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**EVOLUTION OF THE NORTH EAST RIFT ZONE OF TENERIFE, CANARY
ISLANDS: A MULTI-DISCIPLINARY APPROACH**

By Audray Delcamp



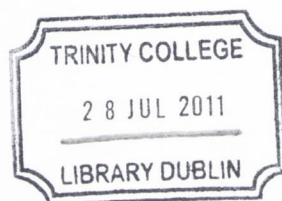
Submitted in accordance with the requirements for the degree of Doctor of Philosophy

The University of Dublin

Trinity College

School of Natural Sciences

January 2010



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To Anaëlle,

This manuscript has not been submitted as an exercise for a degree at this or other university.

It is my entirely own work. It includes, however, geochemical data that have been carried out jointly with F.M. Deegan.

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ABSTRACT

Tenerife Island, Canary archipelago, is at the centre of intense geological research and debate. The island is composed by three principal rift-zones that enclose major collapse structures. This study provides a multi-disciplinary approach to constrain better the development of the North-East Rift-Zone (NERZ) and its link with destabilisation periods manifested in lateral collapse events. Dyke orientation patterns in the field imply that flanks of the NERZ were slowly creeping over extended periods of time to accommodate the emplacement of an intrusive complex(es) and its associated dykes. Paleomagnetic data from NERZ mafic dykes and the associated host rocks of the upper part of the NERZ suggest they were rotated 26° clockwise about a vertical axis. This rotation is interpreted as the consequence of flank creeping that induced strike-slip movements into the edifice. The existence of intrusive complex(es) below the rift and its role in the deformation of the NERZ have been highlighted by AMS analyses and scaled analogue models. The development of the rift and the intrusive complex(es) are inherently linked and bring the system ultimately towards an unstable configuration and at that point, creeping cannot accommodate deformation any longer, and further intrusions will generate a collapse.

The architecture and development of the NERZ thus contrasts with the widely held beliefs that rift-zones are simple linear arrangement of fractures, dyke swarms, and volcanoes. The North-East rift-zone is a dynamic and changing environment, with several intense intrusive phases concordant or alternating with erosion periods and destabilisation events.

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INTRODUCTION

1.1. Tenerife and the Canary Archipelago

1.1.1. Geological setting

The Canary Archipelago comprises seven islands located off the Atlantic Ocean shore of Morocco, at a latitude of 28°N (**Fig.1**). From east to west, they are Lanzarote, Fuerteventura, Gran Canaria, Tenerife, La Gomera, La Palma and El Hierro. The oldest subaerial material has been found in Fuerteventura and corresponds to an Early Miocene age (20.6 Ma; *in Carracedo, 1999*). The archipelago lies on Jurassic oceanic crust locally thickened by intrusive activity (*Ye et al., 1999; Hansteen and Troll, 2003*).

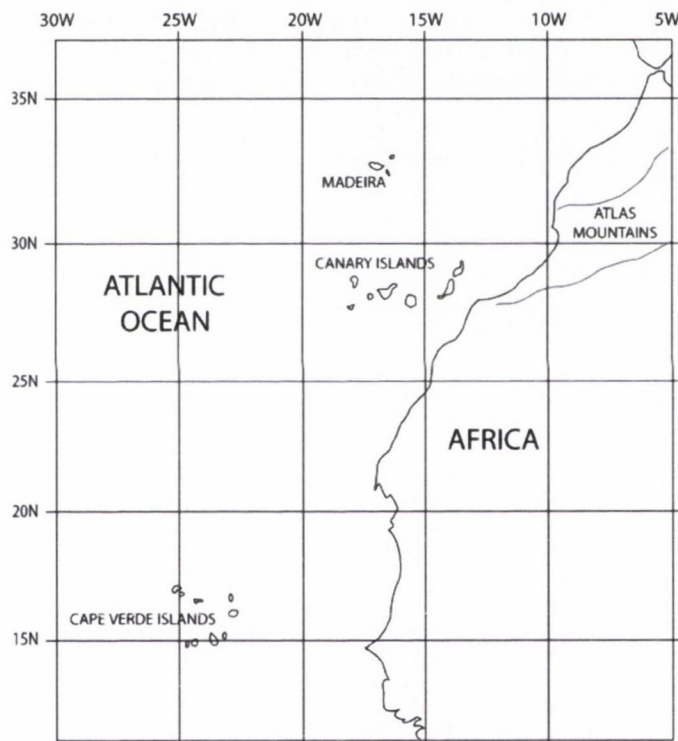


Figure 1: Map of the Canary Islands

The regional tectonics during Jurassic ocean break-up, and thus prior to the island's construction, were characterized by an NW-SE and NE-SW extensional regime in Morocco (*Ait Brahim et al., 2002; Fig.2*), forming the Mesozoic rift systems of the Middle and High Atlas (*e.g. Mattauer et al., 1977*). During the Early Cretaceous, the Moroccan region was affected by N-S extension. In the Late Cretaceous and during the

