogi (1962), who suggested that crustal deformation may be scaled by using laboratory fracture experiments. The author applied static stress to different types of rock and measured what was described as elastic shocks (now known as AEs) using 2 cartridge-crystal transducers. One of these was attached to the specimen and one placed on the ground to record external noise. It was found that the onset of AE activity occurred at a particular stress state, with activity increasing with stress and heterogeneity (in particular before failure), even when the sample is at constant stress regime. Under such conditions, the temporal distribution of AEs matches an exponential distribution. When investigating the magnitude distribution of AEs, Mogi (1962) found that heterogeneities in the Earth’s crust are likely to be the foci of shallow volcanic
earthquakes. In addition, the stress state of a region could be estimated using the Omori frequency distribution of the earthquakes.

Scholz (1968a) improved the recording of micro-fracturing by using newly developed piezoelectric transducers composed of a disk of barium titanate, which have a relatively flat broadband frequency response. In a related paper, Scholz (1968b) noticed that the location of AE events, generated during a uniaxial test on granite, showed evidence for clustering of AEs around the eventual fault. This suggested a coalescence of microcracks at a certain level of stress (Fig. 3.2), coalescing to form a macro-scale fault. However, the fault did not develop from one end of the specimen to the other, but rather a clustering around a point of weakness.

![Figure 3.2: different point of views of a rock sample showing the clustering of the AE locations (circles) around the eventual fault (dashed lines) in (a) static and (b) dynamic cracking region. From Scholz et al. (1968b).](image)

Ohnaka & Mogi (1982) further elucidated on the mechanisms of deformation by recording AE with sensors of different frequency response. Based on a series of uniaxial experiments, five different stages for microseismicity evolution was proposed: 1) compaction at lower stresses causing closure of large cracks, producing signals at 20-400 kHz (low frequency (LF) events), and rupture of asperities, generating signals at 1-1.5 MHz (high frequency (HF) events), 2) all pre-existing cracks are closed, with the stress not high enough to create new fractures, a minimum of
AE activity is seen, 3) stress-induced cracks start to grow as the level of stress increasing causing an exponential build-up of the seismic activity with a constant growth of both LF and HF signals, 4) the AE activity grows supra-exponentially, with a greater increase of LF events due to the formation of larger cracks or dilatancy that attenuates the higher frequencies; 5) failure is anticipated by a rapid acceleration of HF events that continues immediately after. In addition, Ohnaka & Mogi (1982) suggested that the frequency of AEs does not depend on the applied stress, but rather on the change of microstructure within a deforming rock sample, and that the seismic activity starts well before any sign of dilatancy has appeared, therefore disproving the hypothesis of Scholz (1968a). Similarly to the work of Ohnaka & Mogi (1982), Read et al. (1995) carried on triaxial experiments on Darley Dale sandstone under water-saturated conditions, finding a marked shift from lower-amplitude, higher frequency events to higher-amplitude, lower frequency events recorded both pre- and post- peak stress. This behaviour was correlated to the rapid linkage and coalescence of previously isolated dilatant cracks, manifested macroscopically as the transition between strain-hardening and strain-softening. However no increase of HF activity was observed before the failure.

The link between AE activity and microstructure changes (Ohnaka & Mogi, 1982) also leads to the notion that even the linear elastic part of the stress-strain curve is characterized by small cracks that do not deform in the elastic regime, proposing that this section of the stress-strain evolution should be better called pseudoelastic. Related to this concept, the distribution of earthquake magnitudes, where a large number of smaller earthquakes occur more often than larger events, can be quantified by the Gutenberg-Richter frequency-magnitude distribution:

\[ \log N = a - bm \]  \hspace{1cm} (Eq. 3.6)

Where \( N \) is the number of earthquakes with magnitude equal or above \( m \), \( a \) is a constant and \( b \) is the seismic \( b \)-value (Shearer, 2009). This is valid at all scales, and for the particular case of AE signals recorded during rock deformation experiments, the magnitude term can be replaced by the amplitudes of the AEs (Sammonds, 1999), making this a very important scale-invariant analysis tool.

The temporal variations of the \( b \)-value, calculated using AE events during laboratory deformation, was studied by Meredith et al. (1990). During triaxial tests they found that the frequency magnitude distribution was a function of stress history with a double minima in the \( b \)-
value anomaly for the most realistic case, i.e. where dynamic failure is preceded by strain softening. The $b$-value minima are associated with short-term quiescence, because few larger fractures (therefore producing higher magnitude earthquakes) dominate the stress relaxation impeding the formation of several smaller cracks (Main & Meredith, 1991). Main and Meredith (1991) also reported that during this short-term quiescence, the stress intensity factor ($K$) increased despite a decrease in both $b$-value and applied stress. Similarly Sammonds et al. (1992) found an inverse correlation between $K$ and $b$-value throughout their triaxial experiments, but observed a double $b$-value minima only when the pore fluid volume was kept constant. This behaviour has been linked to the onset of dilatancy, as a result of pore pressure decay and causing a prolonged phase of strain softening. When combined with the decrease in axial stress and relaxation of the stress intensity, a net effect of a delay in the major shear zone is observed despite crack growth continuing. A double minima trend is not observed for the experiments run at dry and constant pore pressure conditions (Fig. 3.3).
Figure 3.3: Stress-time curve (continuous line) and evolution of the b-value (discontinuous line) for the triaxial experiments at (a) dry conditions, (b) constant fluid volume and (c) constant pore pressure. Only the case at constant fluid volume shows double minima in the b-value. Modified from Sammonds et al. (1992).
To better understand fault nucleation and growth in the dynamic phase of the stress release that accompanies failure, Lockner et al. (1991) conducted experiments on Westerly granite using the AE hit rate as a control / feedback system, to slow down the fracture process as the dynamic rupture state was approached. The fracture nucleated in the sample centre, finally forming a shear failure plane, after which the AE was greatly reduced (quiescent phase). During these experiments, a minimum in the seismic $b$-value occurred at the time of fault nucleation, followed by a recovery to about half of the initial value and, studying the energy release, portions of the sample characterized by pre-existing strength heterogeneity may be indicated by the preferential direction of propagation of the fault.

Although fault plane solutions are a common tool to describe the sources of the AEs, Moment Tensor (MT) inversions are quite rare. As noted in the work of Pettitt (1998), this is due to i) the response of the transducers not being well calibrated and different sensors behave differently each other, with complicated frequency responses; ii) sensors and wavelength share similar dimensions, yielding to azimuthal dependency at high frequencies; iii) the coupling between sensors and rock significantly affects the recorded amplitudes. The author concentrated on the MT inversions on both laboratory and in situ microseismic data, developing an effective procedure to investigate the source of microcracks. The author focused on the azimuthal amplitude dependence showing that AEs have complex source mechanics with large volumetric components, making them true scaled-down earthquakes.

### 3.4. SIMULATION OF VOLCANIC CONDITIONS IN LABORATORY

The last decade has seen an increased interest in volcano-seismology, in particular the simulation of volcanic conditions in controlled laboratory experiments with the aim of better understanding mechanical and acoustic properties of fluid-saturated volcanic rock samples. In particular, several rock deformation studies (e.g. Vinciguerra et al., 2005; Benson et al., 2008; Heap et al., 2011) have been carried out on an effusive porphyritic alkali basalt from Mt. Etna (hereafter EB, Etna Basalt), for its particularly isotropic distribution of microcracks from thermal cooling (Vinciguerra et al., 2005, Stanchits et al., 2006).
3.4.1. **HYDROSTATIC TESTS**

Hydrostatic tests were performed to investigate seismic velocities and AE features at high pressure conditions. This is important, as Vinciguerra et al. (2005) found a direct correlation between confining pressure and seismic velocities, and an inverse correlation between pressure, permeability and porosity. The presence of microcracks is key to this behaviour, and gives rise to the low value for seismic velocity at low pressures in EB (3.05-3.45 km/s), and consistent with the low-velocity zone observed for the basaltic pile of lava flows at Etna.

Elastic-wave tomography obtained for a saturated, thermally-stressed tuff from the Campi Flegrei Caldera (Italy) showed a general agreement with the velocity structure of the caldera itself (Vinciguerra et al., 2006). This suggested that the gas-saturated or water-saturated conditions for the caldera materials, driven by inelastic pore collapse and thermal cracking was manifested by a significant change of velocity structure at depth, thus demonstrating the power of laboratory experiments.

However, the direct role of temperature had not been fully understood. Burlini et al. (2007) simulated the intrusion of a molten layer into a synthetic olivine aggregate. Here, the amplitude of the seismograms recorded during P-wave velocity measurements decreased with increasing temperature, suggesting a modification of the acoustic properties of the molten layer. The authors associated the AEs from thermal cracking with volcanic LP earthquakes, while the long duration tremor was linked to the migration of the molten layer into the cracks. Finally, shear failure events have been compared with volcano-tectonic earthquakes. Burlini et al. (2009) focused on the dehydration reactions of lizardite, a common alteration product of oceanic lithosphere, which was then inferred to trigger AEs under laboratory conditions. Simulating subduction zone pressures and temperatures, a cascade of multiple and overlapping LF events was recorded, similar to real seismic tremor, and above the equilibrium dehydration temperature.

3.4.2. **UNIAXIAL TESTS**

Cyclical pressurisation and depressurisation of the plumbing system causes an important effect on volcanic edifices, known as stress cycling. Heap et al. (2009) simulated these conditions, with EB, showing that stress cycling effect lowers the Young’s modulus and increases the Poisson’s ratio. It is accompanied by the Kaiser effect, referring to the micro-seismic activity that re-
commences during any loading cycle at the same level of stress at which it ceased during the previous unloading cycle. These data suggest that extra damage to the volcanic edifice, with changed mechanical and seismic properties, may additionally be the result of stress episodes accompanied by volcanic earthquakes (Heap et al., 2009).

Similarly to the work of Burlini et al. (2007) at hydrostatic conditions, Benson et al. (2012) and Lavallée et al. (2012) focused on the interaction between a melt confined in a conduit and the surrounding country rock. Benson et al. (2012) studied the role of temperature on melt viscosity, relaxation, and the delay between rock failure and the number of cycles of acceleration/deceleration required. Lavallée et al. (2012), instead, pointed out how the magma fractured at a strain rate much lower than that one expected with the Maxwell relation for viscoelastic relaxation time (Fig. 3.4).

![Figure 3.4](image.png)

**Figure 3.4:** fractures in the magma after the uniaxial test, as imaged by optical microscope in right bottom image. From Lavallée et al. (2012).

A series of high-temperature, uniaxial experiments on synthetic samples were recently carried out by Vasseur et al. (2015), highlighting the role of microstructural heterogeneity on the style and mechanism of the deformation as well as on the accuracy of the FFM. The results showed that failure is achieved with less strain and stress in high porous material than in low-porosity materials, and requiring less energetic AE increments and rate of AE energy increase. In addition the FFM are more accurate and $b$-value decreases abruptly before failure for highly heterogeneous samples.
3.4.3. TRIAXIAL TESTS

The high level of pre-existing cracks inferred in many volcanic rocks due to cooling cracks (e.g. Vinciguerra et al., 2005; Benson et al., 2007) is thought to be responsible for the randomly-distributed, pore-collapse type AE activity during the initial compaction stages of triaxial deformation. A strong P-wave anisotropy is also derived, driven by the preferential closure of cracks parallel and normal to the compression direction (Stanchits et al., 2006; Benson et al., 2007) which generates a particular Horizontal Transverse Isotropy (Benson et al., 2007). With increasing axial stress and decreasing confining pressure, sources of AE became predominantly tensile-type, but close to failure shear-type events became more important (Stanchits et al., 2006).

Aker et al. (2014) observed the same behaviour in the Vosges sandstone, with the DC component and the seismic moment increasing with the increasing differential stress, with the isotropic component decreasing. The majority of the events were generated by opening fractures, and while small seismic moments are indicative of a tensile mechanism, larger moments are generally related to pure DC. The authors observed 4 stages of AE activity: 1) little activity until a certain stress/strain threshold was achieved; 2) a slow, linear, increase; 3) a continuing linear increase at a steeper gradient, and; 4) an accelerated phase related to the larger scale macroscopic failure (Fig. 3.5).

Figure 3.5: volumetric strain-stress curve (thick line) and cumulative AE events over stress curve (dots), showing the four stages of AE activity. From Aker et al. (2014).
This general trend has been well documented, for example during stress cycling tests (Heap et al., 2011), showing that the deformation will tend to accelerate to failure once a critical level of damage is reached. This result suggests that the seismicity following major intrusive episodes and leading to eruptions at Mt. Etna volcano between 1993 and 2005 can be explained in terms of a time-dependant brittle creep mechanism.

However, the majority of triaxial deformation experiments have to date focused on recording AEs during rock fracture of either dry and water saturated samples. In the context of this study, and with particular focus on the AE frequency content and source mechanism during the coupled rock-fluid mechanics, experiments on EB under pore-saturated conditions (with water) were conducted (Benson et al. 2007). Here, an AE rate acceleration and strengthening was seen as the deformation proceeded, with the nucleation of the major fault imaged by the location of the micro-seismic activity.

To more precisely investigate the nature of the rock-fluid coupling, Benson et al. (2008) rapidly decompressed the sample pore space volume after the failure, resulting in the rapid movement of fluids (stored inside the pore space) via a shallow centrally pre-drilled conduit. During the compression phase the majority of events (mainly HF) originated within the damage zone of the shear fault or its conjugate, with mainly a DC focal mechanism (as expected). However, in the decompression phase AEs had the same location as before but with a high volumetric component and a low shear one, suggesting that these events, analogous with volcanic LP signals, were formed by a fluid flow through a tortuous pathway (Fig. 3.6). The authors found also HB events, which were related to migration of fluids into the fracture as soon as the fluid-filled rock faulted.
Following this, Benson et al. (2010) focused on the transition between HF, LF and HB events during the initial deformational phase. While almost all of the AEs in dry samples belong to the HF signal type, significant numbers of HB events were recorded in the saturated sample, suggesting that the HB events are the general form of seismicity produced by the shearing and fracturing of magma, whereas HF and LF signals are the end-members. When the network of cracks and fractures is sufficiently well developed, HB events become LF signals. De Rubeis et al. (2010) used a statistical approach to classify AEs in terms of their spectra, finding that LF signals occur even during the deformation phase as a result of fluid migration into the growing fractures. Harrington & Benson (2011) sought to better understand the HB events in terms of source mechanism by investigating the changes in scaling between seismic moment and spectral corner frequency in presence of water, concluding that if moment and corner frequency scale as brittle-failure mechanism then fluid process in not involved.

Finally, to better understand the transition between purely liquid-driven events and gas-generated events, a pilot study was conducted at elevated temperature. Under water saturated conditions, a switch from LF to VLF signals was seen at the boiling pressure of water, suggesting that the switch was possibly caused by the change in fluid viscosity (water to steam) in generating different classes of low frequency activity on active volcanoes. This was subsequently verified by using dry Nitrogen gas as the pore fluid. In both tests, AEs started at almost the same pressure,
suggesting the presence of a threshold pressure or pressure change rate affecting the onset of AEs (Fig. 3.7). Although this represents the first study where a gas phase has been used to generate VLF activity, supporting the theory that such seismic activity may be driven by gas movements inside the volcanic edifices, a controversial aspect arose. Kumagai and Chouet (2000, 2001) studied the theory behind the behaviour of a fluid-filled crack when excited by the movement of different fluids (in terms of phase and composition), finding that a gas-rich fluid is responsible for a higher resonating frequency having a narrower bandwidth than a liquid-rich fluid. While a narrower bandwidth is effectively observed both after the phase transition and during the Nitrogen release in Figure 3.7a and 3.7b, an increase in resonating frequency is missing. This experiment subsequently formed the basis of the current study, where fluids of different phases and compositions are been used to produce fluid-induced seismicity and clarify the link between LFS and the characteristics of the fluids.

![Figure 3.7: pore pressure releases at (a) water and (b) nitrogen saturation conditions. (Below) Seismograms and (above) associated spectrograms showing the transition from low-frequency to very-low-frequency occurring at pore pressure = 2 MPa (black line), in water saturation conditions. From Benson et al. (2014).](image)

3.5. SCALING LAWS FOR AEs

One of the great challenges in using laboratory rock physics as a tool is the scaling to the field scale. Many volcanic eruptions are preceded by increasing rate of earthquakes (e.g. Tokarev 1971; Voight 1988) in a similar way rock failure occurs during triaxial rock deformation. Therefore, forecasting models based on the increasing of seismic activity (or via AE, in the laboratory context) can be performed and correlated (e.g. Kilburn, 2003; Vasseur et al., 2015).
Equation (3.6) introduced the concept of magnitude distribution of earthquakes, which is controlled by the seismic b-value. With some basic assumptions about earthquakes (constant stress drop), the b-value and fractal power law exponent (D) can be related by

\[ D = \frac{3b}{c} \]  

(Eq. 3.7)

Where \( c \) is usually equal to 3/2 for earthquakes, making \( D = 2b \) and creating a scaling relationship between laboratory and field deformation data (Main et al., 1990).

Other examples of this scaling include natural time analysis (Vallianatos et al., 2013) where a magnitude correlation between AEs and global seismicity was derived, suggesting that a universal behaviour of the fracturing processes exists at all scales.

Finally, and most importantly, Burlini et al. (2007; 2009) and Benson et al. (2008) used simple size-frequency scaling to compare the results of their laboratory experiments with the volcanic earthquakes, based on the pioneering work of Mogi (1962). Burlini et al. (2007) qualitatively compared 3 types of events at two scales, showing a remarkable resemblance between the two environments (Fig. 3.8). Taking the frequency of the event (\( f \)) and the size of the structures involved (\( d \)), they proposed that

\[ d_n \times f_n = d_l \times f_l \]  

(Eq. 3.8)

Where subscripts \( n \) and \( l \) denote natural and laboratory respectively. In nature, kilometric structures generate event with frequency around 1-2 Hz, while in a rock deformation apparatus AEs resonating at frequency of few MHz are the results of structures of tens or hundreds of µm. Using these values with Eq. 3.8, the authors obtained

\[ \frac{d_n}{d_l} = 5 \times 10^6 \]  

(Eq. 3.9)

and

\[ \frac{f_l}{f_n} = 2.5 - 5 \times 10^6 \]  

(Eq. 3.10)

which show excellent agreement. The same approach in recent laboratory work (e.g. Benson et al., 2008; Burlini et al. 2009) achieved the same agreement, even extending to a size-viscosity scaling (Benson et al., 2008).
Figure 3.8: (above) normalized spectrogram and (below) seismogram of (A, C, E) three different types of AE compared with (B, D, F) three volcanic earthquakes recorded at Mt. Etna. From Burlini et al. (2007).
4. EQUIPMENT AND METHODS

4.1. TEST MATERIAL AND SAMPLES

The rock used for this study is a porphyritic alkali basalt (Fig. 4.1) from Mt. Etna volcano, Sicily (Italy), previously described (EB). Key reasons for this choice are the isotropic distribution of cracks within the rock mass (Vinciguerra et al., 2005; Stanchits et al., 2006), and numerous studies previously done on this rock type, allowing this rock to be used with a degree of confidence. Furthermore, a lack of anisotropy considerably simplifies data analysis: results are not affected by any preferential orientation of cracks or structures for an isotropic material. The physical properties of EB, as described by previous studies, are listed in Table 4.1.

Figure 4.1: optical microscope pictures taken from a pre-test sample in (a) parallel polarised light and in (b) cross polarised light. The rock is formed by millimetric phenocrysts of clinopyroxene (cpx) and submillimetric phenocrysts of olivine (olv), plagioclase (plg) and oxides (black minerals both in PPL and XPL).
Table 4.1: Physical properties of EB

<table>
<thead>
<tr>
<th>PROPERTY</th>
<th>VALUE</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density (kg/m³)</td>
<td>2860 ± 10</td>
<td>Vinciguerra et al., 2005</td>
</tr>
<tr>
<td>Porosity (%)</td>
<td>2.1</td>
<td>Vinciguerra et al., 2005</td>
</tr>
<tr>
<td>Unconfined compressive strength (MPa)</td>
<td>140 ± 5</td>
<td>Heap et al., 2009</td>
</tr>
<tr>
<td>Young’s modulus (GPa)</td>
<td>32</td>
<td>Heap et al., 2009</td>
</tr>
<tr>
<td>Poisson’s ratio</td>
<td>0.22</td>
<td>Heap et al., 2009</td>
</tr>
<tr>
<td>P-wave velocity (km/s)</td>
<td>3.05 – 3.43</td>
<td>Vinciguerra et al., 2005</td>
</tr>
<tr>
<td>Permeability (m⁻²)</td>
<td>7 – 1 x 10⁻¹⁶</td>
<td>Vinciguerra et al., 2005</td>
</tr>
<tr>
<td></td>
<td>at Pₐ of 5 – 80 MPa</td>
<td></td>
</tr>
</tbody>
</table>

Blocks of EB were collected at the quarry located in Camporotondo Etno (Italy) where a pile of effusive lava flows is well exposed, located on the south-western flank of Mt Etna (Fig. 4.2a, b, c). Each block is 200 mm long, 150 mm wide and 120 mm deep, allowing 14 cylindrical samples to be cored. Samples were cored with a 40 mm-wide core barrel along the smallest dimension of the block and cut to a length of approximately 10 cm with a diamond saw. The end faces of the cylinder were then accurately ground using a lathe fitted with a cross-cutting diamond grinding disk. This yields a sample size of 40 mm diameter x100 long (± 0.5 mm), with surface flat and parallel to ± 0.01 mm. The 2.5 ratio between length and diameter was chosen in agreement with ISRM recommendations (Ulusay & Hudson, 2007), and current best practice in triaxial rock deformation research (e.g. Benson et al., 2007).

Experiments were performed on both conventional (solid) sampled, or samples that were prepared with a pre-drilled conduit throughout its axis. The conduit was cored with diamond core drills of two different diameters, either 3mm (Fig. 4.2d) or 10 mm. When machining the axial conduits, to avoid damaging the drill due to the fast rotation and small aperture, shorter drills (5 and 8 cm long) were needed, meaning that a single drilling is not sufficient to create a through-going conduit. Therefore, samples are accurately clamped by a steel v-block and one surface is first drilled by 5-cm-long drill to act as a pilot-hole, followed by the 8-cm-long drill. At this point the sample is overturned and the same drilling procedure is done on the other surface. In this way, the two halfway through conduits reach other to accurately form a full-length conduit. The purpose of
the conduit is to allow the damage zone generated by the triaxial deformation procedure to be quickly depressurized, and thus create the best chance possible of stimulating fluid-induced seismicity. The importance of this is further discussed in chapter 6.

![Figure 4.2](image)

*Figure 4.2: (a) Map of Italy, showing the location of Mt. Etna, (b) the volcano viewed from the east flank, (c) the quarry where the blocks of EB were taken and (d) an EB sample with a 3-mm-wide conduit throughout its axis. The arrow, which start from the conduit, indicates the north for the sensors array.*

4.2. PRELIMINARY TESTS

The density and porosity of EB was measured using standard ISRM saturation and buoyancy techniques (Ulusay & Hudson, 2007). This consists of submerging the samples in water and placing them inside a vacuum chamber for 24 hours, allowing the water to fill the pore space. Once removed from the chamber, samples were weighed \( M_{\text{sat}} \) and then dried in oven at 105 °C for at least 8 hours, followed by at least half an hour in a desiccator. Dry weight \( M_d \) measurements were taken as well the dimensions (length and diameter: \( l \) and \( d \)) of the samples.

The density \( \rho_s \) is calculated by the equation:

\[
\rho_s = \frac{M_d}{\pi \frac{d^2}{4} l} = \frac{M_d}{V},
\]

(Eq. 4.1)
where $V$ is the volume of the sample. The porosity ($n$) is retrieved by the equation:

$$n = \left( \frac{M_{sat}-M_s}{\rho_w} \right) \cdot \frac{100}{V} \% = \frac{100 V_v}{V} \%,$$

(Eq. 4.2)

where $\rho_w$ is the density of water and $V_v$ is the pore volume.

### 4.3. EXPERIMENTAL PROTOCOL

All the experiments in this study are carried out in two consecutive stages, namely the deformation phase and the subsequent pore pressure release (hereafter called venting) phase. Initially the assembly, including the sample, the rubber jacket and the sensors (see section 4.6), is mounted inside the vessel of the triaxial apparatus (see section 4.4) and pressurised hydrostatically. For the experiments run in wet (water saturated) conditions, pore pressure is applied to previously vacuum saturated samples and brought up to the chosen pressure. Note that the effective pressure of 30 MPa is kept constant in all the experiments to avoid different mechanical behaviour at different effective pressure conditions. At this point a standard triaxial deformation test (Ulusay & Hudson, 2007) is run at constant strain rate, i.e. $1 \times 10^{-5}$, until the sample fails and a natural fracture zone is created.

The axial stress is then reduced to hydrostatic conditions in order to prevent any movement along the fracture due to differential stress. If water is being used as the pore fluid for the venting phase, the vessel and the pore fluid pipes are first heated to 175°C for an hour in order to allow temperature equilibrium between the top and the bottom of the vessel. If the venting is conducted at room temperature, or if Nitrogen gas is used as pore fluid, no further action is needed. The vessel is then depressurized and the sample is carefully removed from the rubber jacket, keeping the failed sample in position using elastic bands, and taking care to preserve any fault gauge created during the development of the shear zone and the movement of the two side of the fault. At this point photos are taken at different orientations in order to match them with the location of the AEs, in particular in the final seconds before the failure. Samples are then encapsulated in clear heat shrink tubing.
4.4. THE TRIAXIAL APPARATUS

Triaxial experiments were carried out in an externally heated, servo-controlled triaxial apparatus custom built by Sanchez Technologies (Fig. 4.3 & 4.4). The apparatus is designed to test specimens 100 mm long and 40 mm diameter up to confining pressures of 100 MPa and a temperature 200 °C using an external furnace, composed of heating pads bonded onto the vessel outer walls. An external jacket (insulator) is wrapped around the vessel to avoid loss of heat. A specialist heat-transfer fluid, composed by high-flash point (270°C) oil, is used as confining medium, as well as to provide an axial stress via 100 MPa precision piston pumps. For confining pressure the pump pressure is used directly, whereas, for axial stress a hydraulic booster is fitted (Fig. 4.4). The top of the booster uses a piston of 70 mm diameter on which 100 MPa pressure is applied. This is connected to a smaller 40 mm, amplifying the applied stress by a facture of 6, and allowing the apparatus to achieve a maximum 680 MPa of principal stress on a 40-mm-diameter sample.

Different pore fluids may be used with the system. However, for the purposes of this study, distilled water was used, pressurised via a single 100 MPa precision piston pump. For experiments using gas, a bottle of dry Nitrogen gas was fitted (Fig 4.4) via a simple regulator. The pore pressure pumps are equipped with similar heating pads as described above, including high pressure pipework, to allow high temperature fluids to be injected into the sample and thus avoid cooling the main vessel. For rapid pore pressure venting, an electro-mechanical fast-acting solenoid valve is employed and connected to the lower part of the sample, controlled and signalled via LabView software installed on the AE recorders to ensure accurate synchronization.
Figure 4.3: (a) the Sanchez Technologies triaxial apparatus; (b) a zoom in on the pressure vessel.

To measure axial displacement and strain, the apparatus is equipped with an Eddy Displacement System (EDS) consisting of three contactless transducers mounted on a ring fixed to the vessel. These devices record the change in distance to a steel target placed approximately 2 cm above them and connected mechanically to the top piston driving the sample strain. The transducers create an

Figure 4.4: Schematic overview of the Sanchez triaxial apparatus.
electric field, using target plates to record a magnetic response, which is then a measure of the distance with movement parallel with the axial displacement with sub-micron accuracy. The three readings are averaged and used to calculate the strain according to the sample length.

Sample strain values are first corrected for the Young’s modulus of the apparatus \((E_{\text{app}})\), to yield a corrected strain. To calculate \(E_{\text{app}}\) and thus obtain the corrections for the strain, an aluminium-alloy cylinder (100.05 mm x 40.01 mm) of known Young’s modulus \((E_t = 73 \text{ GPa})\) is deformed and its strain is measured \((\varepsilon_m)\) (Fig. 4.5). The measured Young’s modulus \((E_m)\), calculated on the linear elastic segment of the calibration curve is:

\[
E_m = \frac{\Delta \sigma_m}{\Delta \varepsilon_m} = 49 \text{ GPa}
\]  
(Eq. 4.3)

where \(\Delta \sigma_m\) is the incremental differential stress and \(\Delta \varepsilon_m\) is the incremental strain, both measured along the linear elastic part. \(E_{\text{app}}\) is then calculated over the same incremental stress \((\Delta \sigma)\) by:

\[
E_{\text{app}} = \frac{\Delta \sigma}{\Delta \varepsilon_{\text{m}} - \Delta \varepsilon_t} = \frac{\Delta \sigma}{\Delta \varepsilon_{\text{app}}} = 130 \text{ GPa}
\]  
(Eq. 4.4)

where \(\Delta \varepsilon_t\) is the theoretical incremental strain of the aluminium-alloy cylinder and \(\Delta \varepsilon_{\text{app}}\) is the incremental strain of the apparatus.

The stiffness of the apparatus \((k_{\text{app}})\) is obtained by applying equation 4.5 (Jaeger et al., 2007).

\[
\Delta \varepsilon_t = \frac{k_{\text{app}}}{L \cdot k_{\text{app}} + A E_m} \cdot \Delta z_{\text{app}}.
\]  
(Eq. 4.5)

Where \(L\) is the length of the sample and \(A\) is the cross-sectional area. The incremental deformation accommodated by the apparatus over the linear elastic part \((\Delta z_{\text{app}})\) is calculated through:

\[
\Delta z_{\text{app}} = L \cdot \Delta \varepsilon_{\text{app}} = 0.03 \text{ mm}
\]  
(Eq. 4.6)

Rearranging equation 4.5 for \(k_{\text{app}}:\)

\[
k_{\text{app}} = \frac{A E_m \Delta \varepsilon_t}{\Delta \varepsilon_{\text{app}} - A L \cdot \Delta \varepsilon_{\text{app}}} = 1.2 \times 10^9 \text{Nm}^{-1}
\]  
(Eq. 4.7)
4.5. MECHANICAL DATA

The mechanical data (stress, strain, pore and confining pressures) are recorded on a data logging PC fitted to the triaxial apparatus via proprietary Falcon control software with a data recording rate of 1 sample/second, and stored as a text (.ascii) file. To record data at higher sampling rates (required for the last minutes of the deformation stage and for the full venting stage), external pressure and transducers were connected to a secondary high-speed data logger. These auxiliary devices consist of two Linear Variable Differential Transformers (LVDTs) to measure the axial displacement via cross-head movement, a secondary axial pressure (booster chamber) transducer, and two pore pressure transducers. The activation of the solenoid valve and the data recording was performed by two different LabView applications recording data at 10 sample/second and 5000 sample/second respectively.

The popular MATLAB® package is used to process the mechanical data and then to synchronize mechanical and AE data accurately in time. This is a complex procedure as the data logging PC on the triaxial apparatus is separate to both the Milne and Richter units dedicated to the AE equipment.
Figure 4.6: comparison between the stress-strain curve using the raw values of strain (uncorrected strain, grey symbols) and the stress-strain curve using the values of strain which take into account the stiffness of the apparatus (corrected strain, black symbols).

In addition to the machine stiffness corrections, both axial stress and strain values require corrections due to the sample deformation before the experiment (although the effect is small). Stress correction is achieved via Eq. 4.8 which returns the corrected axial stress at time $i$:

$$\sigma_{1,i,\text{corr}} = \frac{F_i}{A_s} \cdot \frac{\sigma_{1,i} A_d}{A_s}$$  \hspace{1cm} (Eq. 4.8)

where: $F_i$ is the load, $A_s$ is the cross-sectional area of the sample, $\sigma_{1,i}$ is the uncorrected axial stress recorded by the machine at time $i$ and $A_d$ is the default basal area, based on a 40-mm-wide sample.

Strain is corrected according to $E_{\text{app}}$, using equation:

$$\varepsilon_i = d_i \cdot \left( \sigma_{\text{diff},i} \cdot E_{\text{app}} \right)$$  \hspace{1cm} (Eq. 4.9)

Where: $d_i$ is the relative deformation (average from the three EDDY sensors) at time $i$, $l$ is the length of the sample and $\sigma_{\text{diff},i}$ is the differential stress at time $i$ (Fig. 4.6).

Once the values of stress and strain are corrected, the data are then used for further processing, such as deriving Young’s Modulus by interpolating the values which lie in the elastic part of the
stress-strain curve with a regression line created with MATLAB. For experiments using elevated pore pressure it is also possible to evaluate the onset of the dilatancy. To do so, the volume change in the pore pressure pump is plotted against time, showing the onset of the dilatancy when the pump starts to inject water due to expansion of the sample (due to crack opening), and thus requiring more pore volume to maintain the constant (set) pore pressure.

4.6. THE AE SENSORS

To record the AE signals, the vessel is fitted with 20 high pressure/temperature coaxial leadthroughs, components allowing electric signals to travel in and out the vessel. These feature independent connections for both the central (positive) signal recorded from the AE sensors, as well as the ground shields of the coaxial cable and sensor body. Both are isolated from the grounded vessel body, providing optimum signal. These feedthroughs allow the AE sensors to be directly mounted on the sample via waveguides, described below.

The AE sensors used to record AE (passive) and elastic wave velocities (active surveys) were purpose designed in the Rock Mechanics Laboratory. They comprise a Roditi PZT-5A disk (1 mm thick and 3.5 mm wide) to yield a resonant frequency of 1 MHz mounted inside an aluminium-alloy holder (Fig. 4.7b). These were designed to be sensitive to the high frequency (HF) data generated during rock fracture events. To confirm the frequency response, a frequency generator TTi TG1010A, was used to determine that they were most sensitive to the 50–600 kHz band (Fig. 4.8a). Accompanying these, a set of Low Frequency (LF) sensors were also made, to allow the anticipated LF data to be recorded more accurately. These are identical in design but with a longer body, and with a 10 mm long rod of PZT (Fig. 4.7a) to achieve the desired frequency response of 50–200 kHz to capture good signals at the lower frequency end of the range (Fig. 4.8b). In both cases, the PZT is bonded to an aluminium-alloy insert (or waveguide) which is embedded in a Nitrile (high temperature rubber) jacket (Fig. 4.7c), into which the sample is placed. The rubber jacket is designed after an earlier setup pioneered at the Rock & Ice Physics Laboratory at University College London (Sammonds, 1999). The raypaths, created by each couple of sensors, have a wide range of plunges. These varies from 0 to 62°, which is the most inclined raypath through the rock sample in this study.
Before each experiment, 12 sensors (of different frequency response) were checked to verify their response and mounted properly in the rubber jacket.

Figure 4.7: the PZT sensors with flat frequency response up to (a) 200 kHz and (b) up to 600 kHz. The PZT disks are insulated from the sensor body by 2 layers of high-temperature heat shrink tube in (a) and a rubber disk in (b). The rubber jacket, which holds the samples, where 12 PZT sensors are plugged in.