Tertiary and Quaternary, regional tectonics in Morocco were characterized by a tectonic inversion, with a compressive regime (*Ait Brahim et al., 2002*). A phase of E-W extensional regime occurred during the Late Miocene (**Fig.2**).

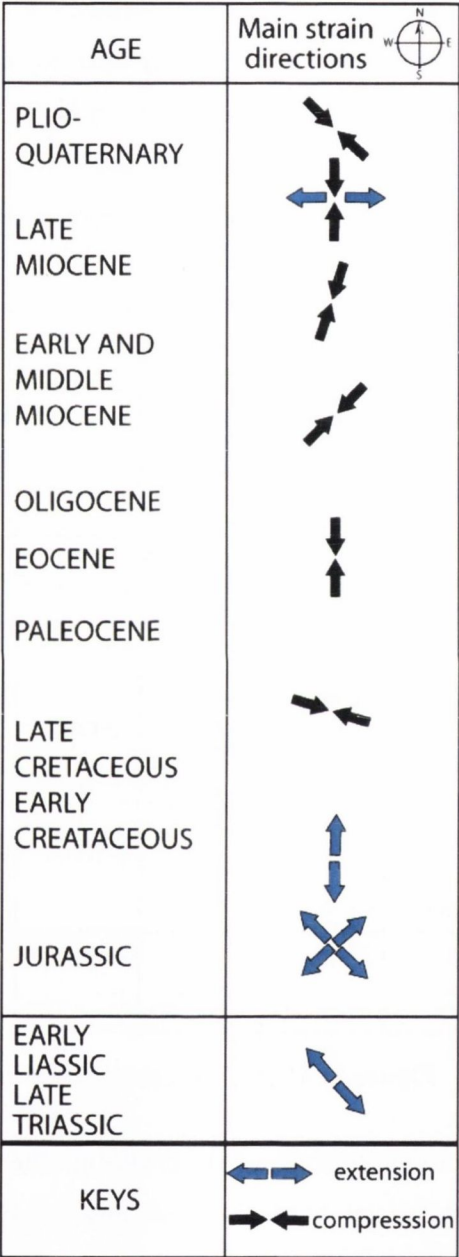


Figure 2: Main strain directions in Morocco from the Mesozoic to Quaternary (*From Ait Brahim et al., 2002*).

A seismic study from *Bosshard and McFarlane (1970)* suggested that the western islands are sited on oceanic crust while the eastern islands formed on the African passive margin, i.e. on a continental crust. These authors suggested that Gran Canaria is located at the continental/oceanic transition. However, more recent studies show that all the islands are based on thick and cold Jurassic oceanic crust (*Banda et al., 1981; Schmincke et al., 1998; Hoernle, 1998; Carbó et al., 2003; Hansteen and Troll, 2003*). The African plate is moving slowly with a rate estimated between 10 and 20mm/year (*Duncan, 1981; Morgan, 1983*).

1.1.2. The Origin

The origin of the archipelago has been the subject of much research and is still the subject of much debate. Recently, the different hypotheses concerning the origin of the Canary Islands have been summarized by *Anguita and Hernán (2000)* and *Garcia Cruz (2001)*. The main hypotheses are:

-**“The propagating fracture model”** where faults are running from the Atlas Mountains to the Canaries (*Beck and Lehner, 1974; Anguita and Hernán, 1975; Robertson and Stillman, 1979*). More recently, *Anguita and Hernán (2000)* suggest that this propagating fracture is a major sinistral strike-slip fault connected to the Canaries through aseismic deformation.

-**“The uplift of tectonic blocks theory”** developed mainly by *Araña and Ortiz (1986, 1991)* where the different islands are sited on lithospheric blocks that were uplifted at different rates in relation to a compressive stress field.

-**“The local rift”** where the Canary formation is related to the opening of a rift during Jurassic and Cretaceous time (*Fúster, 1975; Oyarzun et al., 1997*).

-**“The plume model”** first developed by *Wilson (1963)* for the Hawaiian volcanic chain, has been applied to the Canaries by *Morgan (1971, 1972), Schmincke (1973), Wilson (1973), Burke and Wilson (1976), Carracedo (1994), Carracedo et al. (1998), Dañobeitia and Cañales (2000)*. Building on this concept, *Hoernle and Schmincke (1993)* proposed a modification of this idea, the “blob-model”. Today, the plume or blob model seems to be the most widely accepted theory for the formation of the Canary chain. Indeed, the typical hot-spot correlation between the age of the islands and their

geographic positions has been resolved by recent dating that displayed a west to east continuous age progression (*Guillou et al., 2004*). Moreover, the erosion pattern of the Canary Islands is almost similar to the classic one exhibited by the Hawaiian hot-spot where the western volcanoes represent the shield stage and the eastern volcanoes the post-erosional stage, Tenerife being at the peak of its development (*Carracedo, 1999; Guillou et al., 2004*).

1.1.3. Tenerife evolution

Early constructive stages of Tenerife island are represented by three successive shield volcanoes during the Miocene-Pliocene (Roque Del Conde, Anaga and Teno); (**Fig.3**; *Fúster et al., 1968; Ancochea et al., 1990; Guillou et al., 2004; Carracedo et al., 2007*). After a 2-3 million year gap, renewed volcanism resulted in the central Cañadas volcano (*Ancochea et al., 1990; Martí et al., 1994; Bryan et al., 1998; Carracedo et al., 2007*). An explosive episode (~2-0.5Ma) was followed by a major lateral collapse that occurred at ~200ka, forming the Las Cañadas caldera (*Watts and Masson, 1995; Martí et al., 1997; Cantagrel et al., 1999; Martí and Gudmundsson, 2000; Carracedo et al., 2007*). Following this major event, the stratocone of Teide-Pico Viejo started to build up. Concurrent with the Cañadas volcano evolution and the Cañadas collapse, as well as the initial growth of the Teide-Pico Viejo complex, extensive activity also occurred along the rift zones (*Ancochea et al., 1990; Carracedo et al., 2007*). The present day approximate structure of Tenerife is presented in **figure 4**. Historic volcanic activity has been restricted to the NW rift zone (1704-1706 and 1909) apart from one eruption on the flank of Pico Viejo in the central part of the island (1798). A geochemical distribution in volcanic products has been observed, with mainly basaltic lavas along the rifts and more differentiated products with phonolite restricted to the caldera and the Teide Pico-Viejo complex (*Carracedo, 2004*).