Figure 4.8: frequency response of (a) HF and (b) LF sensors, shown in a semi logarithmic scale.

4.7. THE AE RECORDING SYSTEM

The AE sensors are connected to a sophisticated AE recording system (Fig. 4.9, Itasca-Image) via 12 Pre-amplifiers (Pulser Amplifier Desktop units: PADs), which amplify the incoming signal
in a selectable range from 30 to 70 dB in 10 dB increments. They also act as a band-pass filter, to reduce local (laboratory/building) electrical noise. LF sensors are used with PADs equipped with a 10 kHz - 1 MHz bandpass filter, while the HF sensors are used with PADs with a 45 kHz – 1 MHz bandpass filter. The amplified signal is then split between two independent, separate, outputs for recording on dual AE systems referred to as triggered data and continuous datastream.

The triggered dataset is captured when a particular set of parameters is satisfied, including channel threshold amplitude (in mV, i.e. sensor output, once pre-amplified), and the minimum number of channels to satisfy this criterion. When the parameters are satisfied, the full waveform across each of the channels is recorded by the Milne unit, a 16-bit data acquisition system. The 12 seismograms forming an event are then stored as a BSF file (Binary Storage File: a proprietary data format). Each event consists of a trigger time ($t_0$), with each channel consisting of a pre-set number of points (before $t_0$) of 25% of the total waveform length. This ensures the best opportunity to pick the first P-arrival in subsequent analysis. Generally, in any given experiment there are thousands of events, but due to the interference between wires (cross-talk) or a bad choice of parameters, only 50% are generally used. Finally, the triggered dataset captured by the Milne unit has a maximum number of events per second that may be captured, beyond which the unit saturates. At this point the number of events per second (about 30) exceeds the event capture rate capability of the unit, generally occurring when the sample approaches failure. This is important, as, without other measures, a loss of information would occur during this final stage, especially during the deformation of samples in the brittle regime.

To avoid this issue, a continuous recording of the voltage output across 8 of the 12 channels is performed by two continuously operating digitizers (the Richter units, 16-bit continuous acquisition and streaming system) recording 4 channels each. In this way, the entire signal is recorded during a chosen time period with data sampled at the same 10 MHz sampling rate as for the Milne unit. Importantly, this overcomes the saturation effect, by allowing post-test processing of the datastream using any threshold or trigger criterion needed to extract the optimum event dataset, even during sample failure. In practice, the Richter units are used to record the 4-5 minutes centred on the failure of the sample where the Milne cannot keep up. These data are multiplexed in a single 16 bit file known as a .srm file (stream: a proprietary format), and split in 1-minute files. The Richter units are also activated during the entire venting stage, so that the AE generated by pore
fluid release through the solenoid valve is continuously recorded: this is a key stage as the fluid
induced seismicity has lower amplitude than AE generated during fracture.

Concurrently with the recording of AE data, the Milne and PAD devices are also used in an
active mode to generate a P-wave elastic velocity surveys using the known locations of the sensors,
and pulsing each sensor in turn using a Pulser Interface Unit (PIU). This instrument sends a high
voltage pulse (200 V) to each sensor in turn (which is first isolated from the recorder to avoid
damage), recording the incoming pulse on the remaining 11 channels. This allows the measurement
of P-wave elastic anisotropy and thus a velocity model to be constructed for the sample via the
time-of-flight method (Pettitt, 1998), needed for AE location of the fracturing events. This is
analogous to an active seismic survey at field scale. Each survey is stored in a dedicated folder and
each shot as a .bsf file. These data are highly sensitive to crack damage within the rock sample
(Vinciguerra et al., 2005) and thus is a standard test in rock physics.

Figure 4.9: overview of the AE recording system.

4.8. THE AE DATA

The AE data is acquired and processed using a commercially package known as InSite-Lab™.
A typical workflow would involve picking the first arrival across all channels of data, cross-
correlate these waveforms to evaluate the relative change in velocity to generate a velocity model
for the sample, and finally to locate the events in 3D. A similar procedure is used for the
continuous data, after first extracting discrete events according to a trigger threshold and amplitude.
The advantage of the continuous dataset lies in its ability to have this data collection run many times to extract the maximum data from the continuous datastream.

4.8.1. VELOCITY SURVEY DATA AND PROCESSING OF ACTIVE EVENTS

To study the behaviour of the P-wave velocities during the two-stage experiments, and to build the velocity model that is fundamental for locating the AE hypocentres, velocity surveys are taken every 1-2 minutes. Each survey generates some 144 BSF files (12 transmitters x 12 receivers), which are converted to .ESF files (a proprietary digital data format). These data are grouped in 12 folders each representing one shot or survey, with each survey then containing 12 ESF files, representing the 12 signals recorded by the receivers (the signal of the transmitter is also included, but is not used). Due to the azimuthal dependency (Pettitt, 1998), not all signals are picked (e.g. if both pulser and receiver lie on the same side of the sample, meaning that the pulse arrives at the receiver at 90° to its axis of movement and therefore induces high shear component). Due to these considerations, 96 of the 144 recorded signals are picked and used. In addition, each raypath is represented by two signals (e.g. the signal shot from channel 3 and received by channel 4 and that one from channel 4 to channel 3 cross the same raypath), leaving 48 unique raypaths. However, some raypaths share same trend and azimuth with others, though through different path within the sample. Therefore the stereogram is composed by 27 unique values of P-wave velocity at certain trend and plunge that are then used to derive a model of velocity, and velocity anisotropy, during the experiment.

![Figure 4.10](image)

*Figure 4.10: Schematic figure illustrating a generic PZT transducer and its active element, the ray incident with an angle of $\alpha_i$ to the transducer normal vector. Re-drawn from Pettitt, 1998.*
The processing of the survey data starts with the manual picking of the P-wave arrival time on the 96 signals of the first survey. While an auto-picking function speed up the processing, it has been observed during this study that a correct picking was never achieved and always required a visual check of the arrivals. Therefore, as the number of waveforms to pick was low (96), a manual picking has been preferred. In this way noisy waveforms, which were picked by the auto-picking function, were immediately discarded.

Then, using a cross-correlation function within InSite-Lab, it is possible to pick the other surveys through the similarity of the signals of any survey with a defined master survey (when transmitter and receiver are the same), yielding an extremely accurate elastic velocity change, due to sample deformation. Finally, the P-wave velocity along the raypath \( j \) (\( V_j \)) and recorded by the receiver \( i \) is calculated using MATLAB through the following equation (Pettitt, 1998):

\[
V_j = \frac{l_j + d_i \sin \alpha_{ij}}{t_i - 2t_{al}},
\]

(Eq. 4.10)

where: \( l_j \) is the length of the raypath \( j \), \( d_i \) is the diameter of the pzt disk of the sensor \( i \), \( \alpha_{ij} \) is the angle between the normal to the sensor \( i \) and the raypath \( j \), \( t_i \) is the P-wave arrival time picked at sensor \( i \) and \( t_{al} \) is the time of flight of the signal in aluminium insert placed between the pzt disk and the sample (Fig. 4.10).

The 27 P-wave velocities are plotted in a stereogram (Appendix 2, Matlab code 1) as lineation (strike, dip) and interpolated with V4-griddata method implemented in MATLAB, which is not affected by triangulation and nor “deterioration of the interpolation surface near the boundary” (MATLAB documentation). The stereogram show the variation of velocities as the sample deforms and are used to build the time-dependant transversely-isotropic velocity model which requires the maximum P-wave velocity, the trend and plunge of the maximum velocity axis, as well as P-wave anisotropy (ratio between the minimum and the maximum velocity) for each survey. To verify the goodness of the velocity model, each shot, whose location is known, was located back using the model. Once at least 6 of the calculated locations of the shots coincide with their positions, within an error of 4 mm (10% of the minimum dimension of the sample), the velocity model can be used to locate both the triggered events and the events collected from the continuous data.
4.8.2. TRIGGERED DATA

In addition to the BSF and ESF AE data files, a simple text file known as a HCD file is outputted, containing statistical data for the experiment (Hit Count Data). This contains the number of hits per time interval for every single channel throughout the experiment. Hits are channel-averaged and plotted against time, showing the supra-exponential increase of hits before failure. The ESF data for the triggered AEs are obtained in the same way as for the survey data. In this case, each ESF file represents a passive event occurring either during the deformation stage or the venting stage- and recorded by the PZT sensors. However, the events during the deformation stage are processed separately from those during the venting stage.

4.8.3. CONTINUOUS DATA

Using the continuous stream (dataset) comprising of an .SRM (stream) proprietary digital data, individual triggered event are extracted through a routine that extracts new ESF and BSF datasets matching a criterion, essentially the same as the Milne unit but capably of being re-run as many times as needed. To achieve this, SRM files are first formatted to allow a basic visualization of the continuous data for each channel, making possible the detection of the noise level of each channel (Fig. 4.11). The minimum value of noise is taken as a common value, and the number of channels to be triggered to store an event is input (equal to number of channels with highest noise, plus one channel). The chosen waveform length is 819.2 µs (4 times longer than standard triggered events recorded via the Milne unit) because the coda of the events around the failure time of the sample, and during the subsequent venting may be not fully recovered using the standard 204.8 µs-long waveform.

As the recording occurs in continuous mode, an HCD file containing the hits above a certain threshold is not produced. Instead, a dedicated MATLAB code (Appendix 2, Matlab code 2) was written to extract a hit count file from the raw SRM data. This script takes each waveform (channel) using the same threshold used to extract the events from the continuous stream and then
count the number of hits (threshold crossings) per tenths of second.

Figure 4.11: screenshot of Insite-Lab, showing a continuous 230-ms-long waveform (grey line) with the black dotted line marking the upper limit of the noise level.

4.8.4. PROCESSING OF PASSIVE EVENTS

Once the events are collected from the continuous stream, the processing proceeds in the same way as per the triggered data. In both HCD and MATLAB-generated hit files, the number of hits are averaged per channel and the cumulative hits during the venting is calculated. Noisy channels, with no hits or with strong interference signals, are discarded.

An auto-picking function, based on the RMS amplitude, is first run to approximately pick the P-wave arrival time and to isolate events with noisy data. Events with no picks are immediately discarded leaving a pre-processed dataset populated with events of reasonable quality data that can then be evaluated manually. While during the deformation stage thousands of events occur, during the venting stage only a few tens of AEs are recorded. Therefore while the AEs during the deformation stage are mainly auto-picked, those during the venting stage are always manually picked to ensure high quality.

At this stage, the $b$-value ($b$) is calculated by using the maximum-likelihood estimation (Aki, 1965; Utsu, 1965):

$$b = \log_{10} \frac{e}{M_{av}-M_{min}} = \frac{0.43}{M_{av}-M_{min}}$$ (Eq. 4.11)

Where $M_{av}$ is the average magnitude and $M_{min}$ is minimum magnitude, above the completeness threshold. To get the magnitude, the location is needed, therefore at least 4 first arrivals. For some experiments the number of magnitudes available is not enough to study the $b$-value, as at least 200 magnitudes are required for a good estimation of the $b$-value (Roberts et al., 2015). Therefore in this study the log10 of the maximum amplitudes (log A) of the extracted event are used to calculate...
the \textit{b-value}. The use of log A instead of M in laboratory experiments, where the estimation of accurate magnitudes is difficult, is justified by the assumption M \propto \log A (Lockner et al., 1991).

For each experiment, the \textit{b-value} is calculated for a fixed number of events in the windows, varying from 500 to 2000, in steps of 250 events, with overlap of 10% (Appendix 2, Matlab code 3). The procedure to get the \textit{b-value} for each window consists of:

1. Create binned log A in steps of 0.05, starting from the minimum log A to the maximum;
2. Estimate the completeness threshold. This is achieved by calculating the gradient for each pair of consecutive binned log A. When the gradient overcomes an arbitrary value of 0.01, the second binned log A of the pair corresponds to the completeness threshold;
3. Calculate the \textit{b-value} using the maximum-likelihood estimation;
4. Create synthetic curves with different a-value. For each a-value, the absolute difference between the observed values of binned log A (B) and each synthetic curve (S) as a function of the a-value, \textit{b-value} and log A used (R, Wiemer and Wyss, 2000), is calculated as follows:

\[ R = 100 - (100\left(\frac{\sum_{M_{i}^{\text{max}}}^{M_{i}}|B_{i} - S_{i}|}{\sum_{M_{i}}^{M_{i}}B_{i}}\right) \]  

(Eq. 4.12)

5. If the maximum R (meaning that the power law model R\% of the observed data), is higher than 95, then the \textit{b-value} and a-value used to get that R are kept. Otherwise iteratively, the highest binned log A is removed and the procedure start again from step 3. The removal of the highest log A is justified as the window length is not long enough to cover the recurrence time of the larger log A event. In this way a linear fit does not apply in a magnitude – log frequency distribution (Kulhanek, 2005).

6. The number of removed log A for the whole continuous data is then superimposed on the stress-time curve for each constant number of event window. A trade-off evaluation is used to preserve a high number of bins to evaluate the temporal variation of the \textit{b-value} and at the same time use as many log A as possible. Following this the \textit{b-value} is analysed by using windows of 1000 events.

For event location in 3D, the simplex algorithm is used (Nelder & Mead, 1965; Press et al., 1994), implemented in InSite-Lab. This takes the velocity model previously calculated with the survey data, and locates events using a minimum criterion of 5 arrival times (picks). The AE
locations showing a RMS error higher than 2 µs and those located outside the rock sample, are disabled while all other events are divided in overlapping time-windows to observe the development of the shear zone in the minutes/seconds before the failure.

A relative magnitude based on the location of the events, called Location Magnitude ($M_L$) is produced by the simplex algorithm in Insite-Lab, following the equation:

$$M_L = \log \left( \frac{\sum_{m=1}^{N} (W_{RMS_m} d_m)}{N} \right)$$ (Eq. 4.13)

Where $W_{RMS_m}$ is the RMS waveform amplitude for each sensor $m$, $d_m$ is the ray path length between the source location and the sensor $m$ and $N$ is the number of sensors. Events with at least 6 picks, high signal to noise ratio (SNR), and clear arrivals are used to retrieve the source components (Pettitt, 1998) of the source (run by InSite-Lab).

The final technique used for AE data analysis during the venting stage is the calculation and evaluation of changes in dominant frequencies, and is the key method employed in this research to investigate the role of fluids in generating AE activity. With a MATLAB code, the dominant frequency is retrieved by computing the discrete Fourier transform using the Fast Fourier Transform (FFT) algorithm in MATLAB on the channel having the waveform with the highest amplitude. This waveform is chosen because it is closer to the source, and thus less susceptible to attenuation and path effects that may interfere with the frequency content of the signal. The FFT is performed on all events with at least one arrival.

The spectrogram is calculated for the whole recorded waveform, consisting of pre-signal noise and post-onset (first arrival) signal, using FFT methods. First arrivals and the arrival amplitude of events selected for source component analysis and spectrogram were all manually picked. In particular, for the analysis of the frequencies, the dimensionless Quality factor ($Q$) of the resonator via the relation (Saccorotti et al., 2007):

$$Q = \frac{f}{\Delta f}$$ (Eq. 4.14)

Where $f$ is the frequency of a given spectral peak and $\Delta f$ is the full-width-half-maximum (FWHM) of the spectral peak. $Q$ is calculated with a MATLAB (Appendix 2, Matlab code 4) routine by plotting the normalised spectra of the signal in dB units and picking the two points lying at -6dB around the dominant frequency.
Classes of events are recognized through visual inspection both in time and frequency domain (e.g. Alparone et al., 2010), while families are recognized through the cross-correlation method and the bridging technique (Barani et al., 2007; Alparone et al., 2010) (Appendix 2, Matlab code 5). If a waveform (A) has a cross-correlation above a chosen threshold (0.8) with two different waveforms (B and C), but B and C are not cross-correlated between each other, then A is the bridge waveform and A, B and C are defined as belonging to the same family. All events of the families formed by at least two events are summed and divided by the total number of the events to get the percentage of cross-correlated events. The cross-correlation matrix between each pair of events is then calculating by using different window-length, from 15 to 45 μs, the former being twice the period of the lowest spectral peak and the latter being half of the minimum duration of the short-duration AE events.

4.9. FIELD DATA

Field data consisting of volcanic earthquakes were generously provided by Salvatore Alparone from Istituto Nazionale di Geofisica e Vulcanologia (INGV). Waveforms of tremor, LP and VLP signals were recorded by INGV permanent seismic array operating at Mt. Etna (Fig. 4.12a), composed of broad-band 3-component digital seismometers (Alparone et al, 2013). The waveform containing volcanic tremor was recorded at the station at the North East Crater on December 2, 2015. The LP event was acquired at the station EBCN, near the Bocca Nuova crater on January 10, 2015, while the VLP event was recorded at the station EPLC, lying north-west to the summit craters, on September 5, 2013. The Tornillo signal was acquired by the IVCR station (located north to the La Fossa crater, Fig. 4.12b) of the permanent seismic array on Vulcano Island, on November 7, 2008. All waveforms are recorded by the vertical component of each station and are processed both in time and frequency domain, similar to the processing of laboratory data. Comparison between laboratory and field data are made through direct visual inspection.
Figure 4.12: a) location map of the EBCN (black circle) and EPCL (red circle) seismic stations on the summit of Mt. Etna. Modified from http://www.ct.ingv.it/it/component/content/article.html?id=129. b) location map of the seismic station on Vulcano Island. IVCR is the northernmost station. From Alparone et al. 2010.
5. RESULTS

5.1. DENSITY AND POROSITY

A total of 29 cylindrical samples of EB are tested to determine the density and porosity. An average density of 2860 kg/m$^3$ was determined with a standard deviation of 10 kg/m$^3$.

The average porosity is 2.05%, with a standard deviation of 0.23% showing that the distribution of the porosity measurements has a greater variance than the distribution of the density measurements.

5.2. OVERVIEW OF THE EXPERIMENTS

A series of 22 experiments at different Pressure/Temperature conditions (Table 5.1) were performed to investigate the AE activity during the initial deformation stage (via triaxial deformation apparatus described in Chapter 4) leading to the failure of the sample, followed by the subsequent venting stage with the release of pore fluid through the failure zone, accessed via the axial conduit. All samples underwent the initial failure stage, whilst 14 were selected for the venting stage based on the quality of the AE catalogue and shear failure.

Sample deformation was carried out at ambient room temperature (generally around 25°C) at an effective pressure of 30 MPa, but using different combinations of confining ($p_c$) and pore pressure ($p_p$):

- 7 experiments were performed dry (no pore fluid) at $p_c$ of 30 MPa (EB21, EB23, EB26, EB31, EB32, EB33, EB35);
- 7 experiments were carried out under water saturated conditions with $p_c$ of 35 MPa and $p_p$ of 5 MPa (EB14, EB15, EB16, EB17, EB18, EB19, EB36);
- 8 experiments were run under saturated conditions with $p_c$ of 46 MPa and $p_p$ of 16 MPa (EB20, EB24, EB25, EB27, EB28, EB29, EB30, EB34).

The venting stage consists of the release of water pore pressure at room temperature (EB16, EB29, EB34) or superheated water at 175°C (EB15, EB18, EB19, EB20, EB25, EB27, EB28, EB36). In addition, after the failure under dry conditions, nitrogen gas was used to pressurise the pore space in 4 samples (EB31, EB32, EB33, EB35) and subsequently vented out at room temperature.
Table 5.1: overview of the 22 experiments

<table>
<thead>
<tr>
<th>DEFORMATION STAGE</th>
<th>VENTING STAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry Saturated (5 MPa)</td>
<td>Saturated (16 MPa)</td>
</tr>
<tr>
<td>EB21</td>
<td>EB14</td>
</tr>
<tr>
<td>EB23</td>
<td>EB15</td>
</tr>
<tr>
<td>EB26</td>
<td>EB16</td>
</tr>
<tr>
<td>EB31</td>
<td>EB17</td>
</tr>
<tr>
<td>EB32</td>
<td>EB18</td>
</tr>
<tr>
<td>EB33</td>
<td>EB19</td>
</tr>
<tr>
<td>EB35</td>
<td>EB36</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
</tbody>
</table>

5.3. DEFORMATION UNDER DRY CONDITIONS

Table 5.2 summarises the physical conditions and the mechanical results (strength and strain at peak) for experiments carried out under dry conditions; the full stress-strain evolutions are presented in Figure 5.1 for all experiments. The samples failed at an average maximum differential stress of 457 ± 33 MPa and a corrected strain of 0.0111 ± 0.001. The average tangent Young’s modulus, calculated over the linear elastic part of the stress - strain curve, is 55 ± 5 GPa.
Table 5.2: physical properties and results of the 7 experiments run at dry conditions ($p_c = 30$ MPa, $p_p = 0$).

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Length (mm)</th>
<th>Diameter (mm)</th>
<th>Conduit d (mm)</th>
<th>Duration (mm:ss)</th>
<th>Max. $\sigma_{diff}$ (MPa)</th>
<th>$\varepsilon_{peak}$</th>
<th>E (GPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EB21</td>
<td>100.13</td>
<td>39.78</td>
<td>0</td>
<td>25:52</td>
<td>486</td>
<td>0.0112</td>
<td>58</td>
</tr>
<tr>
<td>EB23</td>
<td>99.98</td>
<td>39.79</td>
<td>3</td>
<td>25:38</td>
<td>490</td>
<td>0.0111</td>
<td>60</td>
</tr>
<tr>
<td>EB26</td>
<td>100.14</td>
<td>39.97</td>
<td>0</td>
<td>25:07</td>
<td>459</td>
<td>0.0109</td>
<td>57</td>
</tr>
<tr>
<td>EB31</td>
<td>99.99</td>
<td>39.90</td>
<td>3</td>
<td>28:00</td>
<td>488</td>
<td>0.0127</td>
<td>50</td>
</tr>
<tr>
<td>EB32</td>
<td>99.87</td>
<td>39.88</td>
<td>3</td>
<td>25:44</td>
<td>449</td>
<td>0.0108</td>
<td>52</td>
</tr>
<tr>
<td>EB33</td>
<td>99.88</td>
<td>40.08</td>
<td>3</td>
<td>25:18</td>
<td>403</td>
<td>0.0114</td>
<td>46</td>
</tr>
<tr>
<td>EB35</td>
<td>99.88</td>
<td>39.89</td>
<td>3</td>
<td>21:54</td>
<td>428</td>
<td>0.0093</td>
<td>60</td>
</tr>
<tr>
<td>Av.</td>
<td></td>
<td></td>
<td></td>
<td>25:21</td>
<td>457</td>
<td>0.0111</td>
<td>55</td>
</tr>
<tr>
<td>St. dev.</td>
<td></td>
<td></td>
<td></td>
<td>01:48</td>
<td>33</td>
<td>0.001</td>
<td>5</td>
</tr>
</tbody>
</table>

Figure 5.1: stress – strain plot of all 7 experiments run at dry conditions ($p_c = 30$ MPa, $p_p = 0$).
5.3.1. **ACTIVE EVENTS (ELASTIC-WAVE VELOCITY)**

The change in P-wave elastic velocities throughout the experiments are shown in Figure 5.2 through 5.7 (6 experiments). Due to technical issues during experiment EB31, the P-wave velocities for this experiment are not available. For the remaining experiments, the horizontal and the most inclined (62°) P-wave velocity are superimposed on the stress-time curve and presented in panel (a) in each figure. The use of P-wave velocity recorded at 62° to the minimum stress plane during triaxial deformation (σ2=σ3) is due to the absence of sensors recording the vertical P-wave velocity (in the σ1 or loading direction) due to technical limitations, which is the maximum elastic velocity during triaxial deformation experiments (e.g. Benson et al., 2007). The arrows in panel (a) mark the 6 velocity surveys shown as stereonets in panels (b) to (g), which represent a projection of velocities plotted as poles in the lower hemisphere. The first stereonet (b) shows the bulk velocity before the start of the deformation phase, with stereonet (c) through (d) illustrating the velocity difference with respect to the previous survey at strain percentages of approximately 0.25, 0.50 and 0.75. Stereonet (f) shows the last survey taken before sample failure, and stereonet (g) the first survey immediately after failure. Yellow and red (hot) colours indicate an increase in velocity from the previous stereonet, whereas cyan and blue (cooler) colours indicate a decrease in velocity.

Table 5.3 summarises the initial P-wave velocities of the 62° and horizontal (0°) raypaths as recorded before the deformation in the intact sample (initial conditions), as well as their maximum through the experiment. At pre-test conditions, the average P-wave velocities are 4200 and 4000 m/s respectively, with the inclined velocity ranging from 3700 to 5000 m/s and the horizontal velocity ranging from 3700 to 4700 m/s. The average maximum P-wave velocities are 5600 m/s at 62° and 4400 m/s horizontally. At their maximum, the velocity at 62° ranges from 5300 to 5800 m/s, while the horizontal velocity from 4100 to 5000 m/s. Note that P-wave velocity for EB33 is somewhat higher than the other samples at the same conditions, and its maximum velocity also the highest of the suite of 6 experiments. This may be the result of a local denser area, such as a zone with fewer pre-existing cracks. This interpretation is supported by the fact that the maximum velocities, which are observed when the cracks are being closed, are similar to the other samples.

All experiments show a more pronounced velocity increase in the axial direction during sample deformation, and a general increase of all velocities (including radial) up to a sample strain of 0.5%. Horizontal velocity (0°) increases by 300 – 500 m/s during this time while the velocities at 62°
increase by 800 – 1700 m/s. Subsequently, horizontal velocities start to decrease, while the sub-vertical velocities decrease from a strain of approximately 0.75%. At sample failure, all velocities undergo a sharp decrease, but while the horizontal velocity reaches values below their initial value, the vertical velocity remains higher than in the intact sample.

Table 5.3: P-wave velocities statistics of the 6 over 7 experiments run at dry conditions ($p_c = 30$ MPa, $p_r = 0$).

<table>
<thead>
<tr>
<th>Sample #</th>
<th>initial 62°vel (m/s)</th>
<th>initial 0°vel (m/s)</th>
<th>Max. 62°vel (m/s)</th>
<th>Max. 0°vel (m/s)</th>
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Figure 5.2: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (solid blue line) and (from b to g) the stereonet of six surveys (indicated by the arrows in panel a) showing b) the velocities of the intact sample and (from c to g) the difference in velocity with the previous survey for EB21 ($p_c = 30$ MPa, $p_p = 0$). b) The intact sample shows a homogeneous P-wave velocity around 3.5 km/s, (c & d) with an increase in all directions up to 0.52% strain. e) At 0.76% strain only vertical velocities are still increasing, (f) followed by a decrease in all directions when strain is 0.85%. g) After failure all velocities decreases between 1 and 2 km/s from the pre-failure velocities.
Figure 5.3: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (solid blue line) and (from b to g) the stereonet of six surveys (indicated by the arrows in panel a) showing b) the velocities of the intact sample and (from c to g) the difference in velocity with the previous survey for EB23 ($\sigma_\text{p} = 30$ MPa, $p_p = 0$). b) The intact sample shows a homogeneous P-wave velocity around 3.5 km/s, (c & d) with an increase in all directions up to 0.53% strain. e) At 0.73% strain a condition of no change in velocities is observed, (f) followed by a decrease in all directions when strain is 0.88%. g) After failure all velocities decreases between 0.5 and 1.5 km/s from the pre-failure velocities.
Figure 5.4: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (solid blue line) and (from b to g) the stereonet of six surveys (indicated by the arrows in panel a) showing b) the velocities of the intact sample and (from c to g) the difference in velocity with the previous survey for EB26 (p_c = 30 MPa, p_p = 0). b) The intact sample shows a cross-shaped distribution of P-wave velocity around 3.5 km/s, with higher velocities at 4 km/s around the edge of the cross, along horizontal raypaths. c & d) All velocities then start to increase up to a strain of 0.49%, (e) followed by a steady condition at 0.74%. f) A general decrease of velocities occurs just before failure, (g) followed by an heterogeneous post-failure change in velocities, with a decrease of about 1.5 km/s at for vertical directions, and an increase up to 0.5 km/s along E-W horizontal directions.
Figure 5.5: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (solid blue line) and (from b to g) the stereonet of six surveys (indicated by the arrows in panel a) showing b) the velocities of the intact sample and (from c to g) the difference in velocity with the previous survey for EB32 ($p_c = 30$ MPa, $p_p = 0$). b) The intact sample shows a non-homogeneous distribution of P-wave velocities, averaging 4 km/s. c & d) All velocities then start to increase up to a strain of 0.50%, (e) followed by a steady condition at 0.73%. f) A general decrease of velocities, with a more marked decrease along E-W horizontal directions, occurs just before failure, (g) followed by an heterogeneous post-failure change in velocities, with a decrease of about 1.5 km/s at for vertical directions, and an increase up to 0.5 km/s along NE-SW horizontal directions.
Figure 5.6: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (solid blue line) and (from b to g) the stereonet of six surveys (indicated by the arrows in panel a) showing (b) the velocities of the intact sample and (from c to g) the difference in velocity with the previous survey for EB33 ($p_c = 30 \text{ MPa}, p_p = 0$). b) The intact sample shows a non-homogeneous distribution of P-wave velocities, averaging 5 km/s. c & d) All velocities then start to increase up to a strain of 0.47%, (e) followed by a steady condition at 0.77%. f) A quasi-homogeneous decrease of 0.5 km/s occurs before failure, (g) followed by a heterogeneous post-failure decrease in velocities, with a sink of about 1 km/s around a SW raypath, plunged at about 40°.
Figure 5.7: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-
wave velocities superimposed on the stress-time plot (solid blue line) and (from b to g) the stereonet of six surveys (indicated by the arrows in panel a) showing (b) the velocities of the intact sample and (from c to g) the difference in velocity with the previous survey for EB35 ($p_c = 30$ MPa, $p_p = 0$). b) The intact sample shows a non-homogeneous distribution of P-wave velocities, averaging 4 km/s. c & d) All velocities then start to increase up to a strain of 0.51%, (e) followed by a steady condition at 0.72%. f) A quasi-homogeneous decrease of 0.2 km/s occurs before failure, (g) followed by a heterogeneous post-failure decrease in velocities, with a sink of about 1 km/s around a NW raypath, plunged at about 40°.
Figure 5.8 shows the P-wave anisotropy for dry experiments EB21 (a), EB23 (b), EB26 (c), EB32 (d), EB33 (e) and EB35 (f). Here, anisotropy is calculated by dividing the 62° P-wave velocities with the average of the horizontal velocities, per survey. All plots indicate a low level of anisotropy (values from 1 to 1.1) which steadily increase as the deformation (strain) increases. In all experiments, except EB26 and EB32, velocity surveys are paused before the failure so as to not interfere with the recording of the continuous data by the Richter units. A general, concave-down-shaped increase of the anisotropy from 200 s to 800 s is followed by a concave-up-shaped increase and a decrease after failure. In general, a maximum P-wave anisotropy between 1.3 and 1.5 was reached before failure, which then decreases to the range 1.2 – 1.4. EB21 and EB35 instead show an increase of anisotropy level even after the failure, reaching 1.65 and 1.5 respectively, which decreases only after the axial pressure is removed. This inverse behaviour can be explained in terms of lack of a pre-failure survey, close to the failure, such as the case of EB26 (c) and EB32 (d). It is also clear from panels (a) and (c) that the presence of a conduit makes little different to either the mechanical or the elastic wave velocity data during the deformation.
Figure 5.8: P-wave anisotropy superimposed on the stress-time plot for experiment EB21, EB23, EB26, EB32, EB33 and EB35, run at dry conditions \((p_c = 30 \text{ MPa}, p_p = 0)\). Both experiments with no pre-drilled axial conduit (a and c) start from an isotropic condition showing an almost identical increase of anisotropy up to 1300 s, when the last pre-failure survey was taken on EB21 (a).

However post-failure behaviour is different, with anisotropy increasing in EB21 (a) and decreasing in EB26 (c). All other experiments (b, d, e and f), are characterized by the presence of a pre-drilled axial conduit and share a similar starting anisotropy of about 1.1, with a less significant increase in anisotropy compared to (a) and (c). EB32 (d), which has surveys taken every minute (rather than paused well in advance before the failure for the benefit of hypocentre locations), reveals a significant increase of anisotropy prior to the failure. In all cases, the anisotropy curves present an inflection point (around 600 s in a and c, around 800 s in b, d, e and f) marking the transition between a concave-down curve to a concave-up pattern.
5.3.2. PASSIVE EVENTS (ACOUSTIC EMISSION)

The AE hit rate, sampled every 5 seconds, is shown in Figure 5.9. All experiments have no AE activity up to 400-700 s, which corresponds to a percent strain of approximately 0.5% (see arrows (d) in Figs. 5.2 – 5.7). At this point, the hit rate increases exponentially (average $R^2$ of the exponential fit: 0.9063) until the last 45-30 seconds. Beyond that, the hit rate grows supra-exponentially (not fitting anymore the exponential fit in black dotted line), reaching a value between 160-200 hits at the time of failure. Once a through-going shear plane has been generated in the sample, the hit rate decreases and AE activity ceases when the differential (axial) stress is removed. In terms of AE onset, pattern and maximum hit rate value, no difference are observed between solid samples (a, c) and samples with a pre-drilled conduit (b, d, e, g). As shown for the velocity and anisotropy plots (Fig. 5.5 and 5.8), EB33 seems to represent an anomaly, having limited AE activity up to 900 s ($\varepsilon_\% = 0.65$) and a maximum hit rate value of 115 hits.

The temporal variation of the $b$-value around the failure time (40 seconds before and 20 seconds after the major stress drop, $\sigma_{\text{drop}}> 50$ MPa), computed from the distribution of logarithmic amplitude vs. log (number) of AE events, is displayed in Figure 5.10. While no clear pattern is observed up to 5 - 10 s before the failure, as failure is approaching a common gentle decrease occurs, followed by a dramatic decrease of $b$-value approximately 1s before failure. After failure the $b$-value increases to pre-failure levels. Here again EB33 (f) show a different pattern, with an increasing trend preceding the sharp decrease before failure. In EB32 (e), the decrease starts about 15 s earlier, and it is characterized by a multiple minima, observed during multiple stress drops.
$p_c = 30 \text{ MPa}; \ p_p = 0; \ p_{\text{eff}} = 30 \text{ MPa}$

Figure 5.9: AE hit rate (line and dots) superimposed on the stress – time curve (solid line) for experiment EB21, EB23, EB26, EB31, EB32, EB33 and EB35, run at dry conditions ($p_c = 30 \text{ MPa}, \ p_p = 0$). Each curve is characterized by i) initial no AE activity (flat part); ii) exponential increase (dotted line, with equation and $R^2$) of hit rate; iii) supra-exponential growth before failure; iv) hit rate maximum at the time of failure; v) port-failure hit rate decrease.
Figure 5.10: temporal variation of the b-value (line and dots) superimposed on the stress-time curve (solid line) for experiment EB21, EB23, EB26, EB32, EB33 and EB35, run at dry conditions ($p_c = 30$ MPa, $p_p = 0$). Each curve is characterized by i) flat or oscillatory pattern up to 6-7 preceding the failure followed by ii) a gentle decrease of b-value and iii) a sharp decrease of b-value 1 s before failure; iv) b-value minima at the time of failure; v) post-failure b-value increase.
To further understand the spatial distribution of the fracture events, Figure 5.11 plots the location of the extracted events recorded around the failure time for samples in dry conditions. To ensure only robust locations are considered, events are first filtered to hypocentres within the sample and with a RMS error < 2 µs (discussed in section 4.8). In addition, the reliability of the locater algorithm (which in turn depends on the velocity model) can be assessed by comparing the preferential orientation of the hypocentres and the photo of the post-test sample in Figure 5.12 and 5.13. This is further discussed in Chapter 6.

Apart from EB31 (which has a number of located events more than twice the average number (i.e. 2092) of located events) and EB26, all other samples display a clear pattern to the damage zone evolution, leaving some portions of the sample with limited AE activity during the minute-long-window. This suggests a clustering of the hypocentres towards the eventual damage zone at least 40 seconds before its full development.

To understand the temporal evolution of the hypocentres, six 11-s-long overlapping windows are made for each experiments, which are shown in Figure 5.12 and 5.13. In all panels, window 4 represents the 11 seconds preceding the primary stress drop, and the last two windows showing the AE generated by the slipping along the newly created damage zone (and therefore more likely to image the fracture).
Figure 5.11: locations of the events recorded around the time of the sample’s failure for experiment EB21, EB23, EB26, EB32, EB33 and EB35, run at dry conditions ($p_c = 30$ MPa, $p_p = 0$). The sample is represented by the cylinder’s frame while each hypocentre is represented by a circle, whose size is proportional to the location magnitude ($M_L$) of the event (middle-right legend), and the colour corresponds to the location error ($\text{Err}_{\text{loc}}$) calculated by the location algorithm (far right legend).
A total of 1910 events are located for sample EB21 (Fig. 5.11a), whose deformation produced a NE-SW fault cutting the whole sample and a shorter NW-SE fracture zone limited to the top part of the sample (Fig. 5.12a). The majority of the AEs occurred in the 21 seconds centred on the stress drop (4th and 5th window), with a preferential clustering in the middle-east sector, close where the two fractures merge. The same clustering, but with fewer events, is visible during fault slip, (also known as stable sliding, 6th window) and in the 10 s before the stress drop, when the stress peak occurs (3rd window). No clear pattern is visible in the first two windows, suggesting that the final damage zone starts to develop 10-20 seconds before.

In EB23 1760 AEs are located (Fig. 5.11b), imaging a NE-SW fracture (Fig. 5.12b). The failed sample reveals also a NW-SE fracture in the middle-east section. Although EB21 is a solid sample and EB23 has a pre-drilled conduit, the temporal evolution of hypocentres of EB23 is similar to that of EB21 suggesting no influence of the conduit on the mechanical failure. While no clear cluster develops up to 20 seconds before the primary stress drop, (1st and 2nd windows), in the final 20 seconds (3rd and 4th windows) the hypocentres image a middle-east area of intense fracturing. During slip along the fault (5th and 6th windows) AE activity is concentrated along the NE-SW fault.

Similarly to EB21, EB26 is a solid sample having 1379 located events (Fig. 5.11c). Post-test inspection reveals a NE-SW fault (Fig. 5.12c), which is not imaged by the hypocentres of the AEs, both with all hypocentres are plotted (Fig. 511c) and with overlapping windows (Fig. 5.12c). However, 10 seconds after the stress drop (6th window), AEs generated by fault slip are then generated in a NE-SW direction. Around the stress drop two distinct areas of intense fracturing are imaged: the south-east corner before the stress drop (3rd and 4th windows), and the middle section with a slight NW-SE pattern (5th window). The first 2 windows, when the peak stress is reached, are characterized by limited and widespread AE activity with little clustering.

Sample EB31 presents the greatest number of located events (4658, Fig. 5.11d), which mask a clear pattern of hypocentres. The deformation caused a complex system of fractures: a localized NE-SW fracture ends against its conjugate widespread NW-SE fracture zone (Fig. 5.12d). The temporal evolution of hypocentres (Fig. 5.12d) does not image such system up to stress drop (first 4 windows), but instead depicts a central area of intense fracturing after the stress peak, which
expands both upwards and downwards as the deformation proceeds. The NW-SE fracture zone is only imaged during fault slip (5th and 6th windows).

EB32 is the sample with the second highest number of located events, clustered along a NW-SE direction (3240, Fig. 5.11e), which is also present in the post-test sample (Fig. 5.13a). Unlike the previous experiments, the overlapping time-windows (Fig. 5.13a) reveal that the NW-SE fault start to develop earlier, at least 30 seconds before the stress drop (1st window), in relation to peak stress. As the deformation proceeds, while the AE activity increases, the clustering is preserved up to and beyond the stress drop.

Sample EB33 is characterized by the lowest AE activity, with only 547 (a quarter of the average of all dry experiments) located events (Fig. 5.11f). Nevertheless, both all combined and temporally divided hypocentres imaged a NW-SE damage zone. Post-test inspection (Fig. 5.13b) shows the presence of this fault which starts to develop early, during the stress peak (1st window), similarly to experiment EB32. However, the following two windows do not display such a trend, which reappears again in the last 11 seconds before the stress drop (4th window) and beyond, during fault slipping (5th and 6th windows).

Finally the hypocentres of EB35, characterized by 1155 located events, image an area of intense fracturing in upper-east part of the sample (Fig. 5.11g), which is not seen visually once the sample is taken out from the triaxial apparatus (Fig. 5.13c). The upper-east of the sample starts to manifest high AE activity from 10 seconds before the stress drop and after the stress peak (3rd window), which continues and intensifies through the stress drop (4th window) and during the first seconds of fault slip (5th window). The earliest and the last windows shows low and widespread (but diffused) AE activity.
Figure 5.12: temporal evolution of the hypocentres of the experiments (a) EB21, (b) EB23, (c) EB26 and (d) EB31 ($p_c = 30$ MPa, $p_p = 0$). For each experiment, the top panel shows 11-s-long overlapping windows corresponding to the X-axis time ticks with hypocentres (for size and colour, see Fig. 5.11); the lower panel is the stress-time curve; the right panel show a photo taken once the sample has failed and removed from the triaxial apparatus.
As seen previously, hypocentre (5.12 and 5.13) clustering starts approximately 10 – 20 seconds before the major stress drop. However, to evaluate the degree of clustering in laboratory data and how this parameter evolves as the sample fails, an approach similar to calculate the $b$-value is used, based on applying bins of equal number of events, to analyse the average inter-event distance. For each pair of consecutive events, their distance is calculated, and the average is obtained by averaging 200 distances with an overlap of 10%. The time of the bin is taken as the average. The
average distance is then plotted superimposed on the stress-time plot (Fig. 5.14). The same
approach is used to study the temporal evolution of the \( M_L \) of the events (Fig. 5.15).

Apart from EB26 and EB33, 5 of the 7 experiments show a clear clustering of the events as the
sample approaches failure. From an average of 30 mm at \( \approx 30 \)s before failure, the distance between
each pair of consecutive events decreases steadily to a value of 26-17 mm at the time of failure,
with post-failure characterized by high degree of clustering. Early trends can be observed in EB21
(a), EB31 (d) and EB35 (g), where a general negative slope appear from the beginning of the plot
(30 – 40 seconds before failure). EB32 (e) shows a negative slope starting 10 seconds before, while
failure in EB23 (b) is preceded by the occurrence of a dramatic clustering 2 seconds before. This is
as distinct from the post-failure behaviour of EB35, where the inter-event distance recovers to pre-
failure levels. EB26 (c) shows clustering 10 seconds before the stress drop, but at the time of
failure a local maximum in the average distance curve appears and inter-event distance keeps
increasing after the stress drop. EB33 (f), has also a maximum at the time of failure, but after the
stress drop the inter-event distance decreases to 26 mm.

Figure 5.15 represents the variations of the average \( M_L \) as the sample approaches the failure and
beyond it. For this parameter, all experiments show a clear increase of the average \( M_L \) (0.5 to 1.5
magnitude), between 1 and 10 seconds before the stress drop. All experiments have an early flat or
slightly increasing curve, meaning no significant variation in average \( M_L \), followed by a supra-
exponential increase which ends at the time of failure, returning to pre-failure level at the same
supra-exponential pace.

This parameter is very sensitive to multiple stress-drops, as in the case of EB32 (e), where two
consecutive stress drops are imaged by two local maxima in the average \( M_L \) trend.
Figure 5.14: average inter-event distance (line and dots) superimposed on the stress-time plot (solid line) for experiment EB21, EB23, EB26, EB32, EB33 and EB35, run at dry conditions ($p_c = 30$ MPa, $p_r = 0$)
Figure 5.15: average location magnitude (line and dots) superimposed on the stress-time plot (solid line) for experiment EB21, EB23, EB26, EB32, EB33 and EB35, run at dry conditions ($p_c = 30$ MPa, $p_o = 0$).
5.4. DEFORMATION UNDER SATURATED CONDITIONS ($p_p = 5$ MPa)

Table 5.4 summarises the sample physical properties and the mechanical results from the experiments carried out under water saturated conditions using pore pressure of 5 MPa and effective confining pressure of 30 MPa: the same as for dry experiments as reported in section 5.3. This represents crustal conditions at a depth of approximately 1.2 km. The full stress-strain data are shown in Figure 5.16 for all experiments. The samples failed at an average maximum differential stress of $435 \pm 28$ MPa and a corrected strain of $0.0102 \pm 0.0011$. The average Young’s modulus, calculated over the linear elastic part of the stress strain curve, is $60 \pm 7$ GPa.

Table 5.4: physical properties and results of the 7 experiments run at saturated conditions ($p_c = 35$ MPa, $p_p = 5$ MPa).

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<th>Duration (mm:ss)</th>
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5.4.1. ACTIVE EVENTS (ELASTIC-WAVE VELOCITY)

Figure 5.17 to 5.23 show the change in the P-wave velocities throughout the water saturated (p_p = 5 MPa), experiments, tabulated in table 5.5 which lists pre-test condition (intact) and maximum value of both 62° and horizontal (0°) P-wave velocities, in the same manner as earlier. For intact samples, an average P-wave velocity of 5300 m/s at 62° is measured, ranging from 5300 to 5600 m/s; horizontally the average P-wave velocity is 5100 m/s, ranging from 4800 to 5300 m/s. Similarly to the dry case, 62° velocities show a higher increase during the deformation than the horizontally, reaching an average of 5900 m/s (5800 – 6200 m/s) along the most inclined direction, and 5300 m/s (5000-5500 m/s) along the horizontal direction. Looking in detail at the P-wave velocity data (5.17-5.23), one can observe a general increase of all velocities up to 0.5% strain, with a higher velocity difference in the axial region. At this point, while the horizontal velocities start decreasing, the vertical velocities keep rising up to 0.75% strain, followed by the eventual decrease up to and beyond failure.

The presence of a constant pore pressure, maintained by the servo-controlled, high pressure pore pump, allows for the measurement of the pore volume change (indicated by the thin black line).
The increase of pore space marked the onset of sample’s dilatancy, which occurs between 0.5 and 0.75% strain. This is simultaneous with the decreasing in P-wave velocity, in essence a mirror image.

Table 5.5: P-wave velocities statistics of the 7 experiments run at saturated conditions ($p_c = 35$ MPa, $p_r = 5$ MPa).

<table>
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Figure 5.17: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB14 (\(p_c = 35\) MPa, \(p_p = 5\) MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) From an average P-wave velocity of 5.5 km/s at intact sample, c) an increase of 0.2-0.3 km/s is measured up to 0.24% and d) 0.49% strain. e) followed by no velocity change at 0.75% strain. f) Before sample failure, velocities are 0.5-0.7 km/s lower than the scenario in panel e. g) with post-failure velocities showing E-W horizontal velocities decreased by 1 km/s, and with local increases along raypath direction of 30-40°. Measured pore volume decreases up to a strain of around 0.5%, and is roughly constant up to 0.75% strain (800 s), and denoting the so-called D’ part of the stress-strain curve, after which it increases until sample failure.
Figure 5.18: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB15 ($p_c = 35$ MPa, $p_p = 5$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) From an average P-wave velocity of 5.5 km/s at intact sample, c) an increase of 0.3-0.5 km/s is visible both at 0.24% and d) 0.53% strain, e) followed by steady conditions at 0.75% strain. f) Before the failure, velocities are 0.2-0.3 km/s lower than the previous scenario. g) Post-failure velocities reveal a quasi-homogeneous decrease in velocities of about 0.8 km/s, with a local increase along N-S horizontal directions. Measured pore volume decreases up to 0.53 % strain ($D'$), followed by a constant state up to 0.75% strain and an eventual accelerating increase.
Figure 5.19: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB16 ($p_c = 35$ MPa, $p_p = 5$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) From an average P-wave velocity of 5 km/s at intact sample, c) a homogeneous increase of 0.5 km/s is visible at 0.24%, d) which become 0.3 km/s at 0.53% strain. e) Following steady conditions occur at 0.75% strain, f) while before failure, velocities are 0.8 km/s lower than the previous scenario. g) Post-failure velocities reveal again a quasi-homogeneous decrease in velocities of about 0.8 km/s, with local increases along directions plunging 40°. Measured pore volume decreases up to 0.48 % strain ($D'$), followed by an accelerating increase.
Figure 5.20: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB17 \( (p_c = 35 \text{ MPa}, p_p = 5 \text{ MPa}) \). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) A non-isotropic P-wave velocity (average = 5 km/s) characterizes the intact sample. c, d & e) The deformation proceeds with homogeneous variations of P-wave velocities of about 0.1-0.2 km/s up to 0.50% strain. f) Before failure, velocities are 0.2-0.3 km/s lower than the previous scenario, g) with post-failure conditions having negative variations of 0-8 km/s, with the highest difference in the SW quadrant. Measured pore volume remains stable up to 0.50% strain \( (D') \), followed by an accelerating increase.
Figure 5.21: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB18 ($p_c = 35$ MPa, $p_p = 5$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) A non-isotropic P-wave velocity (average = 5 km/s) characterizes the intact sample. c, d & e) The deformation proceeds with homogeneous variations of P-wave velocities of about 0.1-0.2 km/s up to 0.46% strain. f) Before failure, velocities are 0.2-0.3 km/s lower than the previous scenario, g) with post-failure conditions having a quasi-homogeneous negative variations of 0-8 km/s. Measured pore volume decreases up to 0.46% strain ($D'$), followed by steady conditions up to 0.6% strain and an accelerating increase.
Figure 5.22: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB19 ($p_c = 35$ MPa, $p_p = 5$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample presents an average P-wave velocity of 5.5 km/s. c & d) As the deformation proceeds, all velocities increased by 0.2-0.5 km/s up to 0.39% strain. e) At 0.57% strain velocities do not change, f) while at 0.71% strain a general decrease of 0.2 km/s occurs. g) Post-failure conditions present negative variations of 1 km/s, with some area having reduced velocity change. These are N-S horizontal directions, axial region and SW quadrant. Measured pore volume decreases up to 0.39 % strain, followed by steady conditions up to 0.5% strain ($D'$) and by an accelerating increase.
Figure 5.23: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB36 ($p_c = 35$ MPa, $p_p = 5$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample presents an average P-wave velocity of 5.3 km/s. c & d) As the deformation proceeds, all velocities increased by 0.2-0.5 km/s up to 0.51% strain. e) At 0.73% strain velocities do not change, f) while at 0.80% strain a general decrease of 0.2 km/s occurs. g) Post-failure conditions present complex P-wave velocity difference, ranging from 0.2 to 1 km/s. Measured pore volume decreases up to 0.35% strain, followed by steady conditions up to 0.5% strain ($D'$) and by an accelerating increase.
Figure 5.24 illustrates the changing P-wave anisotropy for all saturated ($p_p = 5$ MPa) experiments. A low level of anisotropy (values from 1 to 1.1) is initially present in all samples and increases steadily up to approximately 1.2 just before sample failure. In all experiments, apart from EB14 and EB16, velocity surveys are paused before formation of the shear fault (sample failure) so as not to interfere with the recording of the continuous data by the Richter units. However, for these two experiments it is still possible to record a sharp and significant increase in anisotropy right after the sample failure. In addition, although halfway through the experiment the pore volume change (volumetric strain) turns positive (i.e. onset of dilatancy, D’), this is not associated to any variation in the linear increase of anisotropy, reflecting the nature of the continuous nature of crack opening during stress increase.
Figure 5.24: P-wave anisotropy superimposed on the stress-time plot for experiment EB14, EB15, EB16, EB17, EB18, EB19 and EB36, run at saturated conditions ($p_c = 35$ MPa, $p_p = 5$ MPa). All experiments are characterized by a linear increase in P-wave anisotropy, with accelerating only around the time of failure. This is clearly visible in experiments a) EB14 and c) EB16, where the surveys are recorded continuously every minute.
5.4.2. PASSIVE EVENTS (ACOUSTIC EMISSION)

Figure 5.25 shows the hit rate trend for 6 experiments run at saturated conditions ($p_p = 5\text{MPa}$) superimposed on the stress-time curve and pore volume change. No AE activity occurs up to 800-1000 s, coincident with a decrease in the pore volume. As the pore volume increases, the hit rate increase exponentially (average $R^2$ of the exponential fit: 0.7897) up to the last 15-30 s when the trend becomes supra-exponential. After the failure, the hit rate rapidly decreases and ceases. At the failure, an average of 100 hits are recorded, ranging from 80 to 140 hits. Sample EB15 (b), having the lower half conduit 3 mm wide and the upper half 9.5 mm wide, presents the fastest AE build-up after the onset of dilatancy, and the slowest increase before failure.

Figure 5.26 shows the temporal variation of the $b$-value through the final minute of the deformation stage for 6 water saturated ($p_p = 5\text{MPa}$) experiments. As for the dry experiments, the temporal variation of the $b$-value is similar across all experiments using similar effective pressure conditions, presenting a flat – oscillatory pattern during most of the final minutes with a sharp decrease less than a second before the time of failure of the sample. Only EB17 (panel d) manifests an early slower decrease of the $b$-value 3 seconds before the stress drop. Interestingly, under water saturated conditions, the $b$-value remains at a low level even after some 2-20 seconds post failure before recovering to pre-failure values, compared to the $b$-value data at a similar time during dry deformation where recovery is much quicker.
$p_c = 35 \text{ MPa}; p_p = 5 \text{ MPa}; p_{\text{eff}} = 30 \text{ MPa}$

Figure 5.25: AE hit rate (line and dots) superimposed on the stress – time curve (thick line) and pore volume change (thin line) for experiment EB14, EB15, EB16, EB17, EB18 and EB19, run at saturated conditions ($p_c = 35 \text{ MPa}, p_p = 5 \text{ MPa}$). Each curve is characterized by i) little AE activity up to 800-1000 s; ii) exponential increase of hit rate (black dotted line, with equation and $R^2$); iii) supra-exponential growth in the last 30-15 seconds preceding the failure; iv) hit rate maximum at the time of failure; v) post-failure hit rate decrease.
Figure 5.26: Temporal variation of the $b$-value (line and dots) superimposed on the stress-time curve (solid line) for experiment EB14, EB15, EB16, EB17, EB19 and EB36, run at saturated conditions ($p_c = 35$ MPa, $p_p = 5$ MPa). Each curve is characterized by: i) flat or oscillatory pattern up to the last second preceding the failure followed by ii) a sharp decrease of $b$-value in the 1 second before failure; iii) $b$-value minima at the time of failure; iv) a stable $b$-value, but at low value for 2 to 20 seconds; v) post-failure rebound in the $b$-value.
The location of the extracted events recorded around the failure of the sample for water saturated \((p_p = 5 \text{ MPa})\) experiments is displayed in Figure 5.27. The average number of located events is 1174, with EB16 (c) having more than twice this number. As a result, while all experiments show a moderate clustering along a NW-SE direction, in experiment EB16 the high number of events masks any preferential direction. Therefore, the use of overlapping temporal windows (Fig. 5.28, 5.29), together with a post-test photography assists with the analysis. Another common feature for all experiments is the low located AE activity 10 seconds after failure, during the slip of the fault.

Sample EB14 has 990 located events (Fig. 5.27a), imaging a NW-SE damage zone, which actually formed when the sample failed (Fig. 5.28a). A few hundred events, with a widespread distribution, are located up to 10 seconds before the stress drop (first 3 windows). However, in the last 10 seconds (4\textsuperscript{th} window), characterised also by the peak stress, a mild clustering along a NW-SE direction starts to appear, which becomes clear in the first 10 seconds of fault slip (5\textsuperscript{th} window).

In sample EB15, 1353 events are successfully located (Fig. 5.27b) showing a NW-SE cluster. Inspection of post-test sample reveals a complex system of almost vertical fractures (Fig. 5.28b). Looking at this point of view, the overlapping windows depict a concentration of AE in the middle-west sector of the sample, some 10-21 seconds (during which peak stress is reached) before failure (3\textsuperscript{rd} window), with a clearer NW-SE direction both 10 seconds before the stress drop (4\textsuperscript{th} window) and beyond (5\textsuperscript{th} window).

As stated above, sample EB16 has the highest number of located events (2599, Fig. 5.27c), which makes it difficult to recognize the NE-SW main fault and its conjugate oriented NW-SE (Fig. 5.28c). The high number of events located around the failure (4\textsuperscript{th} and 5\textsuperscript{th} windows) masks the faults, which can only be observed between 30 and 10 seconds before the stress drop (2\textsuperscript{nd} and 3\textsuperscript{rd} windows), when also the peak stress is reached.

In sample EB17, the located events (932) image a NW-SE damage zone, from the upper left corner to middle-east area of the sample (Fig. 5.27d). This pattern is identified by the overlapping windows (Fig. 5.28d) around the stress drop (4\textsuperscript{th} and 5\textsuperscript{th} windows), with no clear clustering before and during the peak stress (first 3 windows). However the post-test sample is characterized by a NE-SW main fault and a conjugate, not fully developed fracture oriented NW-SE from the bottom to middle-west.
Sample EB19, characterized by using 555 located AEs, reveals a cluster on the middle-east sector (Fig. 5.27e), which is also visible in the post-test photo, where a system of fractures is present (Fig. 5.29a). In particular the temporal evolution of hypocentres images a prominent cluster after the peak stress, 10 seconds before the failure (4th window) and beyond (5th window).

Finally, 612 events are located in EB36 clustered in the NW-SE direction (Fig. 5.27f). Post-test photos confirmed this pattern (Fig. 5.29b), which is only visible in the second immediately after the failure (5th window). While up to 10 seconds before failure no trend is visible (first 3 windows), during the last second (4th window), when also the peak stress occurs, a concentration of hypocentres is visible in the upper middle sector of the sample.
Figure 5.27: locations of the events recorded around the time of the sample’s failure for experiment EB14, EB15, EB16, EB17, EB19 and EB36, run at saturated conditions ($p_c = 35$ MPa, $p_p = 5$ MPa).
Figure 5.28: Temporal evolution of the hypocentres of the experiments a) EB14, (b) EB15, (c) EB16, (d) EB17 ($p_c = 35$ MPa, $p_p = 5$ MPa) and photos (right) of post-test samples.
The average inter-event distance for the suite of saturated ($p_p = 5$ MPa) experiments is shown in Figure 5.30. In the majority of cases there is a decrease in the inter-event distance with time by about 8 mm during the approximately 20 seconds before stress drop. Within this time frame, some more general variations are seen such as the modest increase in the inter-event distance. This reaches a peak at the moment of peak stress, before decreasing sharply once the sample stress starts to decrease and reaching a minimum value shortly after the sample fails. The outlier to this trend is EB15, which is the only experiment in which inter-event distance again starts to increase after the formation of the main shear-fault. The recovery of inter-event distance to high values is concomitant with a differential stress still decreasing, so it is likely that cracks and still being generated.

Figure 5.31 represents the variations of the average $M_L$ as the sample approaches the failure and beyond. The average $M_L$ remains fairly constant up to 1-10 seconds before the stress drop. Whether the average $M_L$ remains constant (EB15 (b), EB17 (d), EB36 (f)), undergoes a slow increase (EB14 (a) and EB19 (e)) or has a complex pre-failure pattern (EB16 (c)) at the time of failure, the average
$M_L$ always seems to increase by approximately 0.5 magnitude. For the particular case of EB16 (c), an early increase and stabilization of the average $M_L$ occurs around 1580 seconds, simultaneously with a marked strain softening regime.

$p_c = 35 \text{ MPa}; p_p = 5 \text{ MPa}; p_{\text{off}} = 30 \text{ MPa}$

![Graphs showing average inter-event distance and stress-time plots for experiments EB14, EB15, EB16, EB17, EB19, and EB36.](image)

Figure 5.30: Average inter-event distance (line and dots) superimposed on the stress-time plot (solid line) for experiment EB14, EB15, EB16, EB17, EB19 and EB36, run at saturated conditions ($p_c = 35 \text{ MPa}, p_p = 5 \text{ MPa}$).
Figure 5.31: Average location magnitude (line and dots) superimposed on the stress-time plot (solid line) for experiment EB14, EB15, EB16, EB17, EB19, and EB36, run at saturated conditions ($p_c = 35$ MPa, $p_p = 5$ MPa).
5.5. DEFORMATION STAGE UNDER SATURATED CONDITIONS (p_p = 16 MPa)

To explore deformation properties at deeper levels, an additional suite of experiments were conducted at higher pore pressure conditions of 16 MPa and confining pressure of 46 MPa, maintaining an effective confining pressure of 30 MPa corresponding to deeper volcano tectonic conditions of approximately 1.6 km. Table 5.6 shows the mechanical data and basic sample properties for experiments carried out at these conditions. Figure 5.32 shows the basic stress–strain data across all samples tested. The samples failed at an average maximum differential stress of 448 ± 15 MPa and a corrected strain of 0.0112 ± 0.0010. The average tangent (Young’s modulus), calculated over the linear elastic part of the stress - strain curve, is 52 ± 6 GPa.

Table 5.6: physical properties and results of the 8 experiments run at saturated conditions (p_c = 46 MPa, p_p = 16 MPa).

<table>
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<th>Sample #</th>
<th>Length (mm)</th>
<th>Diameter (mm)</th>
<th>Conduit d (mm)</th>
<th>Duration (mm:ss)</th>
<th>Max. σ_diff (MPa)</th>
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Figure 5.32: stress – strain plot of the 8 experiments run at saturated conditions ($p_c = 46$ MPa, $p_p = 16$ MPa).

5.5.1. ACTIVE EVENTS (ELASTIC-WAVE VELOCITY)

The evolution in the P-wave velocity throughout the experiments are shown in Figure 5.33 to 5.40 for saturated ($p_p = 16$ MPa) experiments; table 5.7 lists pre-test condition (intact sample) and maximum value of both $62^\circ$ and horizontal ($0^\circ$) P-wave velocities. Velocities of the intact sample range from 5300 to 5600 m/s (averaging 5400 m/s) along the inclined direction and from 4900 to 5500 m/s (averaging 5200 m/s) along the horizontal direction. The velocities at $62^\circ$ increase more significantly as compared to velocity at $0^\circ$, reaching an average of 6000 m/s (from 5900 to 6300 m/s) along the inclined direction, and 5400 m/s (from 5100 to 5700 m/s) horizontally. These values are similar to those measured for $p_p = 5$ MPa.

The common general trend of the temporal variation of P-wave velocity, and generally similar to that observed under dry / saturated conditions ($p_p = 5$ MPa), is characterized by an increase of all velocities up to a strain of 0.5 %, followed by a velocity decrease after 0.75% strain. All velocities keep decreasing up and beyond the stress drop, with a faster rate at the time of failure.
Another common trend is that dilatancy starts between 0.5% and 0.75% strain, simultaneously with the decreasing of P-wave velocities.

Table 5.7: P-wave velocities statistics of the 8 experiments run at saturated conditions ($p_c = 46$ MPa, $p_r = 16$ MPa).

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<tr>
<th>Sample #</th>
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<th>Initial 0°vel (m/s)</th>
<th>Max. 62°vel (m/s)</th>
<th>Max. 0°vel (m/s)</th>
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Figure 5.33: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB20 ($p_c = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) Early in the experiment, intact sample shows an average P-wave velocity of 5.5 km/s, with the lower velocity orientated along the low inclined W-E direction. c) Differential velocities (Velocity of 62° - Velocity of 0° raypath) then increase to 0.2 km/s at a strain of 0.24%, d) increasing to 0.3 km/s at 0.46% strain. After this point, e) only the axial velocities still increase at 0.74% strain with horizontal velocity decreasing, further increasing the differential velocity. f) At 0.80 % strain all velocities decrease by about 0.3 km/s compared with the previous step, g) a trend that accelerates after failure with a large change of 1 km/s (horizontal velocity: the axial region undergoing a less sharp final velocity jump of 0.5 km/s. Measured pore volume change is always positive thorough sample deformation (possibly a leak of the pore pressure system) with an inflexion point at 800 s likely to correspond to the $D'$ time.
Figure 5.34: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB24 ($p_c = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample present an average P-wave velocity of 5 km/s, with a higher-velocity zone along a NNW-SSE direction, plunging at 45°. c) Differential velocities increase of 0.2-0.3 km/s are observed at 0.25% and d) 0.53% strain, e) with only the axial velocities still increasing at 0.74% strain. f) At 0.89% strain all velocities decrease by about 0.3 km/s, g) followed by a quasi-homogeneous decrease of 0.5 km/s after failure. Measured pore volume change is always positive, suggesting a leak of the pore pressure system.
Figure 5.35: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB25 ($p_c = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample present an average P-wave velocity of 5.3 km/s. c) Differential velocities increase of 0.2-0.3 km/s are observed at 0.25% and d) 0.46% strain, e) with only the axial velocities still increasing at 0.74% strain. f) At 0.89% strain all velocities decrease by about 0.3 km/s, g) followed by a sharper decrease of 0.7 km/s after failure, with area of higher velocity-change (~1 km/s) in the east part of the stereonet. Measured pore volume shows a minimum around 0.54% strain ($D'$), followed by an accelerating increase in the pore space.
Figure 5.36: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) $P$-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB27 ($p_c = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample present an average $P$-wave velocity of 5.3 km/s. c) Differential velocities increase of 0.2-0.3 km/s are observed at 0.25% and d) 0.48% strain, e) with only the axial velocities still increasing at 0.77% strain. f) At 0.92% strain all velocities decrease by about 0.3 km/s, followed by a sharper decrease of 1 km/s after failure, with the axial region showing reduced velocity difference. Measured pore volume shows a minimum around 0.48% strain ($D'$), followed by an accelerating increase in the pore space.
Figure 5.37: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB28 ($p_c = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample present an average P-wave velocity of 5.3 km/s. c) Differential velocities increase of 0.2-0.3 km/s are observed at 0.29% and d) 0.48% strain, e) with only the axial velocities still increasing at 0.77% strain. f) At 0.92% strain all velocities decrease by about 0.3 km/s, g) followed by a sharper decrease of 0.5 km/s after failure, with horizontal E-W directions showing higher velocity change (~1 km/s). Measured pore volume change is always positive thorough sample deformation (possibly a leak of the pore pressure system) with an inflexion point at 500 s likely to correspond to the $D^'$ time.
Figure 5.38: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB29 ($p_c = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample present an average P-wave velocity of 5.3 km/s. c) Differential velocities increase of 0.2-0.3 km/s are observed at 0.25% and d) 0.48% strain, e) with only the axial velocities still increasing at 0.77% strain. f) At 0.92% strain all velocities decrease by about 0.3 km/s, g) followed by a sharper decrease of 0.5 km/s after failure, with the axial region showing reduced velocity difference. Measured pore volume shows a minimum around 0.62% strain ($D'$), followed by an accelerating increase in the pore space.
Figure 5.39: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB30 ($p_c = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample present an average P-wave velocity of 5.5 km/s. c) Differential velocities increase of 0.2-0.3 km/s are observed at 0.25% and d) 0.47% strain, e) with only the axial velocities still increasing at 0.77% strain. f) At 1% strain all velocities decrease by about 0.3 km/s, g) followed by a sharper decrease of 0.5 km/s after failure, with NE-SW directions, plunged at 45°, showing higher velocity difference (~1 km/s). Measured pore volume shows a minimum around 0.62% strain ($D'$), followed by an accelerating increase in the pore space.
Figure 5.40: a) Variation of the horizontal (red line and dots) and the 62° (black line and dots) P-wave velocities superimposed on the stress-time plot (thick blue line) and pore volume change (thin black line) for EB34 ($p_C = 46$ MPa, $p_p = 16$ MPa). The arrows in panel (a) indicate the velocity survey shown in the stereonets from (b) to (g). b) The intact sample present an average P-wave velocity of 5 km/s. c) Differential velocities increase of 0.2-0.3 km/s are observed at 0.23% and d) 0.48% strain, e) followed by a diffused decrease at 0.73% strain (-0.1 km/s) and f) at 0.98% strain (-0.5 km/s). g) Post-failure differential velocities are 0.5 km/s slower than pre-failure velocities. Measured pore volume shows a minimum around 0.36% strain ($D'$), followed by an accelerating increase in the pore space.
Figure 5.41 shows the P-wave anisotropy. A low level of anisotropy (values from 1 to 1.1) is initially present in all samples, increasing slowly but steadily as the deformation experiment proceeds, similarly to the saturated conditions with $p_p = 5$ MPa. After failure, the level of anisotropy tend not to increase any further. Only EB34 (h) has survey recorded every minute, with no interruption before the failure. In this case is it possible to observe a constant increase of the anisotropy up to 1.15 until 200 seconds before the failure. At this point there is an acceleration and the anisotropy goes up to 1.25 just before the failure and remain at this level after the stress drop.

Again, using the pore volume change curve, it is likely that the onset of dilatancy (D′) does not affect the behaviour of the anisotropy trend.
Figure 5.41: P-wave anisotropy superimposed on the stress-time plot for experiment EB20, EB24, EB25, EB27, EB28, EB29, EB30 and EB34 ($p_c = 46$ MPa, $p_p = 16$ MPa). All experiments are characterized by a linear increase of the level of anisotropy, which accelerates only 200 seconds before the time of failure (clearly visible in experiments EB36 (h), where the surveys are recorded continuously every minute).
5.5.2. PASSIVE EVENTS (ACOUSTIC EMISSION)

The hit rate for 5 saturated ($p_p = 16$ MPa) experiments is shown in Figure 5.42. In all cases, microseismic activity starts between 1200 – 1400 seconds after the start of the experiments, and somewhat past the D’ stage as measured by volumetric strain. For experiment EB34 (e), a lower strain rate ($5\times10^{-6}$) was used for samples deformation to confirm no effect of strain rate on strength, lasting 700 seconds. This resulted in a delayed onset of AE activity (at 1600 seconds) and delayed failure. However, the hit rate start to increase 400 seconds before failure, consistent with the other experiments at the same saturated conditions.

While the hit rate has a mild, limited in time exponential growth (average $R^2$ of the exponential fit: 0.7453) during the central part of the deformation stage, as per the previous results ($p_p = 5$ MPa), a supra-exponential increase occurs in the last 15-20 seconds before the sample failure. This brings the hit rate up to an average value of 133 hits, with the maximum value (hit rate of 195) recorded for the solid sample EB24 (b). After failure, hit rate decreases sharply and ceases within few seconds after the stress drop.