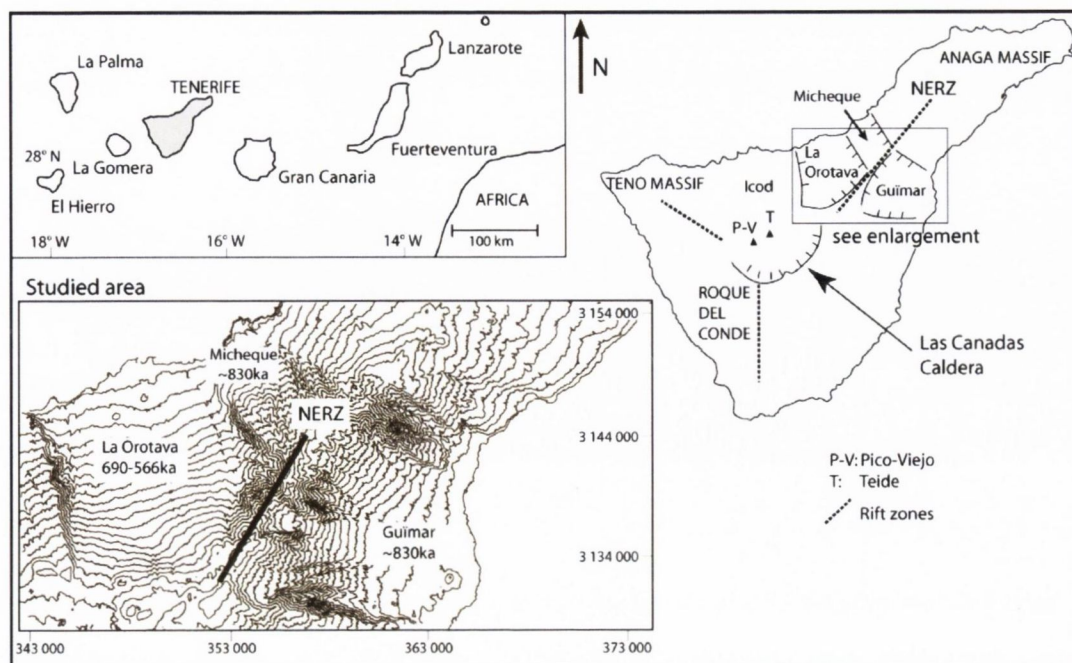


Figure 3: Structural map of Tenerife Island. The studied area is displayed on the bottom left box.

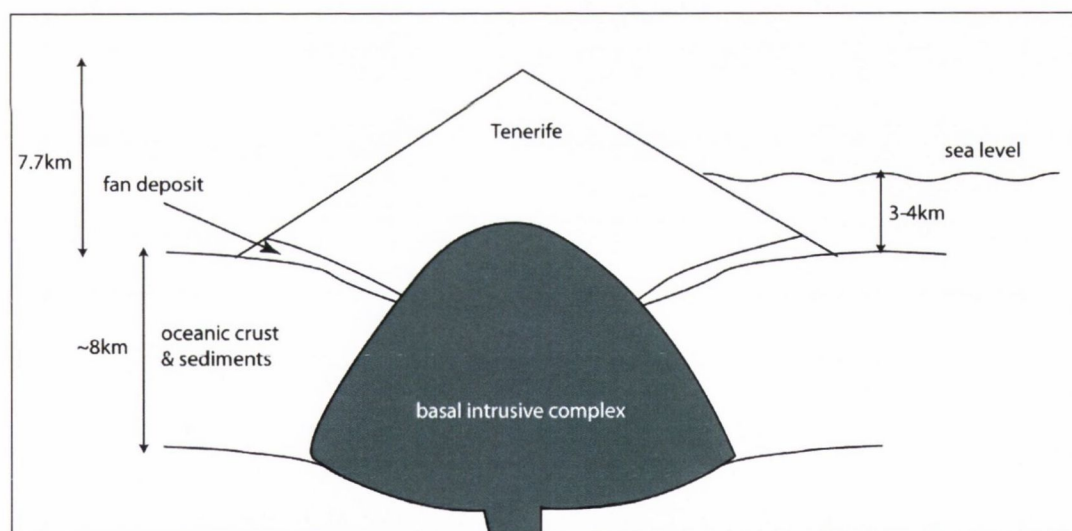


Figure 4: Schematic representation of Tenerife structure. For details see Araña et al., 2000; Cañales et al., 2000; Carracedo et al., 1998; Hürlimann et al., 2000; Schmincke and Segschneider, 1998; Watts, 1994; Watts et al., 1997; Weaver et al., 1998.

1.2.Aim

Tenerife Island is composed of three rift-zones organized in a triple arm pattern. A rift zone is usually described as a well-defined volcanic vent alignment characterized by

intense dyke emplacement along fractures at depth. Consequently, they are usually considered as long-lived, stable volcanic features. The first aim of this PhD is to unravel the mechanism(s) of rift zone formation and its temporal evolution.

Rift-zones usually enclose major collapse structures and a link between rift zone formation and collapse events has thus been postulated. Several models have been proposed to better assess this relationship. *Dieterich (1988)* suggested for the Hawaiian rift zones that intense dyke intrusion within the rift zone induces an outwards flank motion which can trigger collapse (**Fig.5A**). *Walter and Troll (2003, Fig.5B)* and *Walter et al. (2005)* proposed that flank instabilities favor magma trapping along the collapse scar. The second goal of this PhD is to bring new data concerning intrusive basal complexes and dyke emplacement to better assess rift-zone/collapse relationships.

The approach comprises a broad study on the rift zone formation and destabilization associated with lateral collapses and a detailed study on the mechanisms of dyke swarm structures inside the rift.