The effect of dilatancy can be studied in EB25 (panel c) and EB34 (panel e). It is possible to observe, in both cases, that there is no simultaneity between the increase in pore volume and the onset of AE activity, with the onset of dilatancy appearing 300-400 seconds before the increase of hit rate.

The temporal variation of the $b$-value through the final minute of the deformation stage for the saturated ($p_p = 16$ MPa) experiments is shown in Figure 5.43. A flat/oscillatory decrease characterises the $b$-value variation up to a time approximately between 2-10 seconds before sample failure, after which, the $b$-value shows a sharp and rapid decrease to a local minimum (in 7 of the 8 experiments) within the last second around at the time of failure.

However, experiment EB27 (d) shows a slightly more unusual variation. Here, a local minimum occurs between the failure and final stress drop, coincident with an earlier minor stress drop. This is then followed by a sharp increase and an absolute minimum of the $b$-value at the final stress drop of sample failure.
$p_c = 46$ MPa; $p_p = 16$ MPa; $p_{eff} = 30$ MPa

Figure 5.42: AE hit rate (line and dots) superimposed on the stress–time curve (thick line) and pore volume change (thin line) for experiment EB20, EB24, EB25, EB28 and EB34 ($p_c = 46$ MPa, $p_p = 16$ MPa). Each curve is characterized by i) little AE activity (flat part) up to 1000-1200 s (1500 s for EB34 (e)); ii) exponential increase of hit rate (black dotted line, with equation and $R^2$); iii) supra-exponential growth in the last 20-15 seconds preceding the failure; iv) hit rate maximum at the time of failure; v) post-failure hit rate decrease.
$p_c = 46$ MPa; $p_p = 16$ MPa; $p_{\text{eff}} = 30$ MPa

Figure 5.43: temporal variation of the $b$-value (line and dots) superimposed on the stress-time curve (solid line) for experiment EB20, EB24, EB25, EB27, EB28, EB29, EB30 and EB34 ($p_c = 46$ MPa, $p_p = 16$ MPa). Each curve is characterized by i) flat or oscillatory pattern up to the last second preceding the failure followed by ii) a sharp increase of $b$-value 1 second before failure followed by iii) a sharp decrease with iv) $b$-value reaching a minimum at the time of failure; v) $b$-value stable at lower level for 2 to 20 seconds; vi) post-failure $b$-value increase.
Figure 5.44 shows the hypocentres of the extracted events recorded around failure for all saturated (pp = 16 MPa) experiments. The average number of located events is 870, ranging from 426 to 1423. Clustering along a preferential direction can be observed for only 3 experiments: EB25 (c), EB28 (e) and EB30 (g). The temporal evolution of these hypocentres, shown as overlapping time windows (Fig. 5.45 and 5.46), sheds light on the timing of clustering. As for the experiments with pp = 5 MPa, the post-failure fault slip regime features few tens of located events.

EB20 has 912 located events, with their hypocentres not aligned along any direction (Fig. 5.44a). Post-test photography reveals the presence of throughout NE-SW fault with a conjugate NW-SE fracture in the upper half of the sample (Fig. 5.45a). Around the peak stress, no clustering appears (first 3 windows). The two fractures start to be resolved during the 11 seconds preceding the failure (4th window) and beyond (5th window), forming the NW-SE and the NE-SW fracture respectively. The late fault slipping regime (6th window) is characterized by hypocentres along this conjugate fracture.

In EB24, only 426 events were successfully located (Fig. 5.44b), representing the lower number of locations of this particular sample. A NE-SW fault characterizes the post-test sample (Fig. 5.45b), with similar timing as EB20. In fact, during the first 3 windows, no clustering is noticed, which only appears just before failure (4th window); a mild cluster around a NE-SW direction, which becomes less clear beyond the stress drop (5th and 6th windows).

A marked NW-SE fault is imaged by the 1331 located events in EB25 (Fig. 5.44c), which correctly depict the post-test scenario (Fig. 5.45c). In this case, clustering starts between 21-10 seconds before the failure (3rd window), focusing in the middle of the sample. Just before the stress drop (4th window), one can notice a more defined NW-SE fracture, which is migrated upwards, while beyond the failure (5th window) the hypocentres migrates downwards.

Sample EB27 shows a concentration of the hypocentres (576) in the central part of the specimen (Fig. 5.44d). Inspection of the post-test sample reveals a NW-SE (Fig. 5.45d), which is only imaged 20 seconds before the failure and few seconds before the stress peak (2nd and 3rd windows). In the 11 seconds preceding the failure (4th window), characterized also by the occurrence of a minor stress drop, a NE-SW fracture, constrained to the central sector, is located. Immediately after failure (5th window) a cluster of the hypocentres in the middle-east is present, but a clear orientation is not visible.
Similarly to EB25, a NW-SE fault is well imaged by the 730 hypocentres in EB28 (Fig. 5.44e). Post-test photography confirms this fault together with a conjugate fault at the same dip angle, merging in the lower part of the sample (Fig. 5.46a). Temporally, the NW-SE fault is imaged starting 11 seconds before failure (4th window), with clear early fault slipping along the same direction (5th window). However there is no evidence of clustering which traces the conjugate fault.

EB29 features 1423 located events, widespread throughout the specimen, but with few sparse AEs in the bottom half (Fig. 5.44f). Inspection at the post-test sample reveals a NW-SE fault (Fig. 5.46b), which is imaged up to 10 seconds before the failure (3rd window) and during late fault slipping (6th window). Around the stress drop (4th and 5th windows), no clear orientation is visible, possibly masked by the high number of located events (716 and 636 hypocentres respectively).

In EB30 832 located events image a NW-SE fault (Fig. 5.44g), which is confirmed by the post-test photo (Fig. 5.46c). In particular, a mild clustering along such direction appears early, 40-30 seconds before failure (1st window), becoming clearer after the peak stress, 21 to 10 seconds before the failure (3rd window). As the sample approaches failure (4th window) and beyond (5th window), AE activity remains and intensifies along the NW-SE direction.

Finally EB34 presents 713 located events with no preferential orientation (Fig. 5.44h). Post-test sample reveals a system of 2 fractures, crossing each other in the centre of the sample (Fig. 5.46d). Between the peak stress and the failure (2nd, 3rd and 4th windows), the NE-SW fracture is well imaged by the hypocentres. Immediately after the stress drop (5th window), such orientation is no longer visible but reappears again during late fault-slipping (6th window), even if imaged by only a small number of events.
Figure 5.44: locations of the events recorded around the time of the sample’s failure for experiment EB20, EB24, EB25, EB27, EB28, EB29, EB30 and EB34 ($p_c = 46$ MPa, $p_p = 16$ MPa).
Figure 5.45: Temporal evolution of the hypocentres of the experiments a) EB20, b) EB24, c) EB25, d) EB27 (\(p_c = 46\) MPa, \(p_p = 16\) MPa) and photos of post-test samples.
Figure 5.46: Temporal evolution of the hypocentres of the experiments a) EB28, b) EB29, c) EB30, d) EB34 (p_r = 46 MPa, p_p = 16 MPa) and photos of post-test samples.
Figure 5.47 shows the temporal evolution of the average inter-event distance for all experiments at saturated conditions ($p_s = 16$ MPa). The general behaviour is characterized by a high degree of clustering at the time of failure, with the inter-event distance decreasing from approximately 28 mm to 22 mm when the stress drop occurs. The negative slope of the curve starts from the beginning of the dataset and show local minima (EB25 (c) and EB27 (d)) which corresponds to the minor stress drops occurring before the main fracture event. Experiment EB29 (panel f) presents two minima in the average inter-event distance and a major stress drop preceded by an earlier decrease in differential stress. However, for this experiment, a local maximum (between the two minima) occurs at the time failure.

Finally, the average $M_L$ is shown in Figure 5.48. For the majority of experiments the average $M_L$ remains fairly constant up to the moments of failure, when a significant increase is recorded varying from 0.2 (EB24, panel b) magnitude to nearly 1 magnitude (EB28, e). In EB25 (c) and EB27 (d), a previous local maximum occurs 13 and 3 seconds respectively before the failure. As observed for the inter-event distance (Fig. 5.46), a minor stress drop preceded the main stress drop.
Figure 5.47: average inter-event distance (line and dots) superimposed on the stress-time plot (solid line) for experiment EB20, EB24, EB25, EB27, EB28, EB29, EB30, and EB34 ($p_c = 46$ MPa, $p_p = 16$ MPa).
Figure 5.48: average location magnitude (line and dots) superimposed on the stress-time plot (solid line) for experiment EB20, EB24, EB25, EB27, EB28, EB29, EB30 and EB34 (p_c = 46 MPa, p_p = 16 MPa).
5.6. RELEASE STAGE IN LOW TEMPERATURE CONDITIONS (WATER)

Up until this point, the aim of the experimental program was to determine the mechanical and elastic wave properties of Etna Basalt, under two relevant conditions of equivalent depth. With a fracture damage zone established in the sample, and imaged with AE and other rock physics tools, the movement of pore fluids is now carried out to stimulate laboratory Low-Frequency events analogous to those measured in volcanic settings but under known conditions of pressure and temperature.

Table 5.8 summarises the properties and the results of the 3 venting experiments carried out at room temperature using water as pore fluid, while Figure 5.49 shows the cumulative hits superimposed on the pore pressure vs time curve. EB16 was carried out using a pore pressure of 5 MPa, whilst for experiments EB29 and EB34 a higher pressure of 16 MPa was used. In all cases, the effective pressure is 30 MPa and the axial conduit is 3 mm wide.

Venting duration is measured at the top pore pressure transducer (the furthest transducer from the venting point), from initial time $t_0$ to the time when the top pore pressure reached half of the starting $p_p$. While EB16 and EB34 have similar venting duration (0.0126 and 0.0165 s), the venting in EB29 is about 3 times slower (0.0481 s). However, in these three experiments, the top pore pressure steadily decreases to zero within 120 seconds and the bottom pore pressure goes to zero immediately.

Other than EB16, where the cumulative hits grows constantly from $t_0$, AE data from EB29 and EB34 show that well over 50% of the hits occurs in the first second after the pore pressure release (also called “venting”) with a sharp deceleration in hit rate after this.

Table 5.8: conditions and results of the 3 venting experiments run at room temperature ($T = 25°C$), using water as pore fluid.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>$p_p$ (MPa)</th>
<th>Conduit (mm)</th>
<th>Venting duration (s)</th>
<th>Cumul. hits</th>
</tr>
</thead>
<tbody>
<tr>
<td>EB16</td>
<td>5</td>
<td>3</td>
<td>0.0126</td>
<td>108</td>
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<tr>
<td>EB29</td>
<td>16</td>
<td>3</td>
<td>0.0481</td>
<td>137</td>
</tr>
<tr>
<td>EB34</td>
<td>16</td>
<td>3</td>
<td>0.0165</td>
<td>2121</td>
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</table>
Figure 5.49: cumulative hits (grey dashed line) superimposed on the pore pressure – time plot (solid lines) for the 3 experiments with liquid water as pore fluid (T = 25°C). At the time of release (0 s), there is an instant decrease of both pore pressures and the onset of AE hits. a) While EB16 shows a steady increase of hits, b) EB29 and c) EB34 manifest a sharp increase at t₀, followed by a deceleration in the hit rate. The decay of the top pressure is not similar between the three experiments, but they all decreased to 0 within 120 s.
Figure 5.50, 5.51 and 5.52 show a 100-ms-long window with top (black line) and bottom (green line) pore pressure data (Panel A) sampled with a high-speed Data Acquisition Digitizer, together with the same time window of the continuous Acoustic Emission (Richter) data (blue line), using the low-frequency sensor with the highest SNR in each experiment. The spectrogram (Panel B) of the waveform represents the power-time-frequency distribution of the signal in time, across the 100 ms long venting of the pore fluid pressure and the stacked spectrum (Panel C) is the frequency content of the selected waveform in Panel A (dashed vertical lines), from the onset to the end of the signal. The spectrum is calculated by stacking multiple spectrum over a 204.8 μs-long-window with 50% of overlap.

In EB16 (Fig. 5.50), while the bottom $p_p$ reaches zero 10 ms after the release, the top $p_p$ starts decreasing 5 ms after $t_0$, slowly decaying to 1.5 MPa over the following 20 ms, and stabilizing at this pressure. AE activity starts around 10 ms, when the bottom $p_p$ is already at 0 and the top $p_p$ is slowly decaying, lasting for 15 ms. A few transient AE are recorded immediately after this 15-ms-long signal, with a high-amplitude event occurring at 48 ms. The power of the long-duration signal is concentrated over its entire duration, showing a broadband character, lying almost constantly above the -6 dB level from 50 to 180 kHz.

Experiment EB29 (Fig. 5.51) shows a slower decay of the $p_p$ compared to EB16. Once the pressure is released, both top and bottom $p_p$ slowly decrease to zero after 90 ms, following a similar decay rate. Some AE activity occurs between 30 and 40 ms, with a higher-amplitude signal between 42 and 58 ms. Few more transient are recorded after the 16-ms-long signal. This long-duration signal is characterized by having the power concentrated below 110 kHz, with some transient at higher frequency. Up to 140 kHz, the frequency spectrum lies around -6 dB, peaking at 110 kHz. Beyond 140 kHz a steady decrease in amplitude is observed in the spectrum.

Similarly to EB16, while the bottom $p_p$ starts decreasing 5 ms after $t_0$, the top $p_p$ decays around 10 ms in EB34 (Fig. 5.52). However the bottom $p_p$, after a fast decay, slows down around 10 ms (at 7 MPa) slowly decreased to 0 in the following 40 ms. The top $p_p$ instead reaches 5 MPa within 20 ms after $t_0$, slowing down and stabilizing at 1 MPa in the following time. The onset of AE activity is first detected at 11 ms, soon after the top $p_p$ started decreasing. The signal lasts 16 ms and it is followed by several other transients. This long-duration event qualitatively presents a broadband character, with a frequency spectrum lying from 40 to 160 kHz above the -6 dB level.
Figure 5.50: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB16 ($p_p = 5$ MPa, $T = 25^\circ$C).

Figure 5.51: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB29 ($p_p = 16$ MPa, $T = 25^\circ$C).
From the continuous waveforms, AE events are extracted forming a dataset of 70 events in EB16, 34 events in EB29 and 56 events in EB34. The analysed waveforms of each event are taken from the low-frequency sensor recording the highest amplitude. When two sensors recorded similar amplitudes, the waveform showing the highest SNR is considered.

The events show a high-degree of waveforms similarity (Fig. 5.53), which is evaluated with the cross-correlation analysis and using the bridging technique. In all experiments more than 90% of events are correlated with a window-length of 20 μs, which are all grouped in a single family (apart in EB34 with window length of 20μs). The degree of similarity steadily decreases with increasing window length. However, even with the longest length considered (generally corresponding to half of the AE event duration), around 50% of the events are correlated, within 2 – 5 families.
Figure 5.53: percentage of correlated events (top panel) and number of families (bottom panel) recognized for a) EB16 ($p_p = 5$ MPa, $T = 25°C$), b) EB29 and c) EB34 ($p_p = 16$ MPa, $T = 25°C$), using the bridging technique. a) In EB16 more than 70% of the 70 events are correlated, forming 1-3 families; b) in EB29 high percentage of correlated events, cluster in a single family, are presents with window length up to 20 μs, decreasing to around 50% and in 3 families with length of 50μs; c) in EB34 a steady decrease of similarity is observed, starting from 100% of the 56 events (and one family) at length = 10 μs, to less than 50% (and 4 families) at length = 50μs.

The master event and stacked spectra of each experiments in shown in Fig. 5.54. The master event is picked by cross-correlating each pair of events over a window-length of 30 μs using a correlation coefficient of 0.7. The event having the most number of correlated events is the master event (left panel). For each of the correlated events, the frequency spectrum is calculated over a window of 102.8 μs, starting from the arrival time. These spectra are then stacked, forming the stacked spectrum (middle panel). The master event and its spectrum are individually shown on the right panel (upper and lower segment respectively).

In EB16 (Fig. 5.54a) 20 events are correlated with the master event (representing 30% of the dataset), and their spectra form a stacked broadband spectrum. In detail, the master event is a 30-μs-long signal with a broadband spectrum over the frequency band 60-160 kHz.

Experiment EB29 (Fig. 5.54b) has 8 correlated events with the master event (constituting 26% of the total extracted events), forming a stacked single-peak spectrum. The master event is has 80-μs-long signal with a frequency peak at 120 kHz at the centre of a FWHM of 70 kHz (80 – 150 kHz).
In EB34 (Fig. 5.54c) 11 events are correlated with the master event (forming 21% of the dataset), with a stacked spectrum having a broadband character over the lower frequency (below ~120 kHz). The master event is a 60-μs-long signal with a spectrum having a FWHM of 80 kHz (from 40 to 120 kHz).

**Figure 5.54:** correlated event (left panels), their frequency spectra and stacked spectrum (middle panels) and waveforms and spectrum of the master event (right panels) for a) EB16 ($p_p = 5$ MPa, $T = 25^\circ$C), b) EB29 and c) EB34 ($p_p = 16$ MPa, $T = 25^\circ$C).
The dominant frequency and the FWHM during the pore pressure release is shown in Figure 5.55. In the upper part of each panel, the dominant frequency is superimposed on the $p_p$ – time plot, which is designed to better understand the temporal variation of frequency as the pore pressure is released. The FWHM is shown by event number, allowing a better visualisation of the parameter as the majority of the events occur over a narrow time-window.

For experiments EB29 (Fig. 5.55b) and EB34 (Fig. 5.55c) it is clear that the majority of events occur close to the time of venting, with few more events occurring in the following seconds. However, experiment EB16 (Fig. 5.55a) exhibits a slightly longer lasting swarm of activity. In all cases the dominant frequencies are well distributed over the frequency band analysed (40 – 200 kHz), with not a predominant frequency in neither cases.

This characteristic also emerges from the FWHM data. Here, experiment EB16 (Fig. 5.55a) shows large FWHM, covering more than half the bandwidth analysed, from 50 to 180 kHz. Experiment EB29 (Fig. 5.55b) has a smaller FWHM for the majority of the events, lying between 20-30 kHz. The frequency is widely distributed before 0.09 s and after 2.5 s, while between this window time a cluster around 100 kHz can be seen. Finally, experiment EB34 (Fig. 5.55c) straddles these two end-members, with FWHM values around 40-50 kHz, showing also a triangular distribution of the dominant frequency: during the first 0.09 s, the frequency increases from 60 to 150 kHz. Values remain stable up to 2.5 s after $t_0$, followed by a decrease back to 50-60 kHz.
Figure 5.55: temporal evolution of the dominant frequency (top panels) and variation of the FWHM for experiments a) EB16 ($p_p = 5$ MPa, $T = 25^\circ$C), b) EB29 and c) EB34 ($p_p = 16$ MPa, $T = 25^\circ$C).
5.7. RELEASE STAGE IN HIGH TEMPERATURE CONDITIONS (WATER)

Table 5.9 lists the properties and basic output of 7 venting experiments carried out at the highest temperature of 175°C using water as pore fluid (to induce phase change from liquid water to a bubbly liquid and steam). Raw data from these experiments is shown in Figure 5.56 in terms of the cumulative hits superimposed on the pore pressure vs time curve. All experiments use the same effective pressure of 30 MPa, with EB15, EB18, EB19 and EB36 having \( p_p = 5 \) MPa and EB20, EB27 and EB28 having \( p_p = 16 \) MPa.

Three different geometries for the axial conduit are being used. The standard 3-mm conduit was used for the majority of the experiment, however, two additional conduit profiles: of either a straight 9.5 mm-wide conduit, or a combination 3 mm to 9.5 mm step change (half way through the sample axis) to investigate any effect of the conduit size on the resonance mode of the generated AE. Despite this, all experiments have similar venting duration, from 10.6 ms to 18.8 ms, with the top pore pressure stabilizing at 0.8 MPa over the whole period of time analysed.

Compared to venting at room temperature, the cumulative number of hits is higher at these higher temperatures, ranging from 251 to 2626 hits. In all experiments, at least 80% of the hits occurs in the first second after the release, with a sharp deceleration in the following time period.

Table 5.9: conditions and results of the 7 venting experiments run at 175°C, using water as pore fluid.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>( p_p ) (MPa)</th>
<th>Conduit (mm)</th>
<th>Venting duration (s)</th>
<th>Cumul. hits</th>
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<td>0.0188</td>
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</table>

*the recording stops before the 120° seconds after the release.
Figure 5.56: cumulative hits (grey dashed line) superimposed on the pore pressure – time plot (solid lines) for the release stages at high temperature ($T = 175^\circ C$). At the time of release (0 s), both top and bottom $p_p$ decrease instantaneously, stabilizing at different pressures. a) In EB18, b) EB19 ($p_p = 5$ MPa), c) EB27 ($p_p = 16$ MPa) and f) EB36 ($p_p = 5$ MPa), bottom $p_p$ reach 0 within the first second, while the top $p_p$ stabilizes around 0.8 MPa. e) EB15 ($p_p = 5$ MPa), presenting a complex conduit geometry, bottom $p_p$ stabilize around 0.8 MPa, while top $p_p$ immediately decreases to 3 MPa, then slowly decays to 1 MPa within 10 seconds after the release, stabilizing at such pressure. d) In EB28 ($p_p = 16$ MPa) both $p_p$ reach 1.5 MPa within the first second, slowly decrease to 0.9 MPa and stabilize up to 80 seconds after failure. At this point bottom $p_p$ instantaneously decreases to 0.2 MPa, slowly recovering to 0.5 MPa, with the top $p_p$ having a slow decay towards 0.7 MPa. The onset of AE hits is simultaneous with the release, with the majority of hits recorded within the first second and few more hits recorded beyond this point. d) Only EB28 shows a moderate AE activity after the first second, with additional 300 hits (20% of the total).
Figure 5.57 – 5.62 show the 100-ms-long waveforms (Panel A), its spectrogram (Panel B) and the stacked frequency spectrum (Panel C) of the selected signal for the release stages at high temperature. A common feature for all experiments is the concentration of frequency around two or three spectral peaks.

During experiment EB18 (Fig. 5.57) the decay of the bottom \(p_b\) reaches 0 within 10 ms after the release; top \(p_t\) starts decreasing at 8 ms, reaching 0 at 30 ms before recovering at 1.5 MPa approximately 60 ms after the release, and decreasing once more at 0.8 MPa. AE activity starts as the top \(p_t\) decreases from 5 MPa, and lasts for 27 ms. AE transients are recorded after this long-duration signal. Towards the end of the 100-ms-long window (80 ms) and after the top \(p_t\) oscillation, a second, lower-amplitude, long-duration signal occurs, lasting 10 ms. The spectrogram shows that both long-duration signals have power clustered around the 100 and 170 kHz, with a weak cluster around 50 kHz. The stacked spectrum of the first long-duration signals reveals a frequency peak at 105 kHz with a FWHM of 30 kHz, and only the peak at 170 kHz reaching the -6 dB level.

In EB19 (Fig. 5.58) both top and bottom \(p_b\) behaves similarly to the previous experiment, with the instantaneous decrease (5 ms) of the bottom \(p_b\) and the delayed (8 ms) decay of the top \(p_t\), showing oscillations (55 ms). AE activity starts when bottom \(p_b\) reaches 0 (after 5 ms), lasting 15 ms. Two high amplitude transients occur at 20 ms and 25 ms, while low-amplitude long-duration activity is observed at 60 ms, following the top \(p_t\) oscillation. The 15-ms-long signal show concentration of power between 80 and 160 kHz, with dominant peaks at 95 kHz and smaller peak at 150 kHz.

For experiment EB27, the top \(p_t\) temporarily reaches 0 before the bottom \(p_b\) (Fig. 5.59). In particular, bottom \(p_b\) commences a rapid decay at 5 ms, slows down around 9 ms when the pressure is 2 MPa, finally reaching 0 at 30 ms. Top \(p_t\) start decreasing after a short delay (10 ms), temporarily reaching 0 at 19 ms and oscillating afterwards. As seen earlier, AE activity begins together with the decay of the top \(p_t\), with a duration of 25 ms. Some transients are recorded in the immediate 20 seconds after the end of the long-duration signal, with additional events towards the end of the window under investigation. Similarly to the previous case, the power of the 25-ms-long signal lies between 80 and 160 kHz, as shown by the stacked spectrum. However the second peak (160 kHz), does not reach the -6 dB in this experiment.
In EB28 (Fig. 5.60) both \( p_p \) follow a similar path for almost the entire 100-ms-long window, but with a short time offset. The bottom \( p_p \) reaches 1.5 MPa after 8 ms before slowing down and stabilizing at 0.8 MPa at 25 ms. The top \( p_p \) begin to decay at 8 ms, reaching 0.8 MPa at 20 ms and remaining at this value. AE activity initiates at 3 ms and lasts for 38 ms. Some discrete events are recorded later (e.g. 34 ms and 55 ms). The long-duration signal presents a banded spectrogram, with 2 clear frequency cluster at 100 and at 160 kHz and a weaker cluster around 40 kHz. The 100 kHz and the 160 kHz continue to be visible after the long-duration signal ceased. The stacked spectrum shows only the peak at 100 kHz, having a FWHM of 20 kHz, while both the peak at 40 kHz and the peak at 160 kHz do not pass the -6 dB threshold.

In EB15 (Fig. 5.61) the bottom \( p_p \) fast decays within the first 5 ms to 1 MPa, followed by a slower decrease to 0.3 MPa. The top \( p_p \) starts to decrease at 8 ms, reaching 2 MPa in 10 ms, before showing a downwards oscillatory course reaching 3.5 MPa approximately 90 ms after fluid release. The onset of AE activity is simultaneous with the decay of the bottom \( p_p \) and lasts 17 ms. The frequency content of this signal is characterized by a peak around 100 kHz, with a FWHM of 30 kHz (80 – 110 kHz) and, importantly, a new second (narrower) peak around 60 kHz.

Finally, experiment EB36 (Fig. 5.62) presents a sharp decrease of the bottom \( p_p \), which reaches 1 MPa at 5 ms, before oscillating around 1 and 2 MPa (5 – 20 ms) and stabilizing at 0.5 MPa. The top \( p_p \) steadily decreases at 10 ms reaching 2 MPa (20 ms), slowing down maintaining a pressure of 1 MPa. AE activity commences at 11 ms, showing high amplitudes in the first 5 ms, and continues for further 43 ms (more visible in the spectrogram). A second long-duration signal begins at 57 ms. The spectrogram of the entire 100-ms-long window clearly displays 3 frequency peaks, particularly high at the onset of the first long-duration signal. These peaks are at 50, 95 and 160 kHz, with the peak at 95 kHz being the more powerful in both long-duration signals. The three peaks overcome the -6 dB level in the stacked spectra of the first long-duration signal.
Figure 5.57: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB18 ($p_p = 5$ MPa, $T = 175^\circ C$).

Figure 5.58: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB19 ($p_p = 5$ MPa, $T = 175^\circ C$).
Figure 5.59: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB27 ($p_p = 16$ MPa, $T = 175°C$).

Figure 5.60: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB28 ($p_p = 16$ MPa, $T = 175°C$).
Figure 5.61: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB15 ($p_p = 5$ MPa, $T = 175^\circ$C).

Figure 5.62: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 100 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB36 ($p_p = 5$ MPa, $T = 175^\circ$C).
Analysis of the waveform similarity is shown in Figure 5.63 and 5.64. A similar trend of waveforms similarity is calculated for EB18 (Fig. 5.63a) and EB19 (Fig. 5.63b), when respectively 30% (of 14 events) and 20% (of 12 events) of events are correlated when length = 50 μs. However while 2 families are generally recognized in EB18, between 1 and 2 families are detected in EB19.

Experiment EB20 (Fig. 5.63c) has the highest number of extracted events (131), which show a high degree of similarity (more than 80% of correlated events) even at length = 50 μs. For small window length (up to 20 μs) the correlated events are grouped in a single family, which then increase (up to 4 families) with the increase of length.

Experiment EB27 (Fig. 5.63d) and EB28 (Fig. 5.64a) have similar number of extracted events, 38 and 31 respectively, and similar percentages of correlated events with the increase of window length, which remain above 50% even with the maximum window length considered. In terms of families, while in EB27 the length of 30 μs constitute a benchmark, separating the region with one family (< 30 μs) and the region with 3-4 families (> 30 μs), in EB28 a constant increase of families with increasing length is observed.

In experiment EB15 (Fig. 5.64b) 14 events are extracted showing a high degree of similarity only for window length below 20 μs (80% of correlated events), which decrease sharply reaching less than 20% of correlated events for length of 50 μs. Apart at length = 30 μs, all correlated events are grouped in a single family.

Finally, EB36 (Fig. 5.64c) has 72 extracted events and, similarly to EB20, shows a degree of waveforms similarity (around 80% of correlated events) even at long window (50 μs). In this case, however, the correlated events are grouped in up to 2 families.
Figure 5.63: percentage of correlated events (top panel) and number of families (bottom panel) recognized for a) EB18, b) EB19 (pp = 5 MPa, T = 175°C), c) EB20 and d) EB27, (pp = 16 MPa, T = 175°C) using the bridging technique. For details see text.
Figure 5.64: percentage of correlated events (top panel) and number of families (bottom panel) recognized for a) EB28 (pp = 16 MPa, T = 175°C), b) EB15 and c) EB36 (pp = 5 MPa, T = 175°C) using the bridging technique. For details see text.

Figure 5.65 illustrates the master event and its spectrum, with the correlated events and the stacked spectrum. A common characteristic emerging from these figures is the double frequency peak, which is shown by stacked spectrum as well as the master spectrum, and identified using a criterion of half maximum power (dashed horizontal line).

In EB18 (Fig. 5.65a) 6 events are highly correlated with the master event out of a total of 14, with the resulting stacked spectrum revealing a single peak around 100 kHz. The master event is an 80-μs-long signal and its spectrum shows a single peak at 105 kHz and a FWHM of 40 kHz. Two more peaks, at 50 and 150 kHz, can be seen below the line of half maximum (dashed horizontal line).

Experiment EB19 (Fig. 5.65b) presents 3 correlated events (with correlation coefficients not higher than 0.75) with the master event (forming 33% of the dataset), whose stacked spectrum shows a clear double peak characteristic. The spectrum of the master event, a 102.4-μs-long signal, does not reveal this second peak. The single peak lies at 120 kHz and has a FWHM of 30 kHz.

In EB20 the double frequency peak characteristics emerge from both the stacked spectrum of the 45 highly correlated events (35% of the total 131 events) and from the spectrum of the master event (70 μs long) (Fig. 5.65c). These two peaks lie at 100 and 155 kHz, with the higher peak having a FWHM of 35 kHz.
In EB27 (Fig. 5.6d), the master event (a 102.4-μs-long signal) correlates 10 events, forming 29% of the dataset in total. The master spectrum lies constantly above the line of FWHM between 90 and 130 kHz, peaking at 100 kHz.

In EB28 (Fig. 5.66a) the master event and the correlated event represent 39% of the total events. The master signal covers the whole analysed window (102.4 μs) and its spectrum reveals a first peak 105 kHz (and a FWHM between 100 and 140 kHz) and a second smaller peak around 170 kHz.
Figure 5.65: correlated event (left panels), their frequency spectra and stacked spectrum (middle panels) and waveforms and spectrum of the master event (right panels) for a) EB18, b) EB19 (pp = 5 MPa, T = 175°C), c) EB20 and d) EB27 (pp = 16 MPa, T = 175°C) using the bridging technique. For details see text.
Figure 5.66: correlated event (left panels), their frequency spectra and stacked spectrum (middle panels) and waveforms and spectrum of the master event (right panels) for a) EB28 (pp = 16 MPa, T = 175°C), b) EB15 and c) EB36 (pp = 5 MPa, T = 175°C) using the bridging technique. For details see text.
In EB15 (Fig. 5.66b) 3 events have a correlation coefficient above 0.7 with the master events (together representing 29\% of the dataset), whose stacked spectrum shows a powerful peak around the middle of the frequency spectrum plot, with a smaller and narrower peak at higher frequency. The master event, 102.4 μs long, show the same two peaks: a strong peak lies at 120 kHz (having a FWHM of 40 kHz), and a weaker peak at around 160 kHz.

Finally, experiment EB36 (Fig. 5.66c) has a 32\% of highly correlated events (23), whose master event is a relatively short signal (50 μs long) and the master spectrum, though showing multiple peaks, lies constantly above the FWHM line between 50 and 180 kHz.

From the analysis of the dominant frequency and the FWHM of each event (Fig. 5.67 and Fig. 5.68), the double-peak characteristic of these data becomes clearer. In addition, while EB19 (Fig. 5.67b) and EB15 (Fig. 5.68b) have events limited to the immediate instant after the $p_p$ release, all other experiments have several other events recorded some seconds after the venting. In terms of frequency, two clear clusters are visible in all cases: around 100 kHz and around 160 kHz, with the peak at 100 kHz emerging in the majority of the events in all experiments. Temporally, the two peaks start at different times. While in EB27 (Fig. 5.67d) and EB28 (Fig. 5.68a) the cluster at 160 kHz develops later, tending to replace the cluster at 100 kHz towards the end of the 120 s analysed, in all other cases the two clusters appear together since the beginning of the $p_p$ decay, with the higher-frequency cluster tending to disappear at later stage. In addition, experiment EB36 (Fig.5.68c) shows a third cluster, represented by 5 events (out of 72), which lies at 50 kHz and lasts for 10 s (between 20 and 30 s after the release).

Regarding the bandwidth of the dominant frequency peak (shown in terms of FWHM), the majority of the peaks have a FWHM around 20-30 kHz, with few peaks in each experiments having a FWHM up to 100 kHz.
Figure 5.6: temporal evolution of the dominant frequency (top panels) and variation of the FWHM for a) EB18, b) EB19 (pp = 5 MPa, T = 175°C), c) EB20 and d) EB27 (pp = 16 MPa, T = 175°C).
Figure 5.68: temporal evolution of the dominant frequency (top panels) and variation of the FWHM for a) EB28 (pp = 16 MPa, T = 175°C), b) EB15 and c) EB36 (pp = 5 MPa, T = 175°C).
As noted in the previous data, the experiments at high temperature (175°C) with water as pore fluid, show a characteristic double-frequency peak, hereafter called “bimodality”, with dominant frequencies at both 100 and 160 kHz. To better understand this bimodality, and if both peaks are present in the spectrum of each event, the amplitude ratio between the peak around 100 kHz and the peak around 160 kHz is calculated (Fig. 5.69). Bimodality is considered when both peaks reach the -6 dB threshold (equal to the FWHM line), therefore between ratio of 0.5 and 2 (horizontal dashed lines). Note that broadband spectra have ratio close to unity as well, however as observed with the analysis of the FWHM plots, only few events have broadband spectrum under this high temperature conditions.

From Figure 5.69 it is possible to observe that the majority of the ratios fall within the bimodality fields, with the rest of the ratio preferentially lying above a value of 2, meaning a much higher peak at 100 rather than at 160 kHz.
Figure 5.69: amplitude ratio between the 100 kHz peak and 160 kHz peak for a) EB18, b) EB19 ($p_p = 5$ MPa, $T = 175\,^\circ$C), c) EB20, d) EB27, e) EB28 ($p_p = 16$ MPa, $T = 175\,^\circ$C), f) EB15 and g) EB36 ($p_p = 5$ MPa, $T = 175\,^\circ$C), with water as pore fluid.
5.8. RELEASE STAGE IN LOW TEMPERATURE CONDITIONS (NITROGEN)

Table 5.10 summarises basic properties and the results of 4 venting experiments carried out at room temperature using nitrogen gas as pore fluid, while Figure 5.70 shows the cumulative hits superimposed on the pore pressure vs time curve (a) for all experiments. In all cases, the effective pressure is 30 MPa and the axial conduit is 3 mm-wide. Here, Nitrogen gas was added after the sample failure, applying a $p_p$ of 5 MPa for experiments EB32 and EB33, and a $p_p$ of 10 MPa in EB31 and EB35.

As Nitrogen is a compressible fluid, the venting durations are correspondingly longer as revealed by a nearly 100 fold increase ranging from 1.86 to 5.42 s. In experiment EB31 (Fig. 5.70a) the bottom and top $p_p$ decrease to 0 at different time (20 and 30 s after the release respectively), in the other three cases they reach zero simultaneously, between 20 s to 40 s depending on the experiment. In terms of hits, whilst EB32 (Fig. 5.70b) and EB35 (Fig. 5.70d) show a limited growth of hits (less than 10), the other two experiments have a well-defined build-up of AE data.

Experiment EB31 (Fig. 5.70a) displays a rapid increase of hits, distributed over 10 seconds after the release, with few residual hits afterwards. Experiment EB33 (Fig. 5.70c) instead has more than 90% of the hits recorded within a second of the release.

Table 5.10: conditions and results of the 4 venting experiments run at room temperature ($T = 25^\circ C$), using nitrogen as pore fluid.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>$p_p$ (MPa)</th>
<th>Conduit (mm)</th>
<th>Venting duration (s)</th>
<th>Cumul. hits</th>
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</thead>
<tbody>
<tr>
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<td>10</td>
<td>3</td>
<td>2.60</td>
<td>144179</td>
</tr>
<tr>
<td>EB32</td>
<td>5</td>
<td>3</td>
<td>2.70</td>
<td>14</td>
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<tr>
<td>EB33</td>
<td>5</td>
<td>3</td>
<td>1.86</td>
<td>1006</td>
</tr>
<tr>
<td>EB35</td>
<td>10</td>
<td>3</td>
<td>5.42</td>
<td>11</td>
</tr>
</tbody>
</table>
Figure 5.70: cumulative hits (grey dashed line) superimposed on the pore pressure – time plot (solid lines) for the experiment a) EB31, d) EB35 ($p_p = 10$ MPa, $T = 25^\circ$C), b) EB32 and c) EB33 ($p_p = 5$ MPa, $T = 25^\circ$C) with nitrogen gas as pore fluid. At the time of release (0 s), there is an instant decrease of both pore pressures and the onset of AE hits.

Figure 5.71 – 5.74 show the continuous waveforms (Panel A), its spectrogram (Panel B) and the stacked frequency spectrum (Panel C) of the selected signal for the release stages at low temperature with nitrogen as pore fluid. At these conditions, long-duration AE activity (when present) does not die within few milliseconds, but continues over a wide range of time window. Therefore continuous waveforms of different window length are shown for experiments EB31, EB32, EB33 and EB35.

In EB31 (Fig. 5.71), the top $p_p$ takes 18 s to fully decay (with a decelerating trend after 3 s), the bottom $p_p$ reaches 1 MPa within 2 s, before slowing down to decrease to 0 MPa at 8 s. During the first second, the decay curve is interrupted by two increases in $p_p$: AE activity starts simultaneously with the $p_p$ decrease, lasting for the entire duration of the top $p_p$ decay curve, with decreasing amplitude. Two pauses occurred at the time of bottom $p_p$ increases, followed by a quasi-continuous higher-amplitude signal. This signal is characterized by two, well defined spectral peaks of similar
amplitude. The lower frequency peak lies at 60 kHz and has a FWHM of 10 kHz. The higher frequency peak is broader (FWHM = 70 kHz) and centred around 130 kHz.

The behaviour of the \( p_b \) is similar for EB32 (Fig. 5.72) and EB33 (Fig. 5.73) experiments: both bottom and top \( p_b \) follow a quasi-parallel decay trend for most of the duration. Only at the time of release does the bottom \( p_b \) decrease rapidly to 4.5 MPa, slowly decreasing afterwards at a similar decay rate to the top \( p_b \). However, although in experiment EB32 no long-duration AE activity has been recorded over the entire venting stage, in experiment EB33, a significant AE activity appears in the form of three distinct signals lasting between 30 and 60 ms, all within the first 150 ms after the release. The spectrogram reveals similar frequency content for these signals with an area of lower amplitude around 100 kHz. The spectrum of the longer signal shows a higher peak around 50 kHz (FWHM = 40 kHz) and a second smaller peak (just above the -6 dB line) between 105 – 135 kHz.

Finally, experiment EB35 (Fig. 5.74) displays a similar displacement of the two \( p_b \) curves. However, both \( p_b \) tend to slightly increase within the first 500 ms after the release. Over the entire venting stage, only a 10-\( \mu \)s-long signal has been recorded, which occurs almost 400 ms after the time of release. The frequency analysis of this signal reveals a single, narrow (FWHM = 5 kHz) spectral peak at 140 kHz, with two additional smaller peaks (at 65 and 80 kHz) around the -6 dB threshold.
Figure 5.71: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 20 s time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB31 (\(p_p = 10\) MPa, \(T = 25^\circ\)C).

Figure 5.72: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 600 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB32 (\(p_p = 5\) MPa, \(T = 25^\circ\)C).
Figure 5.73: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 300 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB33 ($p_p = 5$ MPa, $T = 25^\circ$C).

Figure 5.74: a) pore pressure (black and green lines) and continuous waveforms (blue line) time series, b) spectrogram over the 600 ms time window and c) stacked frequency spectrum over the selected signal in a) (vertical dashed lines) for EB35 ($p_p = 10$ MPa, $T = 25^\circ$C).
Due to the emergent nature of the generated AE, in experiments EB32 and EB33, only 5 events could be collected. Therefore, further analysis (i.e. waveform similarity, master event and frequency content) cannot be reliably performed for these 2 cases.

However, in EB31 (Fig. 5.75a), 106 events are extracted, showing a high degree of waveform similarity throughout the whole set of window length (more than 90% of correlated events with length up to 30 μs, around 60% at length = 50 μs). As the window length becomes longer, the number of families increases from one family at length = 10 μs to 5 families with length = 50 μs. Experiment EB33 (Fig. 5.75b) has 16 extracted events, having a high degree of similarity up to a length of 20 μs, which falls down to around 40% with a further increase in waveform length. At the maximum length considered, only about 20% of the events are correlated. The correlated events are grouped in 2 – 3 families, showing no correlation with the waveform length.

\[ p_{\text{eff}} = 30 \text{ MPa}, T = 25^\circ \text{C}, \text{ fluid: nitrogen gas} \]

![Graphs showing percentage of correlated events and number of families for EB31 and EB33](image)

*Figure 5.75: percentage of correlated events (top panel) and number of families (bottom panel) recognized for a) EB31 \( (p_f = 10 \text{ MPa}, T = 25^\circ \text{C}) \) and b) EB33 \( (p_f = 5 \text{ MPa}, T = 25^\circ \text{C}) \) using the bridging technique. For details see text.*
In EB31 (Fig. 5.76a), the master event and its correlated events represent 30% of the total event, with the stacked spectrum of this part of the data having a single broad spectral peak. The master event is represented by a 60-μs-long signal, whose spectrum shows 2 spectral peaks above the line of FWHM. The dominant peak is at 140 kHz and has a FWHM of 60 kHz. The second, smaller peak is narrower and lies at 80 kHz.

The master event in EB33 (Fig. 5.76b) correlates 5 signals forming 38% of the total, having a stacked spectrum characterized a single peak at the lower frequencies. The signal of the master events is 102.4 μs long and has too a single spectral peak, lying at 60 kHz and with a FWHM of 20 kHz.

$p_{\text{eff}} = 30 \text{ MPa}, T = 25^\circ\text{C}, \text{fluid: nitrogen gas}$

**Figure 5.76**: correlated event (left panels), their frequency spectra and stacked spectrum (middle panels) and waveforms and spectrum of the master event (right panels) for a) $EB31 (p_p = 10 \text{ MPa}, T = 25^\circ\text{C})$ and b) $EB33 (p_p = 5 \text{ MPa}, T = 25^\circ\text{C})$. 
Finally, the temporal evolution of dominant frequency and the variation of FWHM of EB31 and EB33 are represented in Figure 5.77, which reveals significant differences between the two cases.

In EB31 (Fig. 5.77a) the events are almost equally distributed over 40 seconds after the release, with no a predominant frequency, but instead the spectral peaks are spread over the bandwidth 80 – 180 kHz. The FWHM plot shows a large variability too, with value ranging from few kHz up to 100 kHz. In contrast, EB33 (Fig. 5.77b) has 15 over 16 events occurring within the first second after the release, with the 16° event being recorded after 25 s. While it is not possible to appreciate a trend in the dominant frequency as a function of time, the FWHM plot does reveals a cluster at around 60 kHz, slightly increasing with time. Another characteristic of the events of experiment EB33 is that the FWHM is generally low, between 20 and 30 kHz.

Figure 5.77: temporal evolution of the dominant frequency (top panels) and variation of the FWHM for a) EB31 (p_o = 10 MPa, T = 25°C) and b) EB33 (p_o = 5 MPa, T = 25°C).
6. DISCUSSION

In this chapter, two different stages of the experiments, i.e. the deformation and the venting stages, are compared and discussed with a summary of the whole experimental data (taking into account only samples with the same conduit size of 3mm, numbering 17 experiments) shown in Table 6.1. During deformation, the application of confining and pore pressure (section 6.1) has a first order effect on key parameters including mechanical properties, P-wave elastic-wave velocity and anisotropy, and AE activity. These data are analysed with application to volcanic unrest, and also compared to previous studies from the wider literature. Section 6.2 discusses the role played by the pre-drilled conduit, which is fundamental for the subsequent venting stage.

During venting, the character (both in time and frequency domain) of fluid induced seismicity and the similarity between the events generated by fluid movement (6.3) are recorded. In particular, this allows the effect of pore fluid pressure changes, fluid phases, and decay rate on the AE signals to be analysed. This has proven to be a useful tool in characterising fluids during volcanic unrest, and to infer the volume of volcanic fluids (magma/hydrothermal fluids) involved.

New approaches and experiments such as these are crucial, as many unknowns remain, leading to continued debate on the links between field-measured seismic events and deep seated fluid movements. The analogy between laboratory AE signals and field volcanic earthquakes has contributed to an improved understanding of these deep-seated processes, but has also proven controversial due to the difference in time and space scales. In section 6.4, a qualitative scaling is attempted, in the wake of previous studies (e.g. Burlini et al., 2007; Benson et al., 2014), to address some of these concerns and which confirms some common theories about the generation of fluid-induced volcanic earthquakes.

Finally, the last section (6.5) concludes with a discussion and summary on the different errors occurring during the acquisition and analysis of the laboratory data.
Table 6.1: summary of the experimental data analysed in this study

<table>
<thead>
<tr>
<th>Mechanical properties</th>
<th>DEFORMATION STAGE</th>
<th>Passive events</th>
<th>Location magnitude</th>
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<td></td>
<td>Active events</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>Type</td>
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<td>Sat. experiments (pp = 5 MPa)</td>
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<td>Sat. experiments (pp = 16 MPa)</td>
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<td>At 50% of $\varepsilon_{\text{peak}}$</td>
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<table>
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<th>Fluid Type</th>
<th>T (°C)</th>
<th>$p_p$ decay duration (s)</th>
<th>Long-duration AE Duration</th>
<th>Frequency content</th>
<th>Similarity (30 μs)</th>
<th>Short-duration AE Master event</th>
<th>Frequency Peaks</th>
<th>FWHM</th>
<th>Bimodality</th>
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<td>Water</td>
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<td>0.01 – 0.05</td>
<td>Up to 16 ms</td>
<td>Broadband</td>
<td>73%</td>
<td>Broadband</td>
<td>-</td>
<td>50 – 100 kHz</td>
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<tr>
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<td>175</td>
<td>0.01 – 0.02</td>
<td>Up to 50 ms</td>
<td>Bimodal, 60 and 100 kHz</td>
<td>66%</td>
<td>Bimodal</td>
<td>100 and 160 kHz</td>
<td>20 – 30 kHz</td>
<td>yes</td>
</tr>
<tr>
<td>Nitrogen</td>
<td>25</td>
<td>2 – 6</td>
<td>Up to 14 s</td>
<td>Bimodal</td>
<td>70%</td>
<td>Narrow peaks</td>
<td>-</td>
<td>10 – 50 kHz</td>
<td>no</td>
</tr>
</tbody>
</table>
6.1. THE NATURE OF ROCK-FLUID COUPLING DURING THE DEFORMATION STAGE

6.1.1. ON THE MECHANICAL PROPERTIES

In the previous chapter, the results of the deformation stage are shown with respect to their pore pressure conditions, i.e. dry, 5 MPa and 16 MPa fluid pressure (water). To clarify, a $p_p$ of 16 MPa corresponds to water saturation at 1.6 km, in which $p_c = 46$ MPa and the effective pressure ($p_{eff}$) of 30 MPa. A $p_p = 5$ MPa, combined with a $p_c = 35$ MPa, does not correspond to water saturation at 1.2 km, but was used to keep $p_{eff} = 30$ MPa.

Detailed statistics (peak differential stress, $\sigma_{diff}$, and Young’s modulus, $E$) for these conditions are reported in Tables 5.2, 5.4 and 5.6, while Figures 5.1, 5.16 and 5.32 show the stress-strain data.

The similar peak $\sigma_{diff}$ and $E$ (Table 6.1) between the different samples with same pore pressure conditions provides evidence that the sample material is homogeneous on the scale of the laboratory investigation, and therefore the data is repeatable between samples.

Young’s modulus $E$ does not show any clear effect due to pore pressure. This is consistent with previous studies: this parameter is controlled by the microstructure of the rock (e.g. Ayling et al., 1995; Heap et al., 2009), which is not affected by the presence of pressurized fluids. In fact, this result confirms the isotropic distribution of cracks for this type of rock, as reported by Vinciguerra et al. (2005) and Stachits et al. (2006). Therefore, rock samples drilled from this basalt have similar $E$, regardless the pore pressure conditions.

The behavior of $\sigma_{diff}$ is similar to $E$, showing no significant changes as a response of the increasing $p_p$. However, all experiments were performed at the same effective pressure ($p_{eff} = 30$ MPa), meaning that an increase in $p_p$ is followed by an increase in $p_c$. With the same effective $p_c$, $\sigma_{diff}$ has the same values (respectively 452 MPa in dry conditions, 451 MPa with $p_p = 5$ MPa and 444 MPa with $p_p = 16$ MPa; Table 6.1) within experimental error (or, sample variability). Taken together, therefore, the rock-fluid coupling as measured by mechanical data is described by the standard analysis of the effective medium theory, where the mechanical properties are controlled by the pore-pore interactions, hence the porosity, (Jaeger et al., 2007) and by the effective pressure ($p_c - p_p$), not by the pore pressure itself.
6.1.2. ON THE ELASTIC WAVE PROPERTIES

To analyse the P-wave elastic velocity in both time and space, three experiments having the most frequent velocity surveys and spanning dry and water saturated conditions (5 and 16 MPa) were chosen. These are EB32, EB16 and EB34. To visualise the interaction between the rock mechanical conditions, pore fluid, and measured P-wave data, a series of stereonets were constructed, initially just before the onset of triaxial deformation (Fig. 6.1). Under these hydrostatic starting conditions of $\sigma_1 = \sigma_2 = \sigma_3$ and same $\sigma_{eff}$ (i.e. 30 MPa), compressional velocity increases from an average of 4100 m/s in the dry case (a) to 4900 m/s (b), and 5100 m/s (c) as the pore pressure is increased through 5 MPa to 16 MPa respectively. To maintain the same $\sigma_{eff}$ in all experiments, the confining pressure ($p_c$), is also increased by the same value. It is likely that the higher P-wave velocity is not a result of crack closure, but instead due to the presence of pressurized fluids in the rock cavities (which has a pre-existing microcrack network and a 2% porosity), by promoting elastic wave propagation (Vinciguerra et al., 2005). Another consequence of saturating the pore space is the increase of both bulk modulus and density, with the increase of the former more significant than the latter, thus explaining the increase of the compressional wave under saturated conditions (Jaeger, 2007).

It is also seen that when moving from dry to highly saturated conditions, the initial P-wave anisotropy slightly decreases (Fig. 6.2). Once again, the presence of pore pressure is invoked to explain these data. As water has a higher sound of speed than air (1500 m/s versus 343 m/s, which increases with increasing pressure), its presence not only increases the P-wave velocity (discussed above) but contributes to decreasing the difference (in terms of compressional velocity) between the rock matrix and the pore space, hence decreasing the P-wave anisotropy.
The variation in colour from a) blue to b) yellow/green and c) red represents the increase in P-wave velocity as the pore pressure increase.

The variation of the P-wave anisotropy as a function of time is illustrated in Figure 6.2. To combine different experiments (with different time and strain scale), the strain percentage (ε %) has been normalised with respect to strain at the peak σ_{diff} (x-axis). This allows a general comparison between all datasets reported in the result to be made.

All three conditions (dry, 5 MPa and 16 MPa) start from a similar value of anisotropy (between 1.05 and 1.10). In dry conditions, the P-wave anisotropy has a concave-towards-down shape with increasing strain, followed by an inflexion at approximately 50% of peak-strain, and then a concave-towards-up shape, reaching a level of anisotropy of 1.35 just before peak-strain, which jumps to a maximum of 1.5 immediately before sample failure (blue circles and line). In comparison, the water-saturated cases feature a much more linear trend throughout the deformation stage and beyond peak-strain, and a much lower anisotropy of 1.23 just after the peak stress (sample failure). Similar conclusions were reached by Benson et al. (2010), who noticed that the anisotropy approaching failure in dry samples was higher than that in water saturated samples. This is likely to be a direct consequence of the pressurized fluid, which fill the newly formed cracks and enhance the propagation of seismic wave (Vinciguerra et al., 2005), as well as to homogenize the sample by reducing the difference between vertical and horizontal P-wave velocities. In addition, as noted above, as the pore pressure increases, the anisotropy decreases, which is explained in terms of higher sound of speed of water and therefore reduced difference between the compression wave velocity of the rock matrix and the sound of speed of the pore fluid.
6.1.3. ON THE AE ACTIVITY

The presence of pressurized fluids inside the rock sample has significant impact on the recorded AEs. Figure 6.3 shows hit rate data versus strain for three experiments in a semi-log plot (so exponential patterns appear linear), in this case selected for their best signal-to-noise ratio. These represent a dry sample, and two water saturated samples: EB23, EB16 (5 MPa) and EB25 (16 MPa) respectively. Three clear effects emerge from these data. Firstly, the onset of microseismic activity is increasingly delayed as the pore pressure increases. In dry conditions (blue circles and line) the hit rate builds up early in the deformation stage, at approximately 25% of the strain at peak, while it starts at 50% with pore pressure at 5 MPa (red circles and line) and at 65% with pore pressure at 16 MPa (yellow circles and line). Secondly, the maximum hit rate at the time of peak stress is lowered. In absence of water, the hit rate reached a value of 100 hits/5 s when the peak stress is reached, while it achieves 40 hits/5 s in both water-saturated conditions at the moment of sample failure. However after the peak stress, during the strain-softening regime preceding sample’s failure, EB23
(dry) and EB25 ($p_p = 16$ MPa) reach a similar hit rate (i.e. 100 hits/5 s), with EB16 ($p_p = 5$ MPa) reaching 50 hits/5 s. A velocity survey was performed in EB16 during the strain softening regime, so no AE activity was recorded, likely explaining the lower value in hit rate during such regime.

Finally, the general behaviour of the AE build up is modified by the magnitude of the pore pressure. In dry conditions the hits rate grows exponentially, with a change in the pattern at the time of the peak stress. Conversely, AE recorded in the water saturated samples, the exponential growth is reduced in time and both experiments manifest a supra-exponential increase in the later stage of the deformation, but well before the peak stress, particularly jumping from 10 to 40 hits (300% increase) at 99% strain with pore pressure at 16 MPa.

![Figure 6.3: Hit rate versus strain percentage relative to the strain at peak stress for the dry (blue, EB23, $p_c = 30$ MPa, $p_p = 0$) and water saturated conditions (red for EB16, $p_c = 35$ MPa, $p_p = 5$MPa), yellow for EB25, $p_c = 46$ MPa, $p_p = 16$ MPa). As the pore pressure increase, the onset of AE activity is delayed, the hits rate at the stress peak is lowered and the build-up behaviour is changed.](image)

To better understand what causes such differences in the hit rate pattern, we analysed the pore volume change curve to measure and analyse the onset of dilatancy ($D'$) (Fig. 6.4). $D'$ marks the boundary between compaction and fracturing regime and has been associated with the onset of AE activity (Scholz, 1968b). The compaction regime is characterized by the closure of pre-existing
cracks and pores and does not generate much AE (e.g. Aker et al., 2014). When all pores and crack are closed, the sample starts to increase volume (dilatancy) due to fracturing. Because dry experiments do not have pore pressure, D’ cannot be determined via the methods employed in this research. However, evidence from the two datasets (5 and 16 MPa pore pressure) suggests that as the pore pressure increases, D’ (marked by the local minimum in the pore volume change curve) is delayed, increasing from ~35% strain at 5 MPa to 50% strain at 16 MPa. This means that the duration of the compaction stage depends on the pore pressure. In fact, as the pore pressure increases, a higher stress is required to close the pressurized pores/cracks (Jaeger et al., 2007), so delaying D’ and fracturing.

Once D’ occurs, the aseismic crack closure ends and AE build-up may be expected. However, this is not observed, with AE activity starting 15% strain later than D’ in both cases. This, coupled with the low amplitude of AE events early in this process, limits the AE hit rate that can be detected around D’, unlike a quartz-rich granite (Meredith et al., 1990).

As the deformation ultimately ends as a shear fault (or a set of conjugate faults) cutting the sample at approximately 30°, one can expect that the hit rate around failure is similar for all cases. This is indeed observed, particularly during the strain softening regime when the AE imaging of the fault becomes clear (see Fig. 5.12, 5.28, and 5.45). Hence, while the dry sample experiments exhibit an early and constant exponential build-up towards 100 hits at the time of failure, the saturated samples need to catch up with a faster than exponential growth to reach 100 hits by the time of sample failure, since the pore pressure is acting to delay this process.
Figure 6.4: pore volume change curve for the water saturated experiments with \( p_p = 5 \) MPa (red symbols, EB16, \( p_c = 35 \) MPa) and \( p_p = 16 \) MPa (yellow symbols, EB25, \( p_c = 46 \) MPa).

Crucially, similar patterns (exponential and supra-exponential increase) of hit rate (Fig. 6.3) can be observed in nature, such as the seismic activity preceding volcanic eruptions at Redoubt and Pinatubo volcanoes (Fig. 6.5). Here, the precursory seismic activity is similar to the laboratory hit rate recorded during the water saturated (\( p_p = 16 \) MPa) and dry experiments respectively. This comparison leads to the hypothesis that pressurised fluids play a role in affecting the seismic rate before major eruptions, and therefore might better inform forecasting methods for warning people. In fact, the eruption on Pinatubo was preceded by a full 7-day sequence of VT events, later followed by tremor and LP signals, before the major eruption on June 15, 1991 (Newhall & Punongbayan, 1996). The presence of VT activity indicates a higher proportion of shear-fractures. This may be associated to the dry experiment because of i) the predominance of shear fractures with absence of fluids actively involved in the generation of seismicity; ii) the long-lasting and exponential precursory seismicity growth.

Conversely, the 1989 eruption on Redoubt was preceded by a 23-hour-long sequence of LP events (Chouet et al., 1994), which is generally more likely associated with fluid movement and/or pressurisation. Again, this case may be associated to the water-saturated experiment because i) fluids are actively involved in generating precursory seismic activity; ii) the exponential increase is reduced in time and the supra-exponential is well developed.
These two cases constitute field examples of the same scenario of the dry (where the fracturing does not involve fluids) and water saturated (where the fluids contributes to the fracturing as demonstrated by their influence on both mechanical, acoustic and microseismic properties) laboratory experiments. This naturally leads to the idea that volcanic eruptions with poorly-pressurized fluids are more easily forecasted by the monitoring of the early (VT) precursory seismic activity, while earthquakes generated in volcanoes with highly-pressurized fluids may only be detected much closer to the eventual eruption. This would clearly have wide ranging significance for hazard forecasting, but as yet there is no clear quantitative link.

![Sketch of the precursory seismic rate at different volcanoes showing distinctive and not unique behaviour, making the eruption forecasting extremely unreliable.](image)

*Figure 6.5: Sketch of the precursory seismic rate at different volcanoes showing distinctive and not unique behaviour, making the eruption forecasting extremely unreliable.*

To further analyse and discuss how the AE data can be linked to the fluid saturation, the change in the seismic $b$-value with time is analysed both in the field (e.g. Roberts et al., 2015) and also in the laboratory (e.g. Meredith et al., 1990) (Fig. 6.6). In this figure two minima are present, one around the onset of the strain softening, the second before the stress drop.

Previous studies (Meredith et al., 1990; Lockner et al., 1991; Sammonds et al., 1992) show that under dry or constant pore pressure conditions, a single minimum is expected, thought to be connected to the fault nucleation process (Meredith et al., 1990) occurring during strain softening, where larger fractures generate larger amplitude AE than in the pseudo-elastic part. The presence of two minima is linked to constant pore volume conditions, where short-term pore-pressure
variation created a double minimum in the $b$-value curve, itself as a result of a temporary reduction of stress intensity due to the decreasing differential stress and pore pressure (Sammonds et al., 1992). This relaxation, after the first minimum, then causes a recovery in the $b$-value, finally followed by the second minimum, and sample failure. However, in these three experiments, all have a complex system of at least two fractures (Fig. 6.7), which are imaged by the minima in the temporal evolution of the $b$-value.

A detailed review on the spatio-temporal variations of the $b$-value has been conducted by El-Isa & Eaton (2014), showing that $b$-value generally increases for a certain amount of time before the occurrence of a tectonic earthquake, which is characterized by a drop in $b$-value. This is somewhat similar to the result of these experiments: the sample failure, which is caused by the formation of a major damage zone, is generally accompanied by a global minimum, due to the occurrences of more frequent higher amplitude events. The presence of multiple smaller (local) minima preceding the failure are likely connected to the occurrence of foreshocks, after which the $b$-value tend to increases once again. While the presence of multiple minima indicate the onset of strain-softening regime, which eventually ends up in the formation of the main damage zone, the time preceding such event remains unknown as well as a direct connection of the duration of strain-softening regime between laboratory and field.

Roberts et al. (2015) noted that volcanic $b$-value tend to change prior or during eruptive phases. In particular they observe high $b$-value prior to the 2011 eruption on El Hierro (Canary Islands) and during flank eruption at Mt Etna in 2001-2003. This imply a low $b$-value at the time of failure for both cases. In fact in the experiments shown in this study, high level of $b$-value characterized both the pre-failure and post-failure regime, with minima in the $b$-value only around the time of failure, which may not be the time of eruption (Kilburn, 2003).
Figure 6.6: temporal evolution of b-value, plotted against strain percentage relative to strain at peak stress, for left) the dry (EB23, $p_r = 30$ MPa, $p_p = 0$) and water saturated samples with center) $p_p = 5$ MPa (EB16, $p_r = 35$ MPa) and right) $p_p = 16$ MPa (EB25, $p_r = 46$ MPa). Dotted lines mark the onset of the strain softening, dashed lines indicate the stress drop.

Figure 6.7: failed samples in a) dry (EB23, $p_r = 30$ MPa, $p_p = 0$) and water saturated samples with b) $p_p = 5$ MPa (EB16, $p_r = 35$ MPa) and c) $p_p = 16$ MPa (EB25, $p_r = 46$ MPa).
As highlighted by the *b*-value behaviour, during strain softening events with larger amplitudes characterize the early part of the deformation experiments. Even though the *b*-value is calculated by using the highest amplitudes of the extracted events, the pattern of the average $M_L$ (Fig. 6.8), which is calculated for the same dataset used to retrieve the *b*-value, reflects the *b*-value behaviour. In fact, one can observe two peaks in event magnitude in all experiments, matching the double minimum of *b*-value.

These observations confirm the goodness of the *b*-value achieved through simple signal amplitudes. Although its absolute value cannot be evaluated, the qualitative trend, and particularly points of minima, can be easily found without calculating the seismic magnitude of each event. In other words, a relative magnitude is equally acceptable using trend analysis. Obtaining an accurate event magnitude require certain numbers of arrival times with clear arrivals and a precise velocity model, thus reducing the number of available data points to calculate the *b*-value. In addition, those events with large location errors or located outside the sample (mostly due to automatic miss-picking and uncertainty in the velocity model) are discarded. For instance, in EB23, 15689 events were extracted (meaning 15689 amplitude over which calculate the *b*-value), but only 3170 signals are used to obtain the magnitude, so only a fifth of the entire dataset. Therefore, using the signal amplitude (voltages in this study, but displacement or ground velocity/acceleration for field events) instead of the magnitude to analyse the Richter-Gutenberg distribution of earthquakes not only give similar overall results, but is achieved with less computational time due to less work spent in getting the magnitude, and more details, due to higher number of available data points.

In Fig. 6.8, a clear increase of $M_L$ at the time of failure is evident where the dry experiment manifests an increase of 1.5 magnitude from the level at 95% of strain and both saturated experiments an increase of about 0.7 magnitude. While for the saturated experiment at 5 MPa no high-resolution mechanical data at the time of failure is available, for the other two experiments these are present, as shown in plots in the previous chapter (e.g. Fig. 5.12 and 5.45). These reveal that an instantaneous stress drop of almost 500 MPa occurs in the dry experiment; however, in the saturated experiment ($p_F = 16 \text{ MPa}$) this instantaneous stress drop is about 350 MPa, explaining the difference in the average $M_L$ at the time of failure.
Figure 6.8: temporal evolution of the average location magnitude, plotted against strain percentage relative to strain at peak stress, for left) the dry (EB23, $p_c = 30$ MPa, $p_p = 0$) and water saturated samples with center) $p_p = 5$ MPa (EB16, $p_c = 35$ MPa) and right) $p_p = 16$ MPa (EB25, $p_c = 46$ MPa).

The strain softening stage is characterized by the nucleation of the fault, with the foci of the AE events clustering around the eventual shear zone (Lockner et al., 1991). Sets of conjugate faults were observed both in dry (Thompson et al., 2006) and in wet saturated conditions (Benson et al., 2010). Figure 6.7 shows an example set of post-test photographs of the failed samples at the three different conditions. All samples exhibit a major damage/shear zone which spans from the middle/upper part to lower end of the sample, and is accompanied by a complex conjugate fracture, which stops at major faults, indicating a later nucleation and propagation other than the major structure.

The different stages of fault nucleation is analysed by studying the clustering of the location, represented by the temporal variation of the average inter-event distance (Fig. 6.9). All data exhibit two minima, with the latter likely corresponding to the nucleation of the major fault zone. The nucleation switches from a minor to major fault which appears during the strain softening stage of
deformation (middle and right panels); the timing of the two minima corresponds to the minima in $b$-value and the peaks in event magnitude.

Figure 6.9: Temporal evolution of the average inter-event distance, plotted against strain percentage relative to strain at peak stress, for left) the dry ($EB23$, $p_c = 30$ MPa, $p_p = 0$) and water saturated samples with center) $p_p = 5$ MPa ($EB16$, $p_c = 35$ MPa) and right) $p_p = 16$ MPa ($EB25$, $p_c = 46$ MPa).

6.2. EFFECT OF THE CONDUIT DURING THE DEFORMATION STAGE

6.2.1. ON THE MECHANICAL PROPERTIES

The second stage of the experiments requires access to the damage zone in order to rapidly discharge pore pressure and stimulate rock-fluid interaction. For this reason, a 3.175 mm (1/8”) wide axial conduit, the smallest core drill that could be acquired, was drilled through the whole length in most of the specimens used in the study. Previous work (e.g. Benson et al., 2008) performed experiments with samples characterized by a similar pre-drilled axial conduit, noticing that its presence does not affect the mechanics of deformation and failure. To confirm, two experiments ($EB26$ and $EB32$), performed with respectively solid and pre-drilled samples, are compared. Both experiments are carried in dry conditions, with a $p_c$ of 30 MPa.
Although a small difference in $E$ is observed (the presence of a conduit changes the microstructure, as it represents a discontinuity), their stress/strain curves are very similar, and the peak stress the same within 2% (57 and 52 GPa, Fig. 6.10). This is within the regular experimental error expected for sample variability amongst samples, and thus for the purposes presented, the conduit may be safely neglected in terms of mechanical strength. This is reinforced by a simple calculation of sample area: samples with a conduit have 99.4% of the area of samples with no conduit. Most of the triaxial deformation experiments – whether conduit of solid – produce a fault inclined at 30° to the maximum stress axis, so the conduit, which is parallel to it, does not interfere with the formation of the fault (Fig. 5.12, & 5.45).

![Figure 6.10: Differential Stress versus Strain percentage relative to strain at peak for the solid (blue circles and line, EB26, $p_c = 30, p_p = 0$) and drilled sample (red circles and line, EB32, $p_c = 30, p_p = 0$). Both samples have a similar mechanical behaviour and peak stress.](image)

### 6.2.2. ON THE ELASTIC WAVE PROPERTIES

P-wave velocity of the intact samples (Fig. 6.11) shows that both samples have similar starting velocities. Intuitively, lower velocity may have been expected with samples containing a conduit, with a minimum in the axial region. Instead, solid samples have an average P-wave velocity of 3800 m/s (Fig. 6.11a), while the drilled sample have a velocity of 4100 m/s (Fig. 6.11b). This
difference is likely due to small differences in mineral composition, crystal size and microcrack network rather than the presence of the conduit. To evaluate this assumption, the wavelength ($\lambda$) produced by the PZT sensors while performing velocity survey needs to be calculated as the minimum thickness of a layer to be detected (i.e. seismic resolution) is a quarter of the wavelength (Chopra et al., 2006). Considering frequency ($f$) between 50 kHz and 1 MHz and the P-wave velocities ($v$) of the samples ranging from 3700 – 6000 m/s, with the equation:

$$\lambda = \frac{v}{f}$$

Wavelength $\lambda$ ranges from 3.7 to 120 mm, which could make the conduit detectable ($\lambda$ being smaller than 4 times the conduit size). However, during velocities surveys the receivers record the incoming signals with a dominant frequency of 300 kHz, which make $\lambda > 12$ mm (4*3 mm) for every possible P-wave velocity. In this way the conduit remain seismically undetectable.