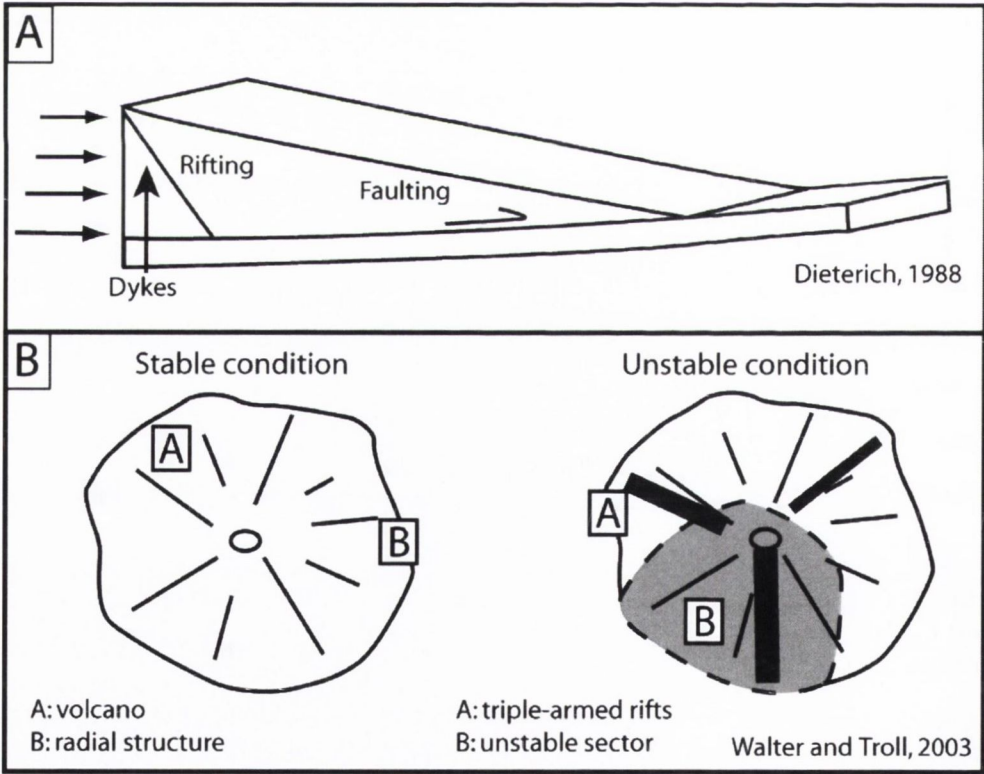


Figure 5: Simplified sketches showing the relationship between dyke emplacement and rift zone formation. **A:** Dieterich (1998) model; **B:** Walter and Troll model (2003).

1.3.Methods

Various methods were employed:

-Field work is a major component of this project and relies on a detailed and large scale study of the rift. The field area will be described later in the section 1.4 of the introduction chapter.

-Petrographic study has been carried out directly in the field and through microscopy. However, due to important variations in mineralogy and in the large mineral nature of the cryptocrystalline groundmass and the occurrence of aphyric dykes, **geochemical analysis** (major and trace element analysis) was needed. Note that this last method was not the main topic of the present PhD. This component was part of a collaboration with another PhD student (F.M. Deegan) who is focusing on petrology and geochemistry of these rocks. My role was to help collect and prepare the samples for XRF analysis and to use the data for broad classification purposes.

-Paleomagnetic analysis was used to group the dykes based on the paleomagnetic conditions that prevailed during their emplacement. This yields a relative time relationship between different dyke groups and can resolve between different batches of magma injection. To complement this a few dykes were dated using **Ar-Ar dating** method to obtain absolute ages.

-AMS (Anisotropy of Magnetic Susceptibility) provides a measure of the sub-microscopic alignment of the magnetic minerals in rocks. This method allows magma flow direction inside individual dyke intrusions to be obtained. Consequently, this technique increases our understanding of how magma migrates in, and how dykes feed rift-zones.

-Analogue modeling is carried out to constrain the range of potential mechanisms of rift and collapse formation, as well as dyke emplacement.

The PhD thesis is structured following the methods used. Each of them provides complementary information allowing a final model to be drawn up. The first chapter concerns the field work (including petrography and geochemistry), the second deals with the paleomagnetic analysis, the third with the AMS data and the fourth with the analogue models. The last chapter is the final conclusions, where all the obtained results using the

various methods are put together in a coherent model of rift zone formation and evolution and its links with collapse structures.

1.4. Field area

I focused on the dykes of the North East Rift Zone (NERZ), also referred as the Cumbre Dorsal (**Fig.3**). This area displays the fresher outcrops on Tenerife. Two major collapses, Guímar and La Orotava, are on the NERZ. La Orotava collapse is thought to have occurred less than 560ka years ago (*Ablay and Hürlimann, 2000*), while the Guímar collapse event is estimated at <83ka (*Ancochea et al., 1990*). However, *Carracedo et al. (2009)* defined a new age of 830ka for Guímar and bracket La Orotava between 690 and 566ka. These authors also identify an older collapse, the Micheque, directly to the north-east of La Orotava and constrained this event to about 830ka. More than 1000 tunnels (locally referred as “galerias”) have been dug to supply Tenerife’s population with water. We had the opportunity to visit and explore two of them in three days. Their very good exposure (100% exposure) allowed us to have a unique access to the interior of the island. We planned to visit more of them to have a better picture of the deeper levels of the rift-zone, but unfortunately, the day after our last visit in the galeria Rio de Igueste (February 2007), a tragic incident happened in one of the galerias. A group of people, unrelated to our team and ill prepared, entered a galeria and got lost in the tunnel system, only to find themselves in an unventilated part of the tunnels. Due to the lack of oxygen, several people in this group died, while others suffered permanent brain damage. Consequently, the local authorities have strictly denied access to all galerias and most of them have been walled-up.

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CHAPTER 1: FIELDWORK

Abstract

The volcanic evolution of Tenerife has been studied recently and the eruptive history of the NE and NW rift zones is now well constrained (*Carracedo et al., 2007 and 2009*). A systematic and detailed mapping study of dykes of the North East Rift-Zone (NERZ) is aimed to complement and to shed new light on the intrusive history related to the three successive collapses (1-Micheque, 2-Güímar and 3-La Orotava), which are located on either side of the rift. Flank creeping seems to be the major mechanism that accommodates the emplacement of the intrusive complex and its associated dykes. Flank spreading mechanisms allows the stabilisation of the rift until a critical point is reached, where further intrusion will generate collapse.

During the first stages of growth, the rift may be controlled by the regional stress field or pre-existing structures, but its future development depends on the local stress field created by the rift-zone evolution (structural maturation). Thus, and in contrast to a widely held misconception, volcanic rift-zones represent a dynamic and changing geological and morphological environment.

1. Introduction

The fieldwork was focused on the Tenerife North-East Rift-Zone (NERZ), orientated N040, and the three collapse scars running along the rift (Micheque, Güímar and La Orotava, cf. **introduction chapter; Fig.1A**). Recent work by *Carracedo et al. (2009)* showed that the rift is characterized by three main stages of activity separated by periods of quiescence: a Miocene (7203 ± 155 ka), a Late Pliocene (2710 ± 58 ka) and the latest stage during the Pleistocene. Each stage seems to be intense and relatively short (few hundred ka) compared with the lifetime of the entire volcano. During the Pleistocene, three collapse events affected the rift. At about 830ka, a sector on the north of the rift collapsed, forming the Micheque embayment. Almost contemporaneously, another collapse occurred on the south-east flank of the rift, producing the Güímar landslide embayment. The last collapse, La Orotava, affected the north of the rift and is estimated to have occurred between 690 and 566 ka. At present, the Micheque embayment is

completely obscured by post-collapse volcanism, while Güímar and La Orotava have been partially filled only.

As the volcanic evolution has been well-established by the recent work of *Carracedo et al. (2009)*, the present chapter concentrates on the structural evolution of the rift and its associated collapse events. The studied zone is divided into twelve sectors (labelled from A to L), where the dykes and structures were systematically recorded and described (**Fig.1B**). The galerias visited (i.e. water tunnels) of Aguas de San José (~1130 a.s.l.) and Rio de Igueste (RdI, ~400m a.s.l) are situated in sectors G and J. Following a brief description of petrography and geochemistry (major elements), I present the different characteristics of the dykes (morphology, structures, thickness etc). Then, the chapter is focused on structural data that help to understand the relationship between rift structures, flank creeping and collapse formation. A comparative study with active oceanic volcanoes elsewhere highlights different stages of rift-zone structural evolution related to basal intrusive complex and dyke emplacement.