Figure 6.11: stereonets of the survey of intact rock of a) a solid (EB26, $p_c = 30$, $p_p = 0$) and b) drilled sample (EB32, $p_c = 30$, $p_p = 0$). Note that both stereonets share same hue of blue, meaning that they have similar P-wave velocities.

The negligible effect of the pre-drilled conduit on the acoustic properties of the sample may be further evaluated by inspecting the temporal variation of the P-wave anisotropy (Fig. 6.12). Although the drilled sample has an initial higher anisotropy (red circles and line), visible also in Fig. 6.11b, both curves in Fig. 6.12 are characterized by having an initial concave-down shape, followed by concave-up, with a point of inflection at 50% of the strain at peak stress. In addition, they both reach a value of anisotropy of approximately 1.5 when the sample approaches peak
stress. Therefore, on balance, the effect of the conduit is assumed negligible in terms of both mechanics, and energy seismic transmission.

![Figure 6.12](image)

**Figure 6.12:** P-wave anisotropy versus Strain percentage relative to the strain at peak stress for the solid (blue, EB26, \(p_c = 30, p_p = 0\)) and drilled (red, EB32, \(p_c = 30, p_p = 0\)) sample. Both curves have an inflection point at 50% and reach similar value of anisotropy at peak stress.

6.2.3. **ON THE AE ACTIVITY**

The recorded AE activity is also not affected by the presence of the pre-drilled conduit. Figure 6.13 compares the hit rate for the solid (blue circles and line) and drilled sample (red circles and line) for the experiment with the best SNR, EB21 and EB31 respectively, both run at dry conditions, with \(p_c = 30\)MPa. As mentioned for the mechanical and acoustic properties, both samples share a similar hit rate build-up, with onset of the AE activity earlier in the deformation stage (about 25% strain) and a constant exponential increase (marked by the dashed lines in the semilog plot) of the activity up to the peak stress with a hit rate of between 100 and 120 before the peak. After failure both experiments reach a maximum value of 160 hits.
Figure 6.13: Hits rate (recorded every 5 seconds) versus Strain percentage relative to the strain at peak stress for the solid (blue, EB21, $p_c = 30$, $p_p = 0$) and drilled (red, EB31, $p_c = 30$, $p_p = 0$) sample plotted as a semilog plot. No significant difference in the build-up of the hits rate are visible. The dashed lines represent the exponential growth (which appears linear in a semilog plot).

The negligible effect of the pre-drilled conduit is also clear when analysing the other two key AE parameters: seismic $b$-value and magnitude. The single minimum presents in both $b$-value evolution (Fig. 6.14) is reflected by the single peak in the normalized average $M_L$ (Fig. 6.15), which shows a similar increase at the time of failure for both cases. Differences between the two cases, however, finally begin to appear when inspecting the average inter-event distance (Fig. 6.16) where the presence of the conduit possible results in a coarser trend and slightly lower value of inter-event distance. A possible explanation for this is the absence of a fully developed $30^\circ$ fault in EB31 and the presence of a much wider area of fracturing, which contribute to increase the distance between pairs of events.
Figure 6.14: temporal evolution of b-value, plotted against strain percentage relative to strain at peak stress, for left) the solid (EB21, $p_c = 30$, $p_p = 0$) and right) pre-drilled samples (EB31, $p_c = 30$, $p_p = 0$). Note the clear decrease during strain softening regime.

Figure 6.15: temporal evolution of the normalized average $M_L$, plotted against strain percentage relative to strain at peak stress, for left) the solid (EB21, $p_c = 30$, $p_p = 0$) and right) pre-drilled samples (EB31, $p_c = 30$, $p_p = 0$). Both curves present an oscillatory/flat part before peak stress, followed by a sharp increase reaching a maximum at the time of failure. Note the similar increase in $M_L$ during the strain softening regime.
Figure 6.16: temporal evolution of the average inter-event distance, plotted against strain percentage relative to strain at peak stress, for left) the solid (EB21, \( p_c = 30, p_p = 0 \)) and right) pre-drilled samples (EB31, \( p_c = 30, p_p = 0 \)). Clear differences are visible during the strain softening regime: while in the solid sample a single minimum is present preceding the time of failure (left panel), a broad minimum at the peak stress characterized the pre-drilled sample (right panel), followed by a general increase in the distance. This can be explained by the presence of a wider area of fracturing which contribute to increase the distance between pair of events.

6.3. ROCK-FLUID COUPLING DURING PORE PRESSURE RELEASE

6.3.1. REQUIREMENTS FOR THE ONSET OF THE LONG-DURATION AE ACTIVITY

The discussion of long-duration AE activity (sections 5.6, 5.7 and 5.8) illustrate how seismicity may be generated when the pore pressure is rapidly released under different pressure/temperature/fluid phase conditions. However, four conditions must occur to generate this AE activity, which are i) the presence of a natural fracture zone, ii) the presence of the axial conduit, iii) a fast pore pressure release and iv) the onset of a pore pressure difference.

Figure 6.17a shows the pore pressure release through the axial conduit of a sample, before the deformation (i.e. generation of fracture damage zone). Using a smooth conduit (no damage zone), and a fast pore pressure release of 5 MPa/20 ms (same rate as in EB16, Fig. 6.17d, black line), no AE activity is generated (Fig. 6.17a, blue line). As postulated by Julian (1994), and tested through event localization in decompression experiments by Benson et al. (2008), a complex, tortuous
pathway is needed for generating tremor-like, long-duration AE activity, whose focus lies at the constriction of the irregular fracture zone. In addition, the absence of a natural damage zone implies the absence of broken and comminuted rock, observed both in the field (Tuffen & Dingwell, 2005) and experimentally (Benson et al., 2008), which changes the rheology of the fluid. In fact, a turbulent viscous fluid is required to generate self-excited oscillations (Julian, 1994).

Although the conduit itself does not produce microseismic events, its presence is fundamental to access the damage zone, allowing the fluids to move easily from the fracture towards the end of the sample, as well as inducing the fluid turbulence. Figure 6.17b shows the venting stage in a deformed sample, where the conduit was not pre-drilled. While the bottom pore pressure is fully released in 20 ms, the top pressure decreased at a much slower rate (still 16 MPa in the 90 ms after the release), creating a long-duration pore pressure difference of 16 MPa. However, as for the previous case, no AE activity is recorded. Without the conduit, the speed of the moving fluids depends entirely on the permeability of the sample. Even though neither the permeability nor the flow speed are measured, one can observe the difference in the top pore pressure decay, which is directly connected to the fluid flow speed, between two deformed samples, one without the conduit (Fig. 6.17b, black line) and one with it (6.17d, black line). Here again the model of Julian (1994) is invoked to explain the importance of the flow speed. Self-excited oscillations, responsible for tremor-like activity, can only occur when the flow speed exceed a critical threshold. Below this threshold, if the flow is disturbed, discrete LP-like events occur.

When a gas phase is used to build the pore pressure, the higher compressibility of the gas compared to the liquid make the pore pressure decay lower by a factor of 3. However, as shown in the previous chapter for experiments EB31 and EB33 (e.g. Fig. 5.71 and 5.73 respectively), long-duration AE is generated in such conditions. Although the venting of all experiments with gas are performed on pre-drilled, deformed samples, and show similar pressure decay rates (meaning similar flow speed), only EB31 and EB33 exhibit a pore pressure difference lasting for several seconds, while in EB32 (here shown in Fig. 6.17c) and EB35, both top and bottom pore pressures decay at the same rate, which does not allowing sufficient pore pressure difference to be generated. As a consequence, no AE activity has been detected above the background noise.

According to the Darcy’s law for fluid flow, the total discharge \(Q_{flow}\), is directly proportional to the pore pressure difference (or drop, \(\Delta p_p\)) via the equation:
\[ Q_{\text{flow}} = -\frac{A \cdot k \cdot (\Delta p_p)}{\mu \cdot L} \quad \text{(Eq. 6.1)} \]

where \( A \) is the cross-sectional area to flow, \( k \) is intrinsic permeability, \( L \) is the actual length where the pressure drop takes place and \( \mu \) is the viscosity of the fluid. If Eq. 6.1 is divided by the cross-sectional area, the flux (\( q \)) can be calculated as:

\[ q = -\frac{k}{\mu} \cdot \nabla p \quad \text{(Eq. 6.2)} \]

where \( \nabla p \) is the pressure gradient vector. Dividing the flux for the porosity (\( n \)) of the medium, then the fluid flow speed (\( v \)) can be calculated as:

\[ v = \frac{q}{n} \quad \text{(Eq. 6.3)} \]

Neither the intrinsic permeability, nor the fluid viscosity nor the porosity of medium (which is different from the porosity measured from the intact sample in section 5.1), are known and the cross-sectional area can be estimated, so it is not possible to calculate the flow speed. In addition, as this is likely to be non-laminar flow, such an approach would be an estimate at best. However, as the pressure drop is directly proportional to the flow speed, the higher the pore pressure difference, the farther the fluid moving through the sample, therefore higher chances to overcome the critical threshold postulated by Julian (1994). Qualitatively, therefore, the higher pressure venting is likely to result in a higher initial flow rate during the venting, which is sufficient for the purposes of this study to induce the AE, even if the precise flow rate cannot be measured in the current setup.
Figure 6.17: Streaming AE data (blue lines) superimposed on the pore pressure decay curves (black and green lines) for a) a sample with a pre-drilled conduit, but no damage zone, b) a solid sample (no conduit) but with a damage zone, c) a combined deformed sample (with a damage zone) and conduit using gas only and d) a pre-drilled, deformed, with liquid phase sample.

6.3.2. PARAMETERS AFFECTING THE LONG-DURATION AE ACTIVITY

Once the AE activity is triggered, its character in the time- and frequency-domain depends on different controls such as initial pore pressure, number and type of fluid phases (which are controlled) as well as a number of uncontrolled variables (e.g. dynamic pressure drop and pressure release rate at the top of the sample). To best evaluate the effects of these variables in the time-domain, the envelope of the signal has been generated (Fig. 6.18) to further analyse its properties with respect to the controlled and uncontrolled variables.
Table 6.2 summarizes the range of variables (both controlled and uncontrolled) that affects the long-duration AE activity and the features of the signal envelope. To this list of experiments, the initial build-up of gas pressure in experiment EB31 (called EB31_f) is added as it was also found to generate AE activity. In this case, AE activity is generated when gas was still supplied to the sample, while all other cases AE activity commenced after the desired starting p_p. The relationship between these parameters are shown in Fig. 6.19, in an attempt at a simple calibration exercise to show that the pore pressure can indeed be used as a proxy to the AE signal, itself acting as a laboratory analogue to field seismic activity. In particular the correlation between the AE signal envelope (Fig 6.19b) and the p_p drop or p_p decay rate (Fig. 6.19a and c) is qualitatively excellent: the relationship (Table 6.2, last two columns) is calculated by normalizing the variables and correlating them over a window length of 100 ms when water is present. For the two cases of gas-only phase (i.e. EB31 and EB31_f) window lengths of 20 s and 1.1 s are used respectively.
Fig. 6.19: a) absolute top $p_p$ decay rate, b) signal’s envelope and c) absolute $p_p$ drop. 1) $p_p$ decay rate peak; 2) $p_p$ decay FWHM; 3) lag between $p_p$ decay and AE activity; 4) envelope’s peak length; 5) envelope peak; 6) envelope FWHM; 7) lag between $p_p$ decay and AE activity; 8) $p_p$ drop peak; 9) $p_p$ drop FWHM.
The next step is to determine what (independent) variables affect the AE activity (i.e. a dependent variable) and to remove those variables which are correlated.

From the data in Table 6.2 a linear correlation may be derived between: i) the maximum top $p_p$ decay rate and the pressure drop peak ($R^2$ coefficient = 0.89), ii) the FWHM of the rate decay curve and the FWHM of the pressure drop ($R^2 = 0.99$), iii) the correlation coefficients between the AE signal envelope – $p_p$ drop and AE signal envelope - top $p_p$ decay rate, with the $R^2$ coefficient between the two correlation coefficients being 0.70.

Combining this information with the lag times between these variables and the AE activity (where the pressure drop always starts before the top $p_p$ decay curve), results is in excellent agreement with Darcy’s law, suggesting that pressure drop (venting) directly controls the fluid flow (and hence, the pressure decay curve), which, surprisingly, is therefore not fast enough (less than 1 m/s overall) to act in the turbulent regime in terms of bulk flow, even though the generation of AE suggests that locally and temporally, pockets of turbulence are generating rock-fluid coupling resulting in AE.
Table 6.2: variables and features of the long-duration AE activity.

<table>
<thead>
<tr>
<th>Test #</th>
<th>Initial p&lt;sub&gt;p&lt;/sub&gt; (MPa)</th>
<th>Fluid Phases</th>
<th>p&lt;sub&gt;p&lt;/sub&gt; Drop Peak (MPa)</th>
<th>Max. Top p&lt;sub&gt;p&lt;/sub&gt; decay rate (MPa/s)</th>
<th>p&lt;sub&gt;p&lt;/sub&gt; drop’s FWHM (ms)</th>
<th>p&lt;sub&gt;p&lt;/sub&gt; decay’s FWHM (ms)</th>
<th>Lag between p&lt;sub&gt;p&lt;/sub&gt; drop and AE (ms)</th>
<th>Lag between p&lt;sub&gt;p&lt;/sub&gt; decay and AE (ms)</th>
<th>Signal’s envelope peak (V)</th>
<th>Envelope’s peak length (ms)</th>
<th>Envelope’s FWHM (ms)</th>
<th>Envelope – p&lt;sub&gt;p&lt;/sub&gt; drop correlation coefficient</th>
<th>Envelope – p&lt;sub&gt;p&lt;/sub&gt; decay rate correlation coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>EB16</td>
<td>5 water</td>
<td>4.48</td>
<td>0.11</td>
<td>11.00</td>
<td>6.00</td>
<td>8.65</td>
<td>7.40</td>
<td>5.33E-04</td>
<td>22.58</td>
<td>8.49</td>
<td>0.84</td>
<td>0.88</td>
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</tr>
<tr>
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<td>7.60</td>
<td>5.40</td>
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<td>3.93</td>
<td>3.21E-04</td>
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<td>7.27</td>
<td>0.93</td>
<td>0.93</td>
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</tr>
<tr>
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<td>0.14</td>
<td>7.60</td>
<td>4.60</td>
<td>3.29</td>
<td>0.49</td>
<td>9.87E-04</td>
<td>32.26</td>
<td>3.82</td>
<td>0.75</td>
<td>0.80</td>
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<td>0.58</td>
<td>6.60</td>
<td>4.20</td>
<td>7.37</td>
<td>3.77</td>
<td>6.24E-03</td>
<td>33.64</td>
<td>4.51</td>
<td>0.93</td>
<td>0.87</td>
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<tr>
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<td>16 water &amp; steam</td>
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<td>0.35</td>
<td>6.00</td>
<td>5.80</td>
<td>4.68</td>
<td>0.88</td>
<td>9.89E-04</td>
<td>38.94</td>
<td>5.59</td>
<td>0.61</td>
<td>0.63</td>
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<tr>
<td>EB29</td>
<td>16 water</td>
<td>6.41</td>
<td>0.17</td>
<td>14.40</td>
<td>10.80</td>
<td>21.53</td>
<td>20.53</td>
<td>1.44E-03</td>
<td>17.28</td>
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<td>2049.80</td>
<td>13.41</td>
<td>n.p.</td>
<td>3.52E-04</td>
<td>18000.00</td>
<td>3504.29</td>
<td>0.94</td>
<td>0.52</td>
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<td>544.00</td>
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<td>n.p.</td>
<td>1.31E-03</td>
<td>200.00</td>
<td>1.96</td>
<td>0.12</td>
<td>0.15</td>
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<td>16 water</td>
<td>9.96</td>
<td>0.32</td>
<td>7.60</td>
<td>8.00</td>
<td>9.88</td>
<td>6.28</td>
<td>7.46E-04</td>
<td>22.12</td>
<td>4.29</td>
<td>0.40</td>
<td>0.44</td>
<td></td>
</tr>
<tr>
<td>EB31_f</td>
<td>0 nitrogen</td>
<td>3.20</td>
<td>2.6E-03</td>
<td>475.00</td>
<td>289.80</td>
<td>20.81</td>
<td>n.p.</td>
<td>1.76E-04</td>
<td>970.50</td>
<td>318.77</td>
<td>0.94</td>
<td>0.68</td>
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Table 6.3: reduced list of variables and features of the long-duration AE activity.

<table>
<thead>
<tr>
<th>Test #</th>
<th>Fluid Phases</th>
<th>$p_p$ Drop Peak (MPa)</th>
<th>$p_p$ drop FWHM (ms)</th>
<th>Signal's envelope peak (V)</th>
<th>Envelope peak length (ms)</th>
<th>Envelope’s FWHM (ms)</th>
<th>Envelope – $p_p$ drop correlation coefficient</th>
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<tbody>
<tr>
<td>EB16</td>
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<td>4.48</td>
<td>11.00</td>
<td>5.33E-04</td>
<td>22.58</td>
<td>8.49</td>
<td>0.84</td>
</tr>
<tr>
<td>EB18</td>
<td>water &amp; steam</td>
<td>4.58</td>
<td>7.60</td>
<td>3.21E-04</td>
<td>26.27</td>
<td>7.27</td>
<td>0.93</td>
</tr>
<tr>
<td>EB19</td>
<td>water &amp; steam</td>
<td>5.00</td>
<td>7.60</td>
<td>9.87E-04</td>
<td>32.26</td>
<td>3.82</td>
<td>0.75</td>
</tr>
<tr>
<td>EB27</td>
<td>water &amp; steam</td>
<td>14.51</td>
<td>6.60</td>
<td>6.24E-03</td>
<td>33.64</td>
<td>4.51</td>
<td>0.93</td>
</tr>
<tr>
<td>EB28</td>
<td>water &amp; steam</td>
<td>13.29</td>
<td>6.00</td>
<td>9.89E-04</td>
<td>38.94</td>
<td>5.59</td>
<td>0.61</td>
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<tr>
<td>EB29</td>
<td>water</td>
<td>6.41</td>
<td>14.40</td>
<td>1.44E-03</td>
<td>17.28</td>
<td>7.32</td>
<td>0.35</td>
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<tr>
<td>EB31</td>
<td>nitrogen</td>
<td>5.14</td>
<td>3100.60</td>
<td>3.52E-04</td>
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<td>3504.29</td>
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<tr>
<td>EB33</td>
<td>nitrogen</td>
<td>1.47</td>
<td>1147.00</td>
<td>1.31E-03</td>
<td>200.00</td>
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<tr>
<td>EB34</td>
<td>water</td>
<td>9.96</td>
<td>7.60</td>
<td>7.46E-04</td>
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<td>EB31_f</td>
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<td>3.20</td>
<td>475.00</td>
<td>1.76E-04</td>
<td>970.50</td>
<td>318.77</td>
<td>0.94</td>
</tr>
</tbody>
</table>
In addition, the lag time between the $p_p$ drop and the AE activity, with the latter starting a few milliseconds after, further confirms that microseismic activity is caused by fluid-movement. Due to the high compressibility of gas the initial pressure decay rate is much slower (3 MPa/2 s) than the initial pressure decay of both low-temperature (LT, 25°C) and high-temperature (HT, 175°C) water (3 MPa/0.01 s). A slow decay of the pressure (0.9 MPa/several minutes) is also observed for HT water, when the mixture of water and steam is released, after $p_p$ stabilizes at around 0.9 MPa. At this pressure/temperature condition, a phase transformation occurs throughout the sample, causing the superheated water to transform in steam. As steam is a gaseous phase, this slow decay after phase transformation is further evidence of gas compressibility. In such a scenario, the properties of the overall mixture has the compressibility of the gas phase (Kieffer, 1977).

However, in EB31 ($p_p = 10$ MPa) and EB33 ($p_p = 5$ MPa) the pressure drops of the same magnitude (few MPa) of the other cases were still achieved. A possible explanation is that such high pressure drops are due to the bottleneck effect, which impedes (but does not stop) the decrease in pore pressure as measured at the top of the sample, but does act to increase the pressure differential between top and bottom. Another variable, which can be discarded for the purpose of this analysis, is the initial pore pressure, as the pore pressure drop better reflects changing conditions during the fluid movements. After this step, the number of variables is reduced to seven (Table 6.3). The relation between these variables are summarised in Figure 6.20 and 6.21.

The peak pressure drop and AE signal envelope analysis is presented in Figure 6.20. Discarding experiment EB33 (which has a small pressure drop and an irregular signal), a positive correlation is derived, linking the pressure drop to the AE envelope peak regardless of the fluid phases involved. Passing from a pure liquid phase to a pure gas phase, a large increase (a factor of 2) in the FWHM of the pressure drop curve is evident (Fig 6.21). As discussed above, this is due to the slower pressure decay of gases. In turn, the FWHM of the pressure drop is in the same order of magnitude (if not the same length) of FWHM of the envelope peak. Finally, as the percentage of gas increases, the correlation coefficient between the envelope of the signal and the pressure drop curve starts to increase, and is clearly visible (Fig. 6.21). For the water-release experiments (Fig. 6.21a, b), the signal envelope (red line) matches the first peak in the pressure drop (black line), but does not follow the other peaks in the pressure drop curve, which follow the highest peak.
When using a pore fluid mixture of water and steam (Fig. 6.21c and d, induced by the release of water pressure above the boiling point of water), new peaks of the pressure drop curve are detected by the signal envelope, but with different length and shape. In EB18 (Fig. 6.21c) a pressure drop transient around 60 ms is followed 20 ms later by a small peak in the AE signal envelope. In EB28 (Fig. 6.21d), a similar situation is present: a pressure transient occurs earlier, at 20 ms, which is imaged by a small envelope’s peak 5 ms later.

Finally, for the gas-release experiments (Fig. 6.21e, f), there is an excellent match between the two curves in particular during the venting of EB31 (Fig. 6.21e), where two smaller pressure transients are detected and quantified by the signal envelope. Overall, the correlation coefficient ranges from an average of 0.53 (water), to 0.81 (water and steam) and 0.94 (nitrogen).

Figure 6.20: pressure drop peak (x axis) versus envelope peak (y axis). The labelled point indicates the discarded experiment.
Figure 6.21: normalized signal envelopes (grey lines) superimposed on the normalized absolute pressure drop (black lines) for a & b) the water-release ($T = 25^\circ\text{C}$), c & d) water/steam-release ($T = 175^\circ\text{C}$) and e & f) gas-release ($T = 25^\circ\text{C}$) experiments. As the gas fraction increases (from top to bottom), the match between the signal’s envelope and the normalized absolute pressure drop increases. In fact while with liquid water (a, b) smaller $p_p$ transients (either preceding or following the main peak) do not match the signal’s envelope, with higher gas fractions (c, d), smaller $p_p$ transients are detected. Finally when only gas is present (e, f), smaller $p_p$ transients perfectly match with the signal’s envelope. From a) to e) gas is released after starting $p_p$ is reached, while in f) gas was still supplied to reach desired starting $p_p$. 
The key difference between long-duration signals produced by different fluid phases can be recognized also in the frequency-domain. In the previous chapter, AE waveforms were accompanied by their spectrograms, showing particular broadband features when signals are generated by a pure liquid phase, and a banded character when a gas fraction is added. Figure 6.22 shows the results of an analysis in terms of the phase-dependent frequency content by comparing the frequency spectra of three selected long-duration AE signals. Here the spectral amplitude is normalized and in decibel units. Zero is defined as the dominant frequency of the signal, and -6 dB line (in red) marks the FWHM of the frequency peak.

The pressure release of a pure liquid phase generates a signal with a broad frequency peak, with FWHM of 90 kHz, and is continuous between 60 and 150 kHz (Fig. 6.22a) with a dominant frequency centred at 80 kHz. When steam is present together with liquid water (Fig. 6.22b), the FWHM of the frequency peak is lower, ranging from 90 to 110 kHz and with a peak at 105 kHz. In the spectrum as shown in Fig. 6.22b, it is also possible to notice a slight increase in amplitude (> -10 dB) at 170 kHz. Finally, for the case of a pure gas phase (Fig. 6.22c), a double peak feature is evident, with a shift in the dominant frequency to around 60 kHz and a shorter FWHM, ranging from 58 to 65 kHz. The second peak is broader, spanning 105 to 155 kHz.

One aspect of these changes warrants further discussion: the narrowing of the spectral peak (which corresponds to an increase of the Quality factor, $Q$, calculated as the ratio between the frequency peak and the width of the frequency peak at FWHM). This effect can be explained by using the Kumagai & Chouet (2000, 2001) model. According to this model, variation in $Q$ depends on the ratio between the compressional wave velocity of the rock matrix ($\alpha$) and the sound of speed of the fluid ($a$):

$$Q \propto \frac{\alpha}{a} \quad \text{(Eq. 6.4)}$$

Assuming that the compressional wave velocity of the rock matrix, then:

$$Q \propto \frac{1}{a} \quad \text{(Eq. 6.5)}$$

Therefore $Q$ is inversely proportional to the density of the fluid. In the experiments described in this study, a switch from liquid phase to a gas phase corresponds to a decrease in the sound of speed (from about 1500 m/s to about 350 m/s) which effectively resulted in an increase of $Q$ of the signals (calculated values are 1, 5.3 and 8.6 for the water, the water-steam mixture and nitrogen.
released respectively) Higher $Q$ are interpreted as reflecting different physical properties of the fluids, particularly an increase in the gas fraction as reported by Chouet (1992), allowing a larger impedance contrast between rock and fluid and resulting in an under-damped signal which slowly decays.
Figure 6.22: top) waveform and bottom) frequency spectrum in decibel units for a) the water (EB34, $p_p = 16$ MPa, $T = 25^\circ$C), b) water-steam mixture (EB28, $p_p = 16$ MPa, $T = 175^\circ$C) and c) nitrogen (EB31, $p_p = 10$ MPa, $T = 25^\circ$C) releases. The dashed horizontal line mark -6 dB level, which corresponds at FWHM of the spectral amplitude. When liquid water is depressurized (a) a 10-ms-long signal is generated, characterized by a broadband spectrum, covering the bandwidth 60 –150 kHz. A longer signal (30 ms) is produced when a fluid formed by a mixture of water and steam is depressurized (b). The resultant spectrum presents a single narrower peak at 100 kHz (however a second peak around 160-180 kHz is visible in the spectrogram in Fig. 5.61) with a FWHM around 90–110 kHz. With the presence of a gas phase only (c), the depressurization of the fluid takes much longer time, producing a 16s long signal. The spectrum under such conditions is characterized by a narrow peak around 60 kHz (FWHM between 58 and 65 kHz) and a much broader peak around 130 kHz (between 100 and 160 kHz).
6.3.3. RELATIONSHIPS BETWEEN LONG- AND SHORT-DURATION AE SIGNALS

Chapter 5 presented a number of short-duration AE events extracted from the continuous datasets, often exhibiting particular timing and dominant frequency characteristics, which are discussed below.

Regarding the timing, it is straightforward to observe that the majority of the events occurs immediately after the pressure release. Most of the extracted events analysed occurs after the long-duration signal, because the superposition of both signals masks the first arrival of the short-duration event. These short-duration AE signals (or pulses, such as in Fig. 6.22c at 11 s after release) appear to be generated simultaneously with the longer-lasting pseudo-continuous AE activity. This type of dual behaviour is interpreted as result of a small region of the sample (for instance a section of the damage zone) where the pressure decay could not be discharged fast enough (see Fig. 6.17b).

As the pressure drop decreases, together with a generally longer-lasting signal, fewer events are extracted. Although both EB34 (Fig. 6.23a) and EB28 (Fig. 6.23b) still have residual pressure drop even one minute after initial pore pressure release, EB31 (Fig. 6.23c), which was pressurised with nitrogen gas, has no pressure drop after 40 s which coincided with an abrupt end of AE activity.

Therefore, as observed for the long-duration AE activity, the fluid speed played a major role in the generation of microseismic activity. When a $p_p$ difference occurred between the two ends of the sample but the flow speed is not fast enough to generate a continuous signal, it caused AE pulses (Fig. 6.23a, b). When the same amount of $p_p$ is present both at top and bottom ends of the sample, a pore pressure drop could not exists, hence the flow speed is zero. As a consequence no AE activity is generated (Fig. 6.23c).
Figure 6.23: temporal evolution of the cumulative hits and pore pressure for a) EB34 (water, $p_p = 16$ MPa, $T = 25^\circ$C), b) EB28 (water and steam, $p_p = 16$ MPa, $T = 175^\circ$C) and c) EB31 (nitrogen, $p_p = 10$ MPa, $T = 25^\circ$C).
In the frequency-domain, the main features observed for the long-duration signals are still preserved in short-duration events, namely a broadband spectrum with low $Q$ for the liquid water case, and banded spectra with high $Q$ for the water-steam and nitrogen gas release with the latter having the highest $Q$ (meaning smallest FWHM, table 6.1). These characteristics are shown in Fig. 6.24. For the liquid water case the average value of $Q$ is 2, whose variation in time (Fig. 6.24a) or in frequency (Fig. 6.24b) does not show any significant trend. When a mixture of water and steam is depressurized, the average value of $Q$ is 4.4. However, for such case, an increase of $Q$ in the latest events appears (Fig. 6.24c), which is thought to a result of an increase in gas-fraction in the fluid as more superheated water is transformed to steam. In addition, as presented in section 5.7, the AEs collected in these conditions show a bimodality, not only in the distribution of dominant frequency, showing two peaks at 100 and 160 kHz respectively (Fig. 6.25a, b), but also within the individual spectrum of each event, with 50% of the events lying in the bimodality field (Fig. 6.25c).
Figure 6.24: variation of $Q$ in left) time and in right) frequency, for the depressurization of a & b) liquid water (EB34, $p_p = 16$ MPa, $T = 25^\circ$C), c & d) water and steam mixture (EB28, $p_p = 16$ MPa, $T = 175^\circ$C) and e & f) nitrogen gas (EB31, $p_p = 10$ MPa, $T = 25^\circ$C).
Figure 6.25: a) temporal evolution of dominant frequency and pore pressure, b) characteristic FWHM and c) bimodality evaluation of the AE spectra of experiment EB28 (water and steam, $p_r = 16$ MPa, $T = 175^\circ$C).
This bimodality is reflected in the distribution of $Q$ (Fig. 6.24d), which presents two clusters: the cluster at lower frequency has an average quality factor of 3.9, while the cluster at higher frequency has a $Q$ of 5.1. This pattern is consistent with Kumagai & Chouet model (2000, 2001), where an increase in gas fraction correspond an increase in the resonance frequency. The increase in frequency ($f$) is the result of a combination of two factors, namely i) the decrease in the ratio between the compressional wave velocity of the rock matrix ($\alpha$) and the sound of speed of the fluid ($a$) and ii) the decrease in the ratio between the density of the fluid ($\rho_f$) and the density of the solid ($\rho_s$) (Kumagai & Chouet, 2000, 2001):

$$\frac{1}{f} \propto \frac{\alpha}{a} \text{ and } \frac{1}{f} \propto \frac{\rho_f}{\rho_s}$$

(Eq. 6.6)

Assuming that the rock properties are constant, then:

$$f \propto a \text{ and } \frac{1}{f} \propto \rho_f$$

(Eq. 6.7)

The $\rho_f$ of water is 1000 kg/m$^3$, while $\rho_f$ of nitrogen is 1.3 kg/m$^3$. $\rho_f$ of the water and steam mixture depends on the fraction of the gas phase in the fluid, which can go down to 0.6 kg/m$^3$ (White, 2002). This decrease in density actually results in an increase of the frequency. The variation of the acoustic velocity of the fluid has been discussed in section 6.3.2, where it is shown that as the gas fraction increases, the sound of speed of the fluid decreases. This should result in a decrease of the frequency. However, while the sound of speed of the gas phase is about 4 times lower than the sound of speed of the liquid phase, the density of the gas phase is around 1000 times lower than the density of the liquid phase. Therefore, in the conditions simulated in these experiments, the density of fluid is the major control on the frequency content, while the sound of speed of the fluid controls $Q$.

The case of joint water and steam release, is of particular interest because it provokes two well-defined peaks are observed during the discharge of fluids with two different phase. Thus far it has been assumed that the type of phases (gas or liquid) has a major role in the generation of different AE signals (in terms of duration and frequency content) and it is now possible to infer that also the number of phases present in the system directly controls the characteristic frequency spectrum of the signals.
The lack of relationship between $Q$ and the dominant frequency is also evident for the case of the nitrogen release, similarly to the water case (single-phase fluid). Neither temporal variations (Fig. 6.24e), nor frequency dependence (Fig. 6.24f) have been observed for $Q$ value.

Under these conditions, bimodality is only observed in the long-duration AE signal (peaks at 60 and 130 kHz, Fig. 6.26), while the distribution of frequencies do not form clusters around any specific value (Fig. 6.27). A lack of bimodality is also present in EB33, where the release of gas nitrogen was not able to generate a long-duration event (Fig. 6.26b) of the same size of signal in EB31 (Fig. 6.26a).

Figure 6.26: top) continuous AE signal superimposed on the pore pressure time plot and bottom) associated spectrogram for a) EB31 (nitrogen, $p_p = 10$ MPa, $T = 25^\circ$C) and b) EB33 (nitrogen, $p_p = 5$ MPa, $T = 25^\circ$C).
Figure 6.27: top) dominant frequency and bottom) FWHM of the AE events of a) EB31 (nitrogen, $p_p = 10$ MPa, $T = 25^\circ$C) and b) EB33 (nitrogen, $p_p = 5$ MPa, $T = 25^\circ$C).

Zobin et al. (2009) noted two, separated in time, frequency peaks in volcanic signals related to Vulcanian explosions at the andesitic volcanoes Colima and Popocatepetl in Mexico. A first low frequency peak is likely due to the passage of a gas bubble and is followed by a high frequency peak, thought to be caused by the vibration of the shallow conduit. While two frequency peaks (60 and 130 kHz) are clearly visible in the long-duration AE (Fig. 6.26), the lack of bimodality in the short-duration AEs is a sign of a non-simultaneity of the frequency peaks (Fig. 6.27). Therefore, Zobin’s interpretation fits very well the laboratory data. First, it explains the presence of two spectral peaks in the long-duration AE, which is particularly evident in EB31 where a high pressure drop likely generates a continuous vibration of the damage zone. Second, it explains the lack of bimodality in short-duration AEs, as the vibration of the conduit is a consequence of the passage of the gas bubble. However, this interpretation does not apply for the water & steam release, because the high degree of bimodality (50%) means that the two frequency peaks (100 and 160 kHz) are
simultaneous. Here it is likely that the cause of the cause of two simultaneous frequency peaks is the coexistence of two fluid phases, which are water and steam.