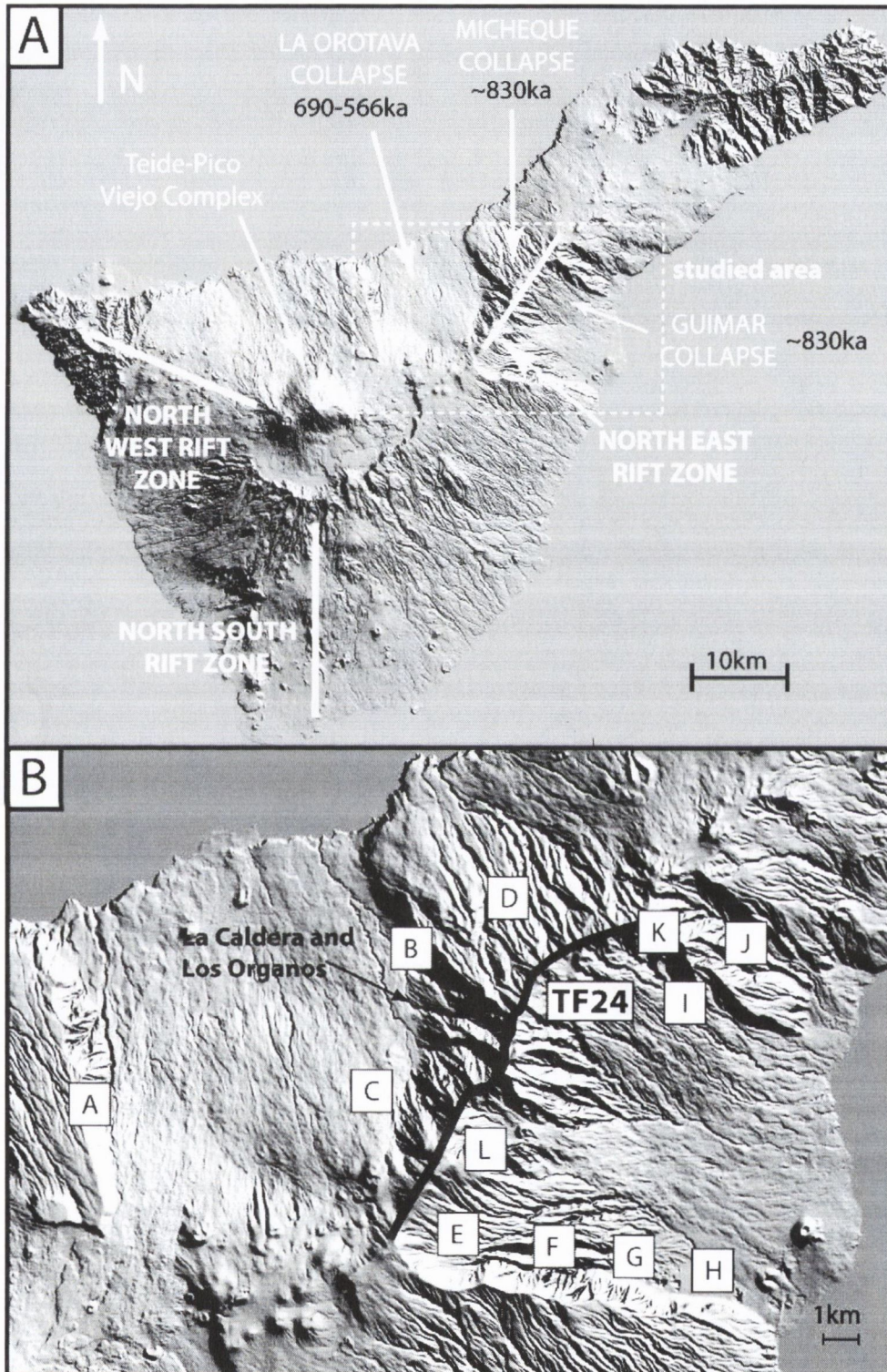


Figure 1: **A:** DEM of Tenerife (courtesy R.Paris, source: Grafcan©). **B:** The sectors studied in this thesis (labelled from A to L).

2. Petrology and geochemistry.

2.1. Petrographic groups

Six main groups of dykes have been recognized through petrographic study of about 90 hand samples and 47 thin sections. This sub-study has been done in collaboration with another PhD student, F.M. Deegan. All the thin sections show a cryptocrystalline groundmass that is composed of glass, feldspar (sometimes in needles), olivine and/or augite in different proportions. The most abundant group (group 1) is the aphyric/subaphyric one with rare or no phenocrysts. The second group (group 2) is composed of feldspar-rich dykes, but rocks of this group can contain rare olivine and/or pyroxene (**Fig.2A**). The third group (group 3) is composed by similar proportions of feldspar and clinopyroxene (augite) phenocrysts (**Fig.2B**). The fourth group (group 4) is composed of ankaramite dykes, i.e. with high crystal content (up to 40%) and clinopyroxene being more abundant than olivine (**Fig.2C**). Often, however, the crystal content is lower and not homogeneously distributed within the dyke (see **section 3.1**). The phenocrysts are usually several-millimetres in size and can be up to 2 cm across. The fifth group (group 5) has lower amounts of olivine and pyroxene phenocrysts (and of smaller-size, ~1 to 2mm) than the ankaramites (**Fig.2D**). The last group (group 6) is pyroxene-rich (**Fig.2E**). We also observed brown hornblende in several thin sections of the different groups. A few dykes fall between these groups and it is not always possible to determine to which group they belong. In general, the opaque mineral content was high. Study under reflective light show that opaque minerals are mainly magnetite. This is consistent with the results from the paleomagnetic study of **chapter 2**.

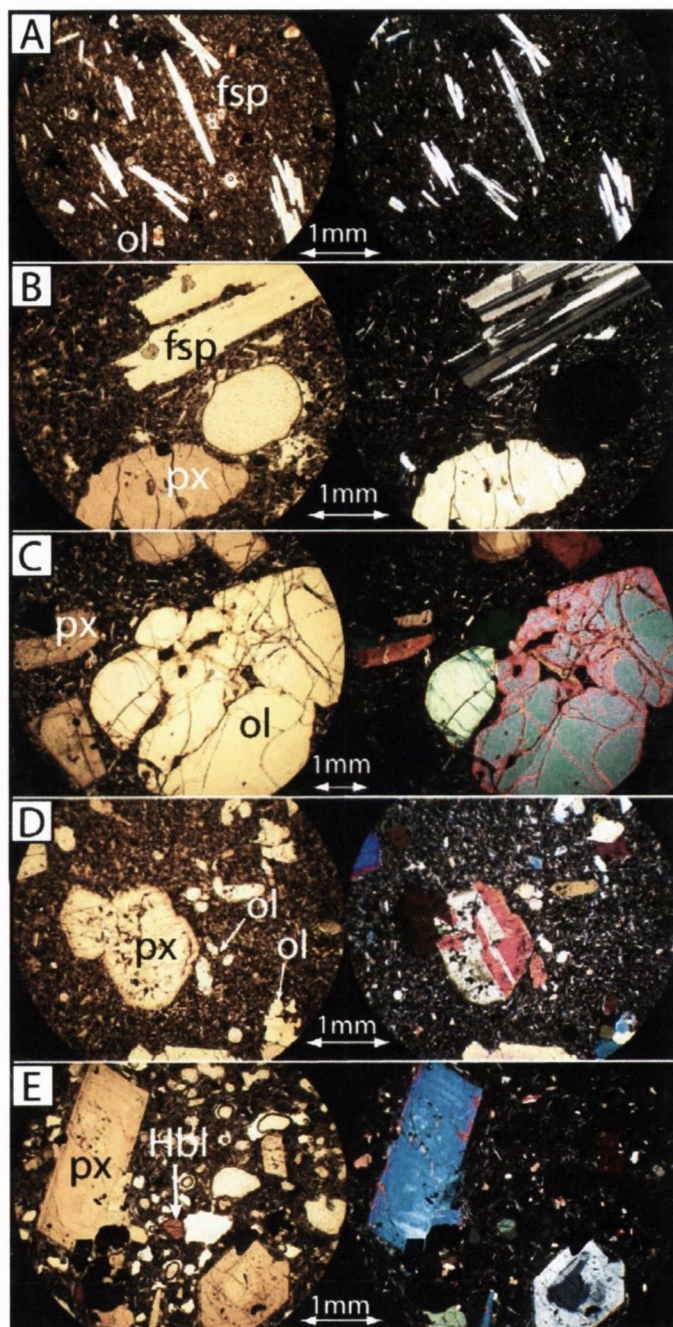