6.3.4. WAVEFORM SIMILARITY: SHORT-DURATION AE EVENTS

Although only few events for each experiment are localized, the location and repeatability of the AE source can be evaluated through the similarity of the waveforms recorded in each experiment. In order to establish the similarity of the events of each experiment, a cross-correlation analysis was performed using the bridging technique (Barani et al., 2007; Alparone et al., 2010), as described in section 4.8. The results for the most representative cases each condition (liquid water, water & steam, nitrogen gas), shown in Fig. 6.28, are put together in Fig. 6.29. Around 50% of the events recorded during venting in each experiment are correlated using the longest window-length (50 μs), with a value above 90% for all experiments with a window-length up to 20 μs.

The high level of waveform similarity revealed by the cross-correlation analysis is a consequence of a non-destructive, stationary source which characterized fluid-induced seismic activity. This further corroborates the hypothesis that short-duration AEs, similarly to the long-duration events, are generated by fluid movement processes.

Figure 6.28: percentage of correlated events recognized for a) EB34 (water, $p_p = 16$ MPa, $T = 25^\circ$C), b) EB28 (water and steam, $p_p = 16$ MPa, $T = 175^\circ$C) and c) EB31 ($p_p = 10$ MPa, $T = 25^\circ$C) using the bridging technique.
Figure 6.29: percentage of correlated events, using a cross-correlation threshold of 0.8, for the events generated by the release of liquid water (triangles, EB34, $p_r = 16$ MPa, $T = 25^\circ$C), water and steam mixture (circles, EB28, $p_r = 16$ MPa, $T = 175^\circ$C) and nitrogen gas (diamonds, EB31, $p_r = 10$ MPa, $T = 25^\circ$C) versus the window length used to calculate the cross-correlation coefficient between each pair of events.

6.3.5. SOURCE COMPONENTS OF THE SHORT-DURATION AE EVENTS

A final analysis method takes the relative percentage of the source components of the short-duration events. This is challenging for the venting stage as the number of events that are well localised is very small, and the source components of the venting at low temperature (with water as pore fluid) is likewise very weak. In fact, as one can see from Fig. 6.30, the amplitude of the signals recorded during the deformation stage (a) is much higher than the signal acquired during the venting (b). As a consequence fewer events are extracted, which are not picked by all sensors and have smaller amplitudes and smaller SNR. Note that the signal in Fig. 6.30b is the same signal shown in Fig. 6.22b, but at different voltage scale.
Therefore, while the locations of the events (10 at best) gives no significant information, the information about their source mechanism is very useful. This is achieved by combining all the localised events of all experiments at the same conditions, with a condition number less than 100 (signifying better resolved solutions, Pettitt, 1998), to derive a solution for the source components.

Figure 6.31 shows the average percentage of each source component (double-couple, DC, compensated linear vector dipole, CLVD, and volumetric, ISO. See section 2.4 for details) for each pressure release conditions. While the DC component still represents between 33 – 48% the source mechanism, in all cases a relative high percentage (between 30 and 40%) of CLVD is present which is consistent for a fluid-movement source process and in agreement with Julian model (1994). The higher percentage of ISO (about 30%) in the water-steam case is attributed to the expansion of the gas after the phase transformation when the overheated water is depressurized.
Figure 6.31: average source components (DC = double-couple, CLVD = compensated linear vector dipole, ISO = volumetric; for details see section 2.4) for all experiments releasing a) liquid water (T = 25°C), b) water and steam mixture (T = 175°C) and c) nitrogen gas (T = 25°C).
6.4. FROM THE LABORATORY TO THE FIELD

During the venting stage of these experiments, signals of different duration, amplitude and spectral contents are produced by the decompression of the pore fluids at different temperature, pressure and phase conditions. What is most crucial is that these parameters are known and measured, whilst in the field most of these are calculated and the source mechanism and other key parameters are often inferred. Therefore, correlating field events with their simulated laboratory events is extremely useful for better understanding the physics behind the generation of the volcanic earthquakes. Here, the focus is on a qualitatively comparison between the long- and short-duration AEs and the volcanic signals generally thought to be produced by the movement of fluids, i.e. Tornillo, tremor, LP and VLP events (Table 6.4).

Table 6.4: summary of the venting stage characteristics and associated field analogue

<table>
<thead>
<tr>
<th>Fluid phase involved</th>
<th>Flow speed</th>
<th>AE duration</th>
<th>Frequency content</th>
<th>Field analogue</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gas</td>
<td>Fast</td>
<td>Long</td>
<td>Bimodal</td>
<td>Tornillo</td>
</tr>
<tr>
<td>Liquid and gas</td>
<td>Fast</td>
<td>Long</td>
<td>Bimodal</td>
<td>Tremor</td>
</tr>
<tr>
<td>Liquid and gas</td>
<td>Slow</td>
<td>Short</td>
<td>Bimodal</td>
<td>LP</td>
</tr>
<tr>
<td>Gas</td>
<td>Slow</td>
<td>Short</td>
<td>Unimodal</td>
<td>VLP</td>
</tr>
</tbody>
</table>

A close similarity is observed between the long-duration AE event generated by the gas release and a classic Tornillo events, as recorded, for example, on Vulcano Island on November 7, 2008 (Fig. 6.32). Two clear features emerge both in both the time- and frequency domain. Firstly, the amplitude modulation (Fig. 6.32c) and secondly, the presence of sharp, well constrained spectral peaks (Fig. 6.32d). These same features appear in the gas-generated long-duration event generated in the laboratory (Fig. 6.32a, b), for which it is known that the signal is caused by the flow speed and the physical state of the fluid involved. The role and contribution of the gas phase has been already discussed in the literature (e.g. Alparone et al., 2010) and in this study in section 6.3.3, but the direct relationship between flow speed and amplitude modulation has never been linked. In fact, while the flow speed can be measured (or calculated through the pressure drop) in laboratory conditions, this is not possible in the field, particularly when the phenomenon occurs inside the volcanic edifice. Even though these events occur under different volcanic phases, such as pre, post
eruption, during seismic crisis and in quiescence phases (Gomez & Torres, 1997), the study of these signals can give important information regarding the speed and volume of the moving fluid. Interestingly, this example also acts over similar time-scales for both laboratory and field events, with duration of few tens of seconds. The pore pressure used to generate the laboratory event correspond to a depth of 1 km, which is in the range of the source locations of Tornillos recorded on Vulcano (Milluzzo et al., 2010). Therefore, by simulating pressure conditions typical of the field scale, it is now possible to generate a signal with duration on the same order of magnitude of its field analogue. However, because the resonating fluid-filled crack is about $10^4$ times smaller in the rock sample (few centimetres, compared to the few hundred meters of natural faults and fractures), the spectral peaks are about $10^4$ times higher (60 – 200 kHz in the laboratory, 6 – 10 Hz in the field), in agreement with the simplified scaling law proposed by Burlini et al. (2007; 2009) and Benson et al. (2008):

$$d_n 	imes f_n = d_l 	imes f_l$$  \hspace{1cm} (Eq. 6.8)

Figure 6.32: a) waveform and b) spectrogram of the long-duration signal generated during the gas release (EB31, $T = 25^\circ$C); c) waveform and d) spectrogram of the Tornillo event recorded on Vulcano Island on November 7, 2008. Note the similar time scale, the same amplitude modulation effect and the narrow spectral peaks.
Long-duration AE during water and steam depressurization (Fig. 6.33a) is compared with volcanic tremor (Fig. 6.33d) recorded at Mt. Etna on December 2, 2015, during strombolian activity preceding lava fountain activity (Alparone, personal communication). Although a banded spectrogram is only visible in the laboratory event (Fig. 6.33c), the analysis of the frequency spectrum reveals that both laboratory (Fig. 6.33e) and volcanic (Fig. 6.33f) signals have few narrow spectral peaks. The laboratory and field signals not only share a similar spectrum, but also a similar source mechanism, which is the decompression of a fluid composed by both a liquid and gas phase. The results suggest that the two spectral peaks found in the laboratory correspond to the two different phases, water and steam, forming the fluid. Therefore it is likely that the same conclusion is applicable to the field event: the number of spectral peaks characterizing the seismic signal can reveal the number of phases that are present in the magma.
Finally, the short-duration AE signals show similarities with discrete LP and VLP events (Fig. 6.34 and 6.35 respectively), particularly the signals generated by the release of a mixture of gas and steam and of nitrogen gas. Figure 6.34 shows the waveform and spectrogram of a typical signal generated during the water/steam release characterized by two spectral peaks, with panels (c) and (d) respectively showing the waveform and spectrogram of an LP event recorded on Mt Etna on January 10, 2015. This event was produced during continuous strombolian and ash emission activity at the Voragine crater (Alparone, personal communication) and is characterized by three
spectral peaks, at 5, 6 and 8 Hz. As for the long-duration AE and tremor activity comparison, the same conclusion applies here, where more than one phase is present in the system. Therefore it is likely that fluid phases play a key role in the character of the signal. In addition, the similarity between long-duration activity and short-duration activity, both in the laboratory and in field is excellent, and further confirms the Julian (1994) model on the role of flow speed. In this way, volcanic tremor and LP events share the same source (meaning same frequency content), but different flow speeds (resulting in different durations).

![Figure 6.34: a) waveform and b) spectrogram of the short-duration signal generated during the water and steam release (EB28, T = 175°C); c) waveform and d) spectrogram of the LP event recorded on Mt Etna on January 10, 2015. Note the similar banded spectrograms.](image)

Though signals at lower frequencies are more difficult to record (due to the background noise, which has the same frequencies in the laboratory, or likewise a very low SNR in the field), the short-duration AEs (Fig. 6.35a, b) generated by the release of gas pressure share similar spectral characteristics with VLP signals (Fig. 6.35c, d) in the field. The VLP event shown here was recorded on Mt Etna on September 5, 2013 before explosive activity at the New South East Crater (Alparone, personal communication). This activity is characterized by higher content of gas
compared to the strombolian activity associated with the LP event. As a result the spectrogram is characterized by narrower spectral peaks, with the dominant frequency lying at 0.5 Hz. In addition a weak later peak at higher frequency (2.5 Hz) is present at 17 s (Fig. 6.35d). A similar behaviour occurs for its laboratory analogue: a narrow spectral peak around 100 kHz dominates the spectrogram, while a weak, later (~150 µs) peak lies at 200 kHz. The narrowing of the spectral peaks and the presence of a later higher frequency peak, confirms the fundamental role played by the gas phase in the generation of VLP activity. Therefore the analysis of such events may be used to constrain the gas volume, hence the mass transport budget, as initially postulated by Chouet (2003).

Figure 6.35: a) waveform and b) spectrogram of the short-duration signal generated during the gas release (EB31, T = 25°C); c) waveform and d) spectrogram of the VLP event recorded on Mt Etna on September 5, 2013. Note the similar narrow spectral peak at lower frequencies.

6.5. UNCERTAINTIES

Throughout the acquisition and analysis of the data a number of uncertainties are generated due to the usual inaccuracies inherent in a laboratory work. A brief discussion of the main uncertainties and errors follows.

First of all, the sensor array consists of 12 PZT sensors placed around the rock sample cylinder (specimen), but no sensors are located on the upper nor the lower basal surfaces, orthogonal to the
maximum stress ($\sigma_1$) during the triaxial deformation stage. As the deformation proceeds, cracks orthogonal to $\sigma_1$ closes followed by the opening of new fracture parallel to $\sigma_1$, making the axis of the cylinder the direction of the fastest elastic velocity until the generation of the damage zone (Benson et al., 2007). Because of the absence of basal PZT sensors, the elastic velocity recorded at 62° (the most inclined) to the horizontal line are used to build a velocity model, relying on data fitting to infer the maximum at 90° via stereonet fitting. This is proven to be a good assumption due to the good overall coverage. To confirm, while locating the shot of the velocity surveys, at least 6 shots are located with errors less than 4 mm (10% of the sample’s diameter).

When the velocity model is applied to locate the events around the time of failure (sections 5.3.2, 5.4.2 and 5.5.2) both static and overlapping windows methods reveal good results, when compared to the actual fault observed in the post-test sample, with location errors generally below 4 mm. Here the simplex algorithm uses a Time-Dependent Transversely-Isotropic velocity model, which takes into account an ongoing deformation along a unique preferential direction, which is the case of triaxial experiments. However, most of the samples do not present a single through-going fault, but a system of conjugate faults and other secondary fractures. In such cases the velocity models applied do not represent the reality. In most cases this velocity model can be assumed to be a good approximation (via hypocentres’ alignment and observed fault match). However, in other cases either a widespread cloud of events or a different alignment of the hypocentres do not image the fault. Therefore a more accurate velocity model, which takes into account different directions of low velocity zone, is used to resolve such issues.

Another aspect concerning the PZT sensors is their sensitivity, which is fundamental to retrieve the MT solution of each AE signal. While a calibration of the sensors to determine the sensitivity has not been exhaustively performed, a relative sensitivity has been conducted. This consists of conducting a velocity survey on an aluminium cylinder, which is an isotropic and perfectly elastic material, with no differential stress applied. To avoid azimuthal dependency (Pettitt, 1998), only the amplitude recorded opposite the transmitter was taken into account. By comparing the 12 amplitudes, the relative sensitivity of each sensor has been calculated. This procedure has been performed before each experiments and the results show that between 8 and 10 sensors have similar sensitivity. The relative sensitivity is then used to calculated a relative MT solution and retrieve the source components, as shown in Figure 6.31.
Secondly, whilst water is injected inside the sample through pumps, which allows precise volume and pressure control, gaseous nitrogen is stored in a bottle and only controlled via a standard regulator (pressure control only). This means that the exact volume of gas injected is unknown and any attempt to correlate the AE amplitude with the precise gas volume is prevented. For the same reason, i.e. the absence of servo-controlled gas pumps, deformation experiments are not run under gas saturated conditions, and the effects of gas pore pressure on the sample failure is not analysed.

By the same token, whilst dilatancy can be studied in the saturated experiments and compared with the microseismic activity, this is not possible for the dry experiments. In fact, the dilatancy measurements may only be achieved through the pore volume change data, which is only recorded when a pressurized fluid is present inside the sample. The lack of information about dilatancy in dry samples prevents a direct comparison between fracturing in dry and saturated conditions.

Finally, in Figure 6.17 one can notice delayed AE onsets in respect of the start of the pore pressure drop, particularly for the shortest time-scale (a – d). This time lag is likely caused by positions of the AE sensors and the pore pressure transducers. While the former are directly in contact with rock sample, the latter are placed about 1 m away from the sample. Therefore there are some practical limits in establish a direct link between the time onsets of AE with the timing of the pore pressure drop.
7. CONCLUSIONS

This research aims to investigate a number of open questions in volcano-seismology, particularly regarding the origin of low-frequency seismicity, which is recorded in numerous forms in volcanoes around the world. It is now widely accepted that fluids are the key to understand such phenomena (e.g. Chouet, 1996) including their role in fracturing and subsequent fluid-enhanced seismicity. In this thesis, new research is presented that tests these ideas in a controlled laboratory setting to better understand details of the fluid pressure and pressure change. This is achieved by performing 2-stage experiments on basalt specimens from Mt. Etna, Italy, using a triaxial apparatus.

During the first stage of the experiments, the specimens underwent triaxial deformation under different confining and pore pressure conditions, where the micro-seismicity (Acoustic Emission (AE) activity) was recorded by state-of-the-art equipment fitted with newly developed low-frequency sensors developed as part of this research. This stage concludes with the failure of the samples and the formation of a shear zone inclined at 30° to the vertical axis. This is then used for the second stage where pressurized fluids are intentionally vented out from the natural fracture, imaged by using AE hypocentre location analysis. In agreement with previous studies (Heap et al., 2009, Benson et al., 2010), the presence of pressurized fluid had no effect on the Young’s modulus, while the peak stress decreases as the pore pressure increases.

A first conclusion on this stage of the experiments is that the presence of pore pressure has a significant effect on the compressional wave velocity and velocity anisotropy, consistent with the findings of numerous previous studies (e.g. Vinciguerra et al., 2005; Benson et al., 2007). However, although pressurized fluids increase the P-wave velocity by enhancing seismic wave propagation, they also have the converse effect: as the pore pressure increases, the dynamic P-wave anisotropy, as the deformation proceeds, is reduced. This aspect, previously not fully understood in past research, is consistent throughout different specimens run at the same pressure conditions. In particular, although a marked shift in the concavity of the anisotropy curve is seen in dry conditions (Fig. 6.2) and with a sharp increase at the time of failure, in saturated conditions, this trend is quasi-linear throughout the whole deformation stage (little or no increase in anisotropy). This effect is
likely linked to same enhanced seismic wave propagation, which is in turn aided by the fluids filling the newly-formed crack, hence reducing the P-wave anisotropy.

The second conclusion from the deformation stage is that pore pressure also has a significant effect on the general fracturing process. Here, this study shows that the onset of dilatancy is delayed as the pore pressure increases, with a consequent delay in the build-up in microseismic event rate. AE activity starts later than dilatancy, with the same lag regardless of pore pressure. This suggests that the fracturing commences with signals below the detection limit/background noise of the instrumentation used in this study. As the pore pressure is introduced and increased, the AE rate passes from an early onset with constantly accelerating AE rate, to a late onset and sudden supra-exponential increase some tens of seconds before the failure. Noticing that the final result of the deformation stage is a shear zone whose character does not depend on the amount of pore pressure, it is then possible to link the sudden increase to the needs of speeding up the fracturing process in order to achieve the same amount of cracking at the time of failure under all conditions. Therefore, around the time of failure properties such as the seismic b-value, magnitude and inter-event distances are not affected by the presence of pressurized fluids.

In order to access the damage zone and rapidly release the pore pressure, samples were pre-drilled throughout their vertical axis with a 3-mm-wide core diamond drill. Due to the limited surface area of the conduit, compared to area of the outer cylinder (which cover 99.4 % of the total surface area), the conduit has no impact on the mechanical and elastic wave properties, or on the AE activity.

During the second stage of the experiments, the pressurized fluids stored inside the sample porosity (including the fracture damage zone form stage 1) are rapidly releases from the previously created damage zone and through the pre-drilled axial conduit. This stage is performed under hydrostatic conditions, to prevent any displacement due to differential stress, and involves either liquid water, a mixture of superheated water and steam, or nitrogen gas.

During this stage, and for the first time, new correlations have been made between the envelope of the long-duration AE and the pressure drop curve, highlighting the key role of the flow speed (here expressed as a pressure drop). However, flow speed is not only responsible for the generation of the long-duration AE activity (as in the Julian model, 1994): it also controls the magnitude (expressed as peak envelope) and the duration of the signal. In addition, when the gas fraction
increases, the AE sensors tends to behave as pressure transducers, picking oscillations of the
pressure variation which last much longer than in the absence of a gas phase. In addition, if the
source location of the signal and the geometry of the fracture were known, an estimate of the
minimum volume of gas involved could be calculated via the magnitude of the pressure drop.
Specifically, the precise determination of the volume of gas could otherwise only be achieved if the
signal envelope was linked to the pore pressure rather than the pore pressure drop. These
information are of particular interest because it is possible to determine the gas fraction and a
minimum volume of gas through the shape of the long-duration AE. In terms of frequency content,
a decrease in quality factor ($Q$) has been calculated, in line with Kumagai & Chouet (2000, 2001)
model for fluid-filled cracks, where $Q$ is thought to be inversely proportional to the acoustic speed
of the fluid.

In conclusion, the presence of short-duration (ten to hundred microseconds) AE events supports
the hypothesis and model put forwards by Julian (1994) on the fundamental role of flow speed.
When flow speed is not fast enough to induce continuous signals, transients are generated. A
further conclusion is that the number of the fluid phases, not just the fluid type, exerts a key
control. This is supported by the presence of a single broadband peak associated to the release of
liquid water, whilst two narrower peaks were consistently linked to the case of superheated water
and steam release. When solely a gas phase is released, two spectral peaks also characterize the
frequency spectrum. In this case, although the higher frequency peak is associated to the vibration
of the fluid-filled crack, the lower frequency peak is related to the passage of the gas bubble which
generates the vibration of the crack, consistent with the model of Zobin et al. (2009). This research
confirms this, from the non-simultaneity presence of the peaks found in the frequency spectrum of
the short-duration AE signals.

SOLVING MACRO-SCALE PROBLEMS FROM MICRO-SCALE CASES

Probably the biggest challenge in geology is to understand the physical processes behind natural
phenomena. While we can observe and interpret these, a direct measurement of fundamental
parameters such as stress, pressure and temperature is often not possible. This can only be done in
controlled environments such as laboratory experiments. This thesis adds to the body of scholarly
activity in this area, by providing new insight and evidence on how micro-scale experiments shed light on the macro-scale problems in relation to volcanic earthquakes.

It has been widely discussed how the water controls the fracturing of rocks. By looking at the seismic hit rate under dry and saturated conditions, one immediately notices the great similarity between these two cases and the two precursory seismic rates at different volcanoes. The 1991 eruption at Pinatubo is preceded by a slow (days to weeks) increasing seismic rate, similar to the fracturing in dry conditions. Conversely, the 1989 eruption at Redoubt Volcano showed a sudden increase (hours) of the seismic rate, similar to the hit rate under saturated conditions. This observation leads to the idea that the warning time for a volcanic eruption maybe controlled by the amount of magmatic/hydrothermal pressure involved.

Another success achieved in this research is the simulation in laboratory of four different types of fluid-induced events which are recorded in different volcanic settings. In doing so, not only are previous models and theories confirmed, but two new concepts are discovered. The first is the role of the number of fluid phases, which emerges from the spectral analysis. By looking at the number of the spectral peaks and their amplitude, one concludes that the number of phases, type and amount of each phase plays a key role. In this way, the volcanic hazard may be better assessed when the nature of the erupted phases is known.

The second aspect, which emerges very clearly, relates the amplitude modulation effect, typical of Tornillo events, to the flow speed. This positive correlation is particular high when only a gas phase is present and may help to understand i) the pressure gradient, ii) the rising rate and time of gas bubbles and iii) the amount of gas volume inside the volcanic edifices.

**FUTURE DIRECTIONS**

This thesis provides new insight into the triggering of volcanic earthquakes through laboratory simulations, particularly, and for the first time, with respect to the nature of the pore fluid (phase) and the speed of the fluid.

While this study does not specifically aim to develop new model of eruption forecasting, it has discussed how micro-seismic rate in the laboratory matches the seismic rate in the field, showing accelerating trends as failure/eruption is approaching. In addition, other parameters such as P-wave velocity anisotropy, $b$-value, magnitude, locations and inter-event distances show a sharp change
before the sample’s failure, suggesting that these parameters can also be used, together with the seismic rate, to build a multi-parametric eruption forecasting model.

In the future, the use of better calibrated PZT sensors will bring information on the actual displacement of each crack, as well as seismic moment and moment magnitude, enabling a direct comparison between laboratory and field signals. Whilst a relative calibration has been performed in this study, verifying the usefulness of such an approach, absolute calibration (via laser interferometry or other means) will be needed in the future to match seismic data to laboratory AE in a quantitative sense. Technological developments in the laboratory are key to the development of appropriate micro-seismic monitoring in the field to yield important insights into volcanic processes, in some cases, and to provide failure precursory patterns useful for failure forecasting and hazard assessment.

The successful simulation of fluid-induced seismic signals opens a wide field of studies where the origin and the characteristics of such signals can be better understood. In particular, due to the technical limitations of our apparatus, depth higher than 1.6 km could not be simulated as well as gas pressures higher than 10 MPa. This is important in the future as low frequency seismicity is also found deeper in the crust. In addition, the direct measurement of the used flow speeds, which can be correlated to the effusion rate at volcanoes, is likely to be an important parameter defining the type of eruption.

Finally, as magmatic fluids are composed of different fluid types and phases, these require to be studied both individually and when combined. In this thesis only water (both liquid and gas phases) and nitrogen gas are investigated. However other fluids (e.g. CO₂) need attention as well as more complex fluid compositions (e.g. 3-phase fluids). The idea would be to approach the natural case as much as possible, to provide answers on the seismicity generated by magma movements. A subject that is still under debate.
8. BIBLIOGRAPHY


INGV website at http://www.ct.ingv.it/it/component/content/article.html?id=129.


MATLAB documentation at https://uk.mathworks.com/help/index.html


Società tipografica modenese.


APPENDIX

APPENDIX 1: ETHIC REVIEW FORM

Certificate of Ethics Review

<table>
<thead>
<tr>
<th>Project Title:</th>
<th>Dynamic laboratory simulations of fluid-rock coupling with</th>
</tr>
</thead>
<tbody>
<tr>
<td>User ID:</td>
<td>590069</td>
</tr>
<tr>
<td>Name:</td>
<td>Marco Fazio</td>
</tr>
<tr>
<td>Application Date:</td>
<td>21/11/2016 18:50:39</td>
</tr>
</tbody>
</table>

You must download your certificate, print a copy and keep it as a record of this review.

It is your responsibility to adhere to the University Ethics Policy and any Department/School or professional guidelines in the conduct of your study including relevant guidelines regarding health and safety of researchers and University Health and Safety Policy.

It is also your responsibility to follow University guidance on Data Protection Policy:
- General guidance for all data protection issues
- University Data Protection Policy

You are reminded that as a University of Portsmouth Researcher you are bound by the UKRIO Code of Practice for Research; any breach of this code could lead to action being taken following the University's Procedure for the Investigation of Allegations of Misconduct in Research.

Any changes on the answers to the questions reflecting the design, management or conduct of the research over the course of the project must be notified to the Faculty Ethics Committee. **Any changes that affect the answers given in the questionnaire, not reported to the Faculty Ethics Committee, will invalidate this certificate.**

This ethical review should not be used to infer any comment on the academic merits or methodology of the project. If you have not already done so, you are advised to develop a clear
protocol/proposal and ensure that it is independently reviewed by peers or others of appropriate standing. A favourable ethical opinion should not be perceived as permission to proceed with the research; there might be other matters of governance which require further consideration including the agreement of any organisation hosting the research.

**Governance Checklist**

**A1-Brief Description Of Project:** Triaxial deformation experiments and subsequent pore pressure releases are performed in order to investigate fluid-rock coupling by recording the acoustic emissions with a state of the art equipment. Experiments are conducted at different pressure, temperature and pore pressure conditions, simulating volcanic, shallow conditions. During the deformation stage microseismic events are analysed in the time-domain, focusing on the role of pore fluid in the fracturing process. Parameters under investigation consist of hits rate, b-value, location and magnitude. In addition, velocity survey were conducted to analyse the variation of P-wave velocity and P-wave anisotropy as the deformation proceeds.

During the pore release stage, where different types of fluid and phases are used, microseismic events are studied in the frequency-domain. Here the role of pore fluid and pore fluid phases in analysed in terms of frequency content, waveform similarity and envelope, quality factor.

**A2-Faculty:** Science

**A3-VoluntarilyReferToFEC:** No

**A5-AlreadyExternallyReviewed:** No

**B1-HumanParticipants:** No

**HumanParticipantsDefinition**

**B2-HumanParticipantsConfirmation:** Yes

**C6-SafetyRisksBeyondAssessment:** No  **D2-PhysicalEcologicalDamage:** No

**D4-HistoricalOrCulturalDamage:** No
E1-ContentiousOrIllegal: No

E2-SociallySensitiveIssues: No

F1-InvolvesAnimals: No

F2-HarmfulToThirdParties: No

G1-ConfirmReadEthicsPolicy: Confirmed

G2-ConfirmReadUKRIOCODEOfPractice: Confirmed

G3-ConfirmReadConcordatToSupportResearchIntegrity: Confirmed

G4-ConfirmedCorrectInformation: Confirmed
FORM UPR16
Research Ethics Review Checklist

Please include this completed form as an appendix to your thesis (see the Postgraduate Research Student Handbook for more information)

<table>
<thead>
<tr>
<th>Postgraduate Research Student (PGRS) Information</th>
<th>Student ID: 681387</th>
</tr>
</thead>
<tbody>
<tr>
<td>PGRS Name: Marco Fazio</td>
<td></td>
</tr>
<tr>
<td>Department: SEES</td>
<td></td>
</tr>
<tr>
<td>First Supervisor: Dr Philip Benson</td>
<td></td>
</tr>
<tr>
<td>Start Date: 01/02/2013</td>
<td></td>
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<tr>
<td>Study Mode and Route:</td>
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<tr>
<td>Part-time ¨</td>
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<td>Full-time ¨</td>
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<td>MD ¨</td>
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<tr>
<td>Professional Doctorate ¨</td>
<td></td>
</tr>
</tbody>
</table>

| Title of Thesis: Dynamic laboratory simulations of fluid-rock coupling with application to volcano seismicity and unrest. |
| Thesis Word Count: 57798 (excluding ancillary data) |

If you are unsure about any of the following, please contact the local representative on your Faculty Ethics Committee for advice. Please note that it is your responsibility to follow the University's Ethics Policy and any relevant University, academic or professional guidelines in the conduct of your study.

Although the Ethics Committee may have given your study a favourable opinion, the final responsibility for the ethical conduct of this work lies with the researcher(s).

UKRIO Finished Research Checklist:
(If you would like to know more about the checklist, please see your Faculty or Departmental Ethics Committee rep or see the online version of the full checklist at: http://www.ukrio.org/what-we-do/code-of-practice-for-research/)

a) Have all of your research and findings been reported accurately, honestly and within a reasonable timeframe?
   - YES
   - NO

b) Have all contributions to knowledge been acknowledged?
   - YES
   - NO

c) Have you complied with all agreements relating to intellectual property, publication and authorship?
   - YES
   - NO

d) Has your research data been retained in a secure and accessible form and will it remain so for the required duration?
   - YES
   - NO

e) Does your research comply with all legal, ethical, and contractual requirements?
   - YES
   - NO

Candidate Statement:
I have considered the ethical dimensions of the above named research project, and have successfully obtained the necessary ethical approval(s).

Ethical review number(s) from Faculty Ethics Committee (or from NRES/ScREC):

If you have not submitted your work for ethical review, and/or you have answered ‘No’ to one or more of questions a) to e), please explain below why this is so.

Neither human beings nor animals are involved in this research project.

UPR16 – August 2015
APPENDIX 2: MATLAB SCRIPTS

MATLAB CODE 1

% PLOT THE P-WAVE VELOCITIES OF INTACT SAMPLE IN A STEREONET

clear
addpath('E:\Milne_data\Portsmouth_tests\CORRECTED_MECH_DATA')

k = input('Type in test#:\n');

if k==17||k==18||k==19||k==20
    % type the path of the file with trend and plunge of each raypath
    % of sensor array before February 2015
    azi_pl = importdata ('E:\PhD files\Sensors_calibration\azimuth_plunge.xlsx');
    azi_pl (143,:) = [];
else
    % type the path of the file with trend and plunge of each raypath
    % of sensor array after February 2015
    azi_pl = importdata ('E:\PhD files\Sensors_calibration\new_azimuth_plunge.xlsx');
    azi_pl (143,:) = [];
end

trend = azi_pl(:,1); plunge = azi_pl(:,2);
trend(143) = trend(132);
plunge(143) = plunge(132);
eval(sprintf('addpath(''E:\milne_data\Portsmouth_tests\EB_%.0f_40'')',k));

summary = eval(sprintf('importdata(''EB_%.0f_40_summary.txt'')',k));
summary (:,11) = summary (:,11)*100;
velocity = eval(sprintf('importdata(''EB_%.0f_40_velocity.txt'')',k));
velocity = velocity/1000;

max_vel = round(max (max(velocity)));
min_vel = round (min (min(velocity)));

I = input ('choose 6 surveys to plot:\n');
velocity = velocity (I,:);
summary = summary (I,:);
report = eval(sprintf('importdata(''mech_report_%.0f.txt'')',k));
fail_strain = report.data;
fail_stress = cell2mat(report.textdata(2,1));
fail_stress = str2num(fail_stress(1:end-3));

figure('Position',[1 1 1200 600])

for i = 1:length (velocity (:,1))
    if summary (i,11)<0
        summary (i,11) = 0;
    end
    v = velocity(i,:);
    % get rid of the nan
    survey = [];
    for w=i:length(v)
        if isnan(v(w))
            else survey = [survey;trend(w),plunge(w),v(w)];
        end
    end
    t = survey(:,1); p = survey(:,2); v = survey(:,3);
    hold on; axis equal, axis off, box off, hidden off % equal scaling in x and y, no axes or box
    axis([-1 1 -1 1]) % sets scaling for x- and y-axes
    title (sprintf('\sigma_{d_i_f} = %.0f MPa - \epsilon_{%%} = %.2f',summary(i,12), summary(i,11)*100)); % plot x- and y-axes
    plot([-1 1],[0 0],[0 0],[-1 1],[k])
end

r = 1; % radius of reference circle
TH = linspace(0,2*pi,360); % polar angle, range 2 pi, 1/10 degree increment
[X,Y] = pol2cart(TH,r); % Cartesian coordinates of reference circle
plot(X,Y,k') % plot reference circle

263
s2 = sqrt(2);

% schmidt -- Script for plotting a Schmidt net to plot points, first calculate theta = pi*(90-azimuth)/180, then rho = sqrt(2)*sin(pi*(90-dip)/360), and finally the components xp = rho*cos(theta) and yp = rho*cos(theta)
or use snetplot to plot from a file
theta = pi*(90-t)/180;
rho = sqrt(2)*sin(pi*(90-p)/360);
x = rho.*cos(theta);
y = rho.*sin(theta);
% plot
plot(x,y,’kd’)
LA = linspace(0,pi,180); % longitudes for pts on parallels
for n = 1:8 % loop to plot parallels at 10 deg increments
    ph = n*(10*pi/180); % latitude of particular parallel
cph = cos(ph); sph = sin(ph);
    X = r*cph*cos(LA); % x-coords of pts on parallel
    Y = r*sph*ones(size(X)); % y-coords of pts on parallel
PHP = acos(sqrt(X.^2 + Y.^2)/r); % plunge of pts on parallel
ALP = (pi/2) - atan2(Y,X); % azimuth of pts on parallel
XX = (r*s2)*sin(pi/4 - PHP/2).*sin(ALP); % x-coords of projected pts
YY = (r*s2)*sin(pi/4 - PHP/2).*cos(ALP); % y-coords of projected pts
plot(XX,YY,’k’,XX,’k’) % plot two sets of parallels
end
PH = linspace(-pi/2,pi/2,180); % latitudes for pts on meridians
for n = 1:8 % loop to plot meridians at 10 degree increments
    la = n*(10*pi/180); % longitude of particular meridian
    cla = cos(la);
    X = r*cos(PH).*cla; % x-coords of pts on meridian
    Y = r*sin(PH); % y-coords of pts on meridian
PHP = acos(sqrt(X.^2 + Y.^2)/r); % plunge of pts on meridian
ALP = (pi/2) - atan2(Y,X); % azimuth of pts on meridian
XX = (r*s2)*sin(pi/4 - PHP/2).*sin(ALP); % x-coords of projected pts
YY = (r*s2)*sin(pi/4 - PHP/2).*cos(ALP); % y-coords of projected pts
plot(XX,YY,’k’,XX,’k’) % plot two sets of meridians
end
% do the interp
xv=x; yv=y; zv=v;

%stem3(xv,yv,zv)
    xi=linspace(-1,1,1000); yi=xi;
    [Xi,Yi,Zi]=griddata(xv,yv,zv,xi,yi,’v4’);
% delete point outside hemisphere
for w =1:1000
    for j=1:1000
        if Xi(w,j)^2 + Yi(w,j)^2>1
            Zi(w,j)=NaN;
        end
    end
end
% fix overlay
x2 = Zi;
h=surf(Xi,Yi,Zi);
hold on
z = get(h,’ZData’);
set(h,’ZData’,z-10);
pcolor(Xi,Yi,Zi)
shading interp
% set fixed colormap for velocity only
caxis ([3 6.5]);
h2=colorbar;
colormap (jet)
% set title for velocity only
title (h2,'P-vel (km/s)');
end
% save picture
eval (sprintf('saveas (gcf,''EB_%.0f_40_survey.tiff'')',k));
eval (sprintf('saveas (gcf,''EB_%.0f_40_survey'')',k));

MATLAB CODE 2
% EXTRACT THE HIT COUNT FROM A SRM FILE
clear
% import file with voltage thresholds
thresholds = importdata('D:\richter_thresholds_venting.xlsx');
thresholds = thresholds.data;
thresholds = thresholds(:,2:end);
% import file with SRM file info
info = importdata('D:\SRM_venting_info.txt');
k = input('Type in test number:\n');
tic
a = info(k-13)+2;
if a<10
    a = eval(sprintf('num2str(''0%.0f'');',a));
else a = num2str(a);
end
% import wve file and get length of SRM file
if k == 16
    eval(sprintf('data = importdata (''D:\Portsmouth_tests\EB_%.0f_40\Venting\RT_auto\rct-uop-090414.data.000%s.wve'');',k,a))
else eval(sprintf('data = importdata (''D:\Portsmouth_tests\EB_%.0f_40\Venting\rct-uop-090414.data.000%s'');',k,a))
end
time_start = data.data(10);
time_finish = data.data(11);
hhmmss_start = num2str(time_start);
hhmmss_finish = num2str(time_finish);
start_time_ss = str2double(hhmmss_start(1:2))*3600 + str2double(hhmmss_start(3:4))*60 +
str2double(hhmmss_start(5:end));
length_srm = (str2double(hhmmss_finish(1:2))*3600 + str2double(hhmmss_finish(3:4))*60 +
str2double(hhmmss_finish(5:end))))-
    (str2double(hhmmss_start(1:2))*3600 + str2double(hhmmss_start(3:4))*60 +
str2double(hhmmss_start(5:end)));
last_step = (floor(length_srm*10) -1.000001)/10;
rest = ((length_srm*10 - floor(length_srm*10))/10)+0.000001;
% Read 16 bit, 4 channel SRM data
% import Master Richter first 0.1s
if k == 16
    eval(sprintf('datafile=''D:\Portsmouth_tests\EB_%.0f_40\Venting\RT_auto\rct-uop-090414.data.000%s'';',k,a))
else eval(sprintf('datafile=''D:\Portsmouth_tests\EB_%.0f_40\Venting\rct-uop-090414.data.000%s'';',k,a))
end
namefile = 'data';
if k == 20
fname_bin1=char([datafile '.h5']);
else fname_bin1=char([datafile '.srm']);
end
fname_inx=char([datafile '.wve']);
Fs=10E6; % sampling frequency
L = 1E6; % time window
% open binary data file, read data
fid = fopen(fname_bin1,'r');
secs=0; %number of seconds to skip
skipNbytes=secs*Fs*2^4; %4 bytes per int32, 2 bytes per int16, 4 samples per time step (multiplexed)
fseek(fid,skipNbytes, 'bof'); % skip secs seconds
data = fread(fid,[4,L],'uint16'); %1E6=0.1s
data = (((data/2^16)*5)-2.5)*2;'
fclose(fid);
% apply bandpass Butterworth filter of order 10
HPfilter = 1e4;
LPfilter = 1e6;
d = fdesign.bandpass('N,F3dB1,F3dB2',10, HPfilter,LPfilter,Fs);
Hd = design(d,'butter');
data1 = filter (Hd,data);
% Import Slave Richter first 0.1s
if k == 16
   eval(sprintf('datafile=''D:\Portsmouth_tests\EB_%.0f_40\Venting_RT_auto\rct-uop-180113.data.000%s'';',k,a))
else eval(sprintf('datafile=''D:\Portsmouth_tests\EB_%.0f_40\Venting\rct-uop-180113.data.000%s'';',k,a))
end
namefile = 'data';
if k == 20
   fname_bin2=char([datafile '.h5']);
else fname_bin2=char([datafile '.srm']);
end
fname_inx=char([datafile '.wve']);
Fs=10E6; % sampling frequency
L = 1E6; % time window
% open binary data file, read data
fid = fopen(fname_bin2,'r');
secs=0; %number of seconds to skip
skipNbytes=secs*Fs*2^4; %4 bytes per int32, 2 bytes per int16, 4 samples per time step (multiplexed)
fseek(fid,skipNbytes, 'bof'); % skip secs seconds
data = fread(fid,[4,L],'uint16'); %1E6=0.1s
data = (((data/2^16)*5)-2.5)*2;'
fclose(fid);
% apply bandpass Butterworth filter of order 10
data2 = filter (Hd,data);
% get absolute values of the waveform
data = [abs(data1),abs(data2)];
% set threshold for each channel
threshold = thresholds(k-13,1:8);
% find points of relative maxima on each channel using derivative concept
(by Eduardo Rossi, University of Geneva)
derivative_1 = data(2:end-1,:)-data(1:end-2,:);
derivative_2 = data(3:end,:)-data(2:end-1,:);
dev_1_pos = derivative_1>0;
dev_2_pos = derivative_2<0;
clear derivative_1
clear derivative_2
matrix_signs = dev_1_pos+dev_2_pos;
clear dev_1_pos
clear dev_2_pos
% Find elements with matrix_signs == 2 (meaning local maximum)
[elementi_i, elementi_j] = find(matrix_signs==2);
clear matrix_signs
% Correct indexes because coming from submatrixes
hits1 = zeros(1, 8); % 8 channels
ampl = linspace(0.01,2.3,20); % CREATE A 3D MATRIX, WITH SECONDS, CLUSTERS AND CHANNELS AS DIMENSIONS
for i=1:length(elementi_j)
    if data(elementi_i(i) + 1, elementi_j(i)) >= threshold(elementi_j(i));
        hits1(elementi_j(i)) = hits1(elementi_j(i))+1;
    end
end
time_step = 0.0999999:0.1:last_step;
for secs = time_step; % add 0.1 minus 0.1us each time
%import Master Richter second 0.1s
L = 1E6+1; % time window (plus last point of the previous window time)
% open binary data file, read data
fid = fopen(fname_bin1,'r');
secs=secs; %number of seconds to skip
skipNbytes=secs*Fs*2*4; %4 bytes per int32, 2 bytes per int16, 4samples per time step (multiplexed)
% apply bandpass Butterworth filter of order 100
data1 = filter (Hd,data);
% Import Slave Richter second 0.1s
L = 1E6+1; % time window
% open binary data file, read data
fid = fopen(fname_bin2,'r');
secs=secs; %number of seconds to skip
skipNbytes=secs*Fs*2*4; %4 bytes per int32, 2 bytes per int16, 4samples per time step (multiplexed)
% apply bandpass Butterworth filter of order 100
data2 = filter (Hd,data);
data = [abs(data1),abs(data2)]; %get absolute values of the waveform
derivative_1 = data(2:end-1, :)-data(1:end-2,:);
derivative_2 = data(3:end, :)-data(2:end-1,:);
dev_1_pos = derivative_1>0;
dev_2_pos = derivative_2<0;
clear derivative_1
clear derivative_2
matrix_signs = dev_1_pos+dev_2_pos;
clear dev_1_pos
clear dev_2_pos
[elementi_i, elementi_j] = find(matrix_signs==2);
clear matrix_signs
hits1 = zeros(1, 8);
for i=1:length(elementi_j)
    if data(elementi_i(i) + 1, elementi_j(i)) >= threshold(elementi_j(i));
    hits1(1,elementi_j(i)) = hits1(1,elementi_j(i))+1;
    end
end
hits = [hits;hits1];
end
%
get hits of the incomplete 0.1 second
%import Master Richter rest time
L = rest*10E6; % time window (plus last point of the previous window time)
% open binary data file, read data
fid = fopen(fname_bin1,'r');
seccs=length_srm - rest; %number of seconds to skip
skipNbytes=secs*Fs*2*4; %4 bytes per int32, 2 bytes per int16, 4 samples per time step (multiplexed)
fseek(fid, skipNbytes, 'bof'); % skip secs seconds
data = fread(fid,[4,L], 'uint16'); %1E6=0.1s
fclose(fid);
%apply bandpass Butterworth filter of order 100
data1 = filter (Hd,data);
% Import Slave Richter
rest time
L = rest*10E6; % time window
% open binary data file, read data
fid = fopen(fname_bin2,'r');
seccs=length_srm - rest; %number of seconds to skip
skipNbytes=secs*Fs*2*4; %4 bytes per int32, 2 bytes per int16, 4 samples per time step (multiplexed)
fseek(fid, skipNbytes, 'bof'); % skip secs seconds
data = fread(fid,[4,L], 'uint16'); %1E6=0.1s
fclose(fid);
%apply bandpass Butterworth filter of order 100
data2 = filter (Hd,data);
data = [abs(data1),abs(data2)]; %get absolute values of the waveform
derivative_1 = data(2:end-1,:)-data(1:end-2,:);
derivative_2 = data(3:end,:)-data(2:end-1,:);
dev_1_pos = derivative_1>0;
dev_2_pos = derivative_2<0;
clear derivative_1
clear derivative_2
matrix_signs = dev_1_pos+dev_2_pos;
clear dev_1_pos
clear dev_2_pos
[elementi_i, elementi_j] = find(matrix_signs==2);
clear matrix_signs
if k == 15 || k == 37
    hits1 = zeros(1, 12);
else
    hits1 = zeros(1, 8);
end
for i = 1:length(elementi_j)
    if data(elementi_i(i) + 1, elementi_j(i)) >= threshold(elementi_j(i));
        hits1(1,elementi_j(i)) = hits1(1,elementi_j(i))+1;
    end
end
hits = [hits;hits1];
hits = [zeros(length(hits(:,1)),1),1 hits];
for i = 1:length (hits(:,1))
    hits(i,1) = (start_time_ss*10+i)/10;
end
% get hit rate count per second
hits_second =[];
for i=1:floor(length_srm) % complete seconds in the file
    hits_second = [hits_second;sum(hits(i*10-9:i*10,2:end))];
end
hits_second = [hits_second; sum(hits(floor(length_srm)*10+1:end,2:end))];
for i=1:floor(length_srm)+1
    hits_second(i,1) = i+start_time_ss;
end
eval(sprintf('save(''EB_%.0f_40_hitsxstenth_%s.txt'',''hits'',''ascii'')',k,datafile(end-1:end)))
eval(sprintf('save(''EB_%.0f_40_hitsxsecond_%s.txt'',''hits_second'',''ascii'')',k,a))
toc

MATLAB CODE 3
% GET B-VALUE BY USING FORMULAS TAKEN FROM UTSU 1965 (Shi and Bolt, 1982) and WIEMER AND WYSS 2000

clear
k = input('Type in test number:
');
% define AE unit
STR1 = 'richter';
% import PED file with peaks
para = eval(sprintf('importdata (''E:\Richter_data\Portsmouth_tests\Export_files\EB_%.0f_40_Apeaks_richter_fail.ped'','' ',3)',k));
event_t = para.textdata(4:end,4);
para = para.data;
% get event time
event_time = [];
for i = 1:length(event_t)
    str = cell2mat(event_t(i));
    h = str2num(str(1:2));
    m = str2num(str(4:5));
    s = str2num(str(7:8));
    event_time = [event_time;h*3600+m*60+s i];
end
% get max amplitude of each event
if strcmp(STR1,'richter')
    if k == 37||k==15
        peaks = para(:,2:2:24);
        peaks(:,13) = max(abs(peaks(:,1:12))')';
    elseif k == 27
        peaks = para(:,2:2:22);
        peaks(:,13) = max(abs(peaks(:,1:11))')';
    end
else
    if k == 16
        peaks = para(:,2:2:20);
        peaks(:,13) = max(abs(peaks(:,1:10))')';
    elseif k == 27
        peaks = para(:,2:2:22);
        peaks(:,13) = max(abs(peaks(:,1:11))')';
    else
        peaks = para(:,2:2:24);
        peaks(:,13) = max(abs(peaks(:,1:12))')';
    end
end
% get max amplitude of all channels and convert in logarithmic scale
MIN = min(peaks(:,13));
MAX = max(peaks(:,13));
magn_total = floor(log10(MIN)):0.025:ceil(log10(MAX));
occur_total = zeros(length(magn_total),1);
for j =1:length(magn_total)
    for i = 1:length(peaks(:,13))
        if log10(peaks(i,13))>=magn_total(j)
            occur_total(j) = occur_total(j) + 1;
        end
    end
end

occur_total = log10(occur_total);
peaks(:,13) = log10(peaks(:,13));

% find magnitude threshold
count = 2;
grad = abs(occur_total(count) - occur_total(count-1));
while grad<0.01
    count = count+1;
    grad = abs(occur_total(count) - occur_total(count-1));
end
Mmin_total = magn_total(count);

% define length of the bins according to dataset size
if length (peaks(:,1))<1500
    NAMPS = [500 750];
elseif length (peaks(:,1))>1500&&length (peaks(:,1))<2000
    NAMPS = [500 750 1000];
elseif length (peaks(:,1))>2000&&length (peaks(:,1))<2500
    NAMPS = [500 750 1000 1250];
elseif length (peaks(:,1))>2500&&length (peaks(:,1))<3000
    NAMPS = [500 750 1000 1250 1500];
elseif length (peaks(:,1))>3000&&length (peaks(:,1))<3500
    NAMPS = 500:250:1750;
else NAMPS = 500:250:2000;
end
for Namps = NAMPS;
tic
    bin = floor(1+(length (peaks(:,13))-Namps)/(Namps/10));
b_val = [];
    H = waitbar(0,'Please wait...', 'Name', sprintf('EB%.0f, progress bar for Namps = %.0f,k,Namps'));
    for n = 1:bin
        % Report current estimate in the waitbar's message field
        magn_bin = [];
        for i = n*Namps/10-(Namps/10-1):(Namps+n*Namps/10)-Namps/10
            magn_bin = [magn_bin;peaks(i,13)];
        end
        time_ss_b = event_time((Namps+n*Namps/10)-Namps/10);
        % get frequeny of events in the bin
        MIN = min(magn_bin);
        MAX = max(magn_bin);
        magn = MIN:0.1:MAX;
        occur = zeros(length(magn),1);
        for j =1:length(magn)
            for i = 1:Namps
                if magn_bin(i)>=magn(j)
                    occur(j) = occur(j) + 1;
                end
            end
        end
        %
    end
end
end
occur = log10(occur);
occur = occur(find(occur>=0));
% find magnitude threshold in the bin
count1 = 2;
grad = abs(occur(count1) - occur(count1-1));
while grad<0.01 % arbitrary value
count1 = count1+1;
grad = abs(occur(count1) - occur(count1-1));
end
Mmin = magn(count1);
magn_thr = [];
for i = 1:Namps
    if magn_bin(i)<Mmin
        else magn_thr = [magn_thr;magn_bin(i)];
    end
end
% calculate b-value using Aki's maximum likelihood method
bval = 0.43/(mean(magn_thr)-Mmin);
% get a-value minimizing the error (Wiemer & Wyss, 2000) for a set of lines with a b-value slope
aval = [];
for c = -occur_total(1):0.001:occur_total(1)
sint = -magn*bval + c;
    R = 100 - (100*(sum(abs(occur-sint))/sum(occur)));
    aval = [aval; c R];
end
[M I] = max(aval(:,2));
% remove high magnitudes if R < 95
count2 = 0;
while M < 95
    count2 = count2+1;
    % Report current estimate in the waitbar's message field
    waitbar(count2/length(magn),H,sprintf('Bin: %.0f of %.0f, Number of removed amplitudes: %.0f',n,bin,count2))
    magn_thr2 = [];
    for i = 1:length(magn_thr)
        if magn_thr(i)>magn(end-count2)
            else magn_thr2 = [magn_thr2; magn_thr(i)];
        end
    end
    bval = 0.43/(mean(magn_thr2)-Mmin);
    aval = [];
    for c = -2*occur_total(1):0.001:2*occur_total(1)
        sint = -magn(1:end-count2)*bval + c;
        R = 100 - (100*(sum(abs(occur(1:end-count2)-sint))/sum(occur(1:end-count2))));
        aval = [aval; c R];
    end
    [M I] = max(aval(:,2));
end
sint = -magn*bval + aval(I,1);
if count2==0
    magn_thr2 = magn_thr;
end
% calculates the standard error of meanAmp then calculates the standard error of b from this
Adiffs2=(magn_thr2-mean(magn_thr2)).^2;
S2A = sum(Adiffs2)/(length(magn_th2)*(length(magn_th2)-1));
stdEmA = S2A^0.5;
stdEb = 2.3*(bval^2)*stdEmA;
stdEB_per = stdEb/bval*100;

% b_val file is composed by: time (seconds), b_val, correlation
% coefficient, completeness threshold, highest binned amplitude used, number of removed high binned
% amplitude, number or binned amplitude used to get the b-value, standard %
% error of b-value as calculated above
b_val = [b_val; time_ss_b bval M Mmin magn(end-count2) count2 length(magn) stdEB_per];
end
eval(sprintf('save ('"b_value_%.0f_%s_MLE_Aki_%.0f.txt"","b_val","-ascii"'),k,STR1,Namps))
% import hit_file to zero the b-value time
hit_data =
eval(sprintf('importdata("E:\Milne_data\Portsmouth_tests\hits_files\EB_%.0f_hits_milne.txt"'),k));
if k == 14
    hit_data(:,1) = hit_data(:,1)+1400;
elseif k == 16
    hit_data(:,1) = hit_data(:,1)+2800;
elseif k == 32
    hit_data(:,1) = hit_data(:,1)+2600;
end
b_val(:,1) = b_val(:,1) - hit_data(1,1);
%import mech data to plot b-value and stress over time
data =
eval(sprintf('importdata("E:\Milne_data\Portsmouth_tests\CORRECTED_MECH_DATA\data_mech_%.0f.txt"'),k));
data(:,1) = data(:,1) - hit_data(1,1);  
% plots
[AX,H1,H2] = plotyy(data(:,1),data(:,18)-data(:,14),b_val(:,1),b_val(:,2));
if k == 34
    xlim(AX(1),[0 2200])
    xlim(AX(2),[0 2200])
else
    xlim(AX(1),[0 1800])
    xlim(AX(2),[0 1800])
end
ylim(AX(1),[0 500])
set(get(AX(1),'Xlabel'),'String','Time (s)')
set(get(AX(1),'Ylabel'),'String','Differential stress (MPa)')
set(get(AX(2),'Ylabel'),'String','b-value')
set(AX(1),'YTick',0:50:500)
box off
H2.Marker = 'o';
H1.LineWidth = 2;
H2.LineWidth = 1;
H2.Color = 'k';
legend('Diff Stress','b-value','Location','northwest')
eval(sprintf('saveas (gcf,"EB_%.0f_b_value_%s_MLE_Aki_%.0f.tiff"'),k,STR1,Namps))
eval(sprintf('saveas (gcf,"EB_%.0f_b_value_%s_MLE_Aki_%.0f"'),k,STR1,Namps))
% plot number of removed binned amplitude vs stress and time
[AX,H1,H2] = plotyy(data(:,1),data(:,18)-data(:,14),b_val(:,1),b_val(:,6));
if k == 34
    xlim(AX(1),[0 2200])
    xlim(AX(2),[0 2200])
else
    xlim(AX(1),[0 1800])
    xlim(AX(2),[0 1800])
end
ylim(AX(1),[0 500])
ylim(AX(2),[0 40])
set(get(AX(1),'Xlabel'),'String','Time (s)')
set(get(AX(1),'Ylabel'),'String','Differential stress (MPa)')
set(get(AX(2),'Ylabel'),'String','No. of removed amplitude')
set(AX(1),'YTick',0:50:500)
set(AX(2),'YTick',0:10:40)
box off
H2.Marker = 'o';
H1.LineWidth = 2;
H2.LineWidth = 1;
H2.Color = 'k';
legend('Diff Stress','removed amplitude','Location','northwest')
eval(sprintf('saveas (gcf,''EB_%.0f_del_ampl_%s_MLE_Aki_%.0f.tiff'')',k,STR1,Namps))
eval(sprintf('saveas (gcf,''EB_%.0f_del_ampl_%s_MLE_Aki_%.0f'')',k,STR1,Namps))
% plot histogram of amplitude lower thresholds
histogram (10.^b_val(:,4))
ylabel('Occurrences')
xlabel('Completness (amplitude)')
eval(sprintf('saveas (gcf,''EB_%.0f_completness_%s_MLE_Aki_%.0f.tiff'')',k,STR1,Na
mps))
eval(sprintf('saveas (gcf,''EB_%.0f_completness_%s_MLE_Aki_%.0f'')',k,STR1,Namps))
% plot histogram of amplitude upper thresholds
histogram (10.^b_val(:,5))
ylabel('Occurrences')
xlabel('High amplitude removed')
eval(sprintf('saveas (gcf,''EB_%.0f_high_ampl_%s_MLE_Aki_%.0f.tiff'')',k,STR1,Namps))
eval(sprintf('saveas (gcf,''EB_%.0f_high_ampl_%s_MLE_Aki_%.0f'')',k,STR1,Namps))
toc
delete(H)
end

MATLAB CODE 4

% GET FWHM AND CALCULATE QUALITY FACTOR
addpath('E:\Richter_data\Portsmouth_tests\Freq_venting\Files')
k = input('Type in test number:\n');
fft_low = eval(sprintf('importdata(''fft_low_%.0f.txt'')',k));
dom_freq = eval(sprintf('importdata(''dom_freq_%.0f.txt'')',k));
f = importdata('frequency_28.txt'); % all frequency file are the same
X = eval(sprintf('importdata(''E:\Milne_data\Portsmouth_tests\EB_%.0f_40\venting_time_%.0f.txt'')',k,k));
X = cell2mat(X);
X = str2num(X(44:end-2));
dom_freq(:,end) = dom_freq(:,end) - X;
[I1 m1] = max(find(f<1e4));
[I2 m2] = min(find(f>3e5)); % take a longer window to get resample data without edge problems
f = f(I1:I2);
fft_low = fft_low(I1:I2,:);
f = resample(f,100,1);
fft_low = resample(fft_low,100,1);
%[I1 m1] = max(find(f<4e4));
%[I2 m2] = min(find(f>2e5));
I1 = 310;
I2 = 1950;
f = f(I1:I2);
fft_low = fft_low(I1:I2,:);
[I3 m3] = min(find(dom_freq(:,end)>eps));
if isempty(I3)
    I3 = 1;
end
[I4 m4] = max(find(dom_freq(:,end)<120));
if isempty (I4)
    I4 = length(dom_freq(:,1));
end
fft_low = fft_low(:,I3:I4);
dom_freq = dom_freq(I3:I4,:);
range = [];
Q = [];
for i =1:length(fft_low(1,:))
    fft_low(:,i) = fft_low(:,i)/max(fft_low(:,i));
x1 = find(fft_low(:,i)==1);
x2=x1-10;
while fft_low(x2,i)>=0.5
    if x2 > 2
        x2 = x2-1;
    else
        fft_low(x2,i)=0.499;
    end
end
x3=x1+10;
while fft_low(x3,i)>=0.5
    if x3 < length(f)-1
        x3 = x3+1;
    else
        fft_low(x3,i)=0.499;
    end
end
range = [range; f(x2) f(x3)];
% calculate Q
Q = [Q dom_freq(i,1)/(f(x3)-f(x2))];
end
eval(sprintf('save(''range50_%.0f.txt'',''range'',''ascii''),k))

MATLAB CODE

% CROSS-CORRELATION OF EVENT WITH BRIDGE METHOD, % USING COEFFICIENT THRESHOLD OF 0.8 AND % WINDOW LENGTH OF 30 POINTS AFTER ARRIVAL

clear
addpath('E:\Richter_data\Portsmouth_tests\Freq_venting\Files')
k = input('Type in test number:\n');
onsets = eval(sprintf('importdata(''onsets%.0f_chosen.txt''),k));
all_low  = eval(sprintf('importdata(''all_low_%.0f_chosen.txt''),k));
f = importdata('frequency_28.txt'); % all frequency file are the same
if ne(k,[28 34])
    dom_freq = eval(sprintf('importdata(''dom_freq_%.0f.txt''),k));
else
    dom_freq = eval(sprintf('importdata(''dom_freq_%.0f_chosen.txt''),k));
end
if k == 28
    all_low = all_low(:,1:31);
end
% zero the time to venting instant
X =
eval(sprintf('importdata(''E:\Milne_data\Portsmouth_tests\EB_%.0f_40venting_time_%.0f.txt''),k,k));
X = cell2mat(X);
X = str2num(X(44:end-2));

% normalize waveforms
for i = 1:length(all_low(1,:))
    all_low(:,i) = all_low(:,i)/max(all_low(:,i));
end

% discard events before release and after 120 s
[I3 m3] = min(find(dom_freq(:,end)>eps));
if isempty(I3)
    I3 = 1;
end
[I4 m4] = max(find(dom_freq(:,end)<120));
if isempty(I4)
    I4 = length(dom_freq(:,1));
end

dom_freq = dom_freq(I3:I4,:);
onsets = onsets(I3:I4,:);
all_low = all_low(:,I3:I4);
total = length(all_low(1,:));

% create matrixes of equal length of FFT

% correlate events
n_families = [];
stat_cross_corr = [];
min_corrs_tt = [];
mean_corrs_tt = [];
for L = [100:100:500];
    mat_events = [];
    for i = 1:length(onsets(:,1))
        mat_events = [mat_events all_low(onsets(i,1):onsets(i,1)+L,i)];
    end
end
% correlate
[row column] = size(mat_events);
events_corr = zeros(row, column);
for i=1:column
    for j=1:column
        if i==j
            events_corr(i,j) = 0;
        else
            c = mat_events(:,j);
            d = mat_events(:,i);
            a = corrcoef(c,d);
            events_corr(i,j) = abs(a(1,2));
        end
    end
end
% find master event (highest number of correlated events)
% correlate events by using different cross-correlation coefficients threshold
threshold = 0.8;
events_thr = [];
for i = 1:column
    J = find(events_corr(:,i)>=threshold);
    if isempty(J)
        else corr1 = [i*ones(length(J),1) J events_corr(J,i)];
        events_thr = [events_thr; corr1];
    end
end
if isempty(events_thr)
count = 1;
l_fam = 0;
min_corrs = 0;
mean_corrs = 0;
n_pair = 0;
else eval(sprintf('save (''events_corr_%.0f_%.0f.txt'',''events_corr'',''
-ascii'')',k,L))
a = find(events_thr(1:end,2) == events_thr(1,1));
b = find(events_thr(1:end,2) == events_thr(1,2));
c = find(events_thr(1:end,1) == events_thr(1,1));
d = find(events_thr(1:end,1) == events_thr(1,2));
fam1 = [events_thr(1,1); events_thr(1,2)];
fam1_a = [events_thr(1,1); events_thr(1,2)];
for num = 1:length(a)
    find(fam1_a==events_thr(a(num),1));
    if sum(ans) == 0
        fam1_a = [fam1_a; events_thr(a(num),1)];
    end
end
for num = 1:length(b)
    find(fam1_a==events_thr(b(num),1));
    if sum(ans) == 0
        fam1_a = [fam1_a; events_thr(b(num),1)];
    end
end
for num = 1:length(c)
    find(fam1_a==events_thr(c(num),2));
    if sum(ans) == 0
        fam1_a = [fam1_a; events_thr(c(num),2)];
    end
end
for num = 1:length(d)
    find(fam1_a==events_thr(d(num),2));
    if sum(ans) == 0
        fam1_a = [fam1_a; events_thr(d(num),2)];
    end
end
incr = length(fam1_a)-length(fam1);
while incr > 0
    fam1 = fam1_a;
c = [];
    for i=3:length(fam1_a)
        I = find(events_thr(:,1:2)==fam1_a(i));
        if isempty(I)
            else
                for k1=1:length(I)
                    if I(k1)>length(events_thr(:,1))
                        I(k1) = I(k1)-length(events_thr(:,1));
                    end
                end
                c = [c; I];
            end
        for i=1:length(c)
            c(i) = a;
            if sum(ans)>0
                else pair = events_thr(c(i),1:2);
            end
        end
end
end

for k1 = 1:2
    pair(k1) == fam1_a;
    if sum(ans) > 0
        else fam1_a = [fam1_a; pair(k1)];
    end
    end
end
incr = length(fam1_a) - length(fam1);
end
count = 2;
eval(sprintf('events_thr%.0f = [];', count))
for i = 1:length(fam1)
    events_thr(:,1) == fam1(i);
    I = find(ans > 0);
    events_thr(I,:) = NaN;
end
for i = 1:length(events_thr(:,1))
    if isnan(events_thr(i,1))
        else eval(sprintf('events_thr%.0f = [events_thr%.0f; events_thr(i,:)];', count, count))
    end
end
while eval(sprintf('length(events_thr%.0f)>0', count))
    eval(sprintf('events_thr = events_thr%.0f;', count))
a = find(events_thr(2:end,2) == events_thr(1,1));
b = find(events_thr(2:end,2) == events_thr(1,2));
c = find(events_thr(1:end,1) == events_thr(1,1));
d = find(events_thr(1:end,1) == events_thr(1,2));
fam = [events_thr(1,1); events_thr(1,2)];
fam_a = [events_thr(1,1); events_thr(1,2)];
for num = 1:length(a)
    find(fam_a == events_thr(a(num),1));
    if sum(ans) == 0
        fam_a = [fam_a; events_thr(a(num),1)];
    end
end
for num = 1:length(b)
    find(fam_a == events_thr(b(num),1));
    if sum(ans) == 0
        fam_a = [fam_a; events_thr(b(num),1)];
    end
end
for num = 1:length(c)
    find(fam_a == events_thr(c(num),2));
    if sum(ans) == 0
        fam_a = [fam_a; events_thr(c(num),2)];
    end
end
for num = 1:length(d)
    find(fam_a == events_thr(d(num),2));
    if sum(ans) == 0
        fam_a = [fam_a; events_thr(d(num),2)];
    end
end
incr = length(fam_a) - length(fam);
while incr > 0
fam = fam_a;
c = [];
for i=3:length(fam_a)
    I = find(events_thr(:,1:2)==fam_a(i));
    if isempty(I)
        else
            for k1=1:length(I)
                if I(k1)>length(events_thr(:,1))
                    I(k1) = I(k1)-length(events_thr(:,1));
                end
            end
            c = [c; I];
        end
    end
for i=1:length(c)
    c(i) == a;
    if sum(ans)>0
        else pair = events_thr(c(i),1:2);
            for k1 =1:2
                pair(k1) == fam_a;
                if sum(ans)>0
                    else fam_a = [fam_a; pair(k1)];
                end
            end
        end
    end
    incr = length(fam_a)-length(fam);
end
eval(sprintf('fam%.0f = fam;',count))
count = count + 1;
eval(sprintf('events_thr%.0f = [];',count))
for i = 1:length(fam)
    events_thr(:,1) == fam(i);
    I = find(ans>0);
    events_thr(I,:) = NaN;
end
for i =1:length(events_thr(:,1))
    if isnan(events_thr(i,1))
        else eval(sprintf('events_thr%.0f = [events_thr%.0f; events_thr(i,:)];',count,count))
    end
end
end
% get minimum correlation and average correlation for each family
min_corrs = [];
mean_corrs = [];
l_fam = [];
n_pair = [];
for n = 1:count-1
    eval(sprintf('fam = fam%.0f;',n))
l_fam = [l_fam; length(fam)];
min_corr = [];
sum_corr = 0;
n_pair_fam = [];
for i=1:length(fam)
    corrs = events_corr(fam(i),fam);
    [m I] = find(corrs>threshold);
n_pair_fam = [n_pair_fam; fam(i) length(I)];
corrs1 = [];
for k1=1:length(corrs)
    if corrs(k1) == 0
    else corrs1 = [corrs1; corrs(k1)];
    end
end

min_corr = [min_corr min(corrs1)];

[m I] = max(n_pair_fam(:,2));
n_pair = [n_pair; n_pair_fam(I,:)];
min_corrs = [min_corrs; min(min_corr)];
if length(fam) > 1
    mean_corrs = [mean_corrs; sum_corr/((length(fam)^2)/2)];
else mean_corrs = [mean_corrs; sum_corr];
end
end

end
correlated_events = [correlated_events sum(l_fam)];
stat_cross_corr = [stat_cross_corr; L 100*sum(l_fam)/total count-l max(l_fam)];
end
eval(sprintf('save(''EB%.0f_stat_cross_corr_bridge_.%.0f_%0f.txt'',''stat_cross_corr'',''-ascii''),k,threshold*100,L))
% plot correlated events VS window length
plot (stat_cross_corr(:,1)/10, stat_cross_corr(:,2),'k^','LineWidth',2)
xlabel('Window length (\mu s)', 'FontSize', 12)
ylabel('% correlated events', 'FontSize', 12)
xlim([5 55]); ylim([0 100])
set(gca,'YTick',0:20:100)
set(gca,'XTick',10:10:500)
set(gca,'FontSize',12)
eval(sprintf('str = ''%.0f events'';',total))
h = text(40,80,str);
h.FontSize = 12;
eval(sprintf('saveas(gcf,''EB%.0f_correvents_%.0f.tiff'')',k,threshold*100))
eval(sprintf('saveas(gcf,''EB%.0f_correvents_%.0f'')',k,threshold*100))
% plot number of families VS window length
plot (stat_cross_corr(:,1)/10, stat_cross_corr(:,3),'k^','LineWidth',2)
xlabel('No. of families', 'FontSize', 12)
xlim([5 55]); ylim([0 10])
set(gca,'YTick',0:1:10)
set(gca,'XTick',10:10:50)
set(gca,'FontSize',12)
eval(sprintf('str = ''%.0f events'';',total))
h = text(10,8,str);
h.FontSize = 12;
eval(sprintf('saveas(gcf,''EB%.0f_Nfam_%.0f.tiff'')',k,threshold*100))
eval(sprintf('saveas(gcf,''EB%.0f_Nfam_%.0f'')',k,threshold*100))
% plot size larger family VS window length
plot (stat_cross_corr(:,1)/10, stat_cross_corr(:,4),'k^','LineWidth',2)
xlabel('Size larger family', 'FontSize', 12)
xlim([5 55]); ylim([0 total+1])
set(gca,'XTick',10:10:50)
set(gca,'FontSize',12)
eval(sprintf('str = ''%.0f events'';',total))
h = text(40,total/2,str);
h.FontSize = 12;
eval(sprintf('saveas(gcf,''EB%.0f_Sizefam_%.0f.tiff'')',k,threshold*100))
eval(sprintf('saveas(gcf,''EB%.0f_Sizefam_%.0f'')',k,threshold*100))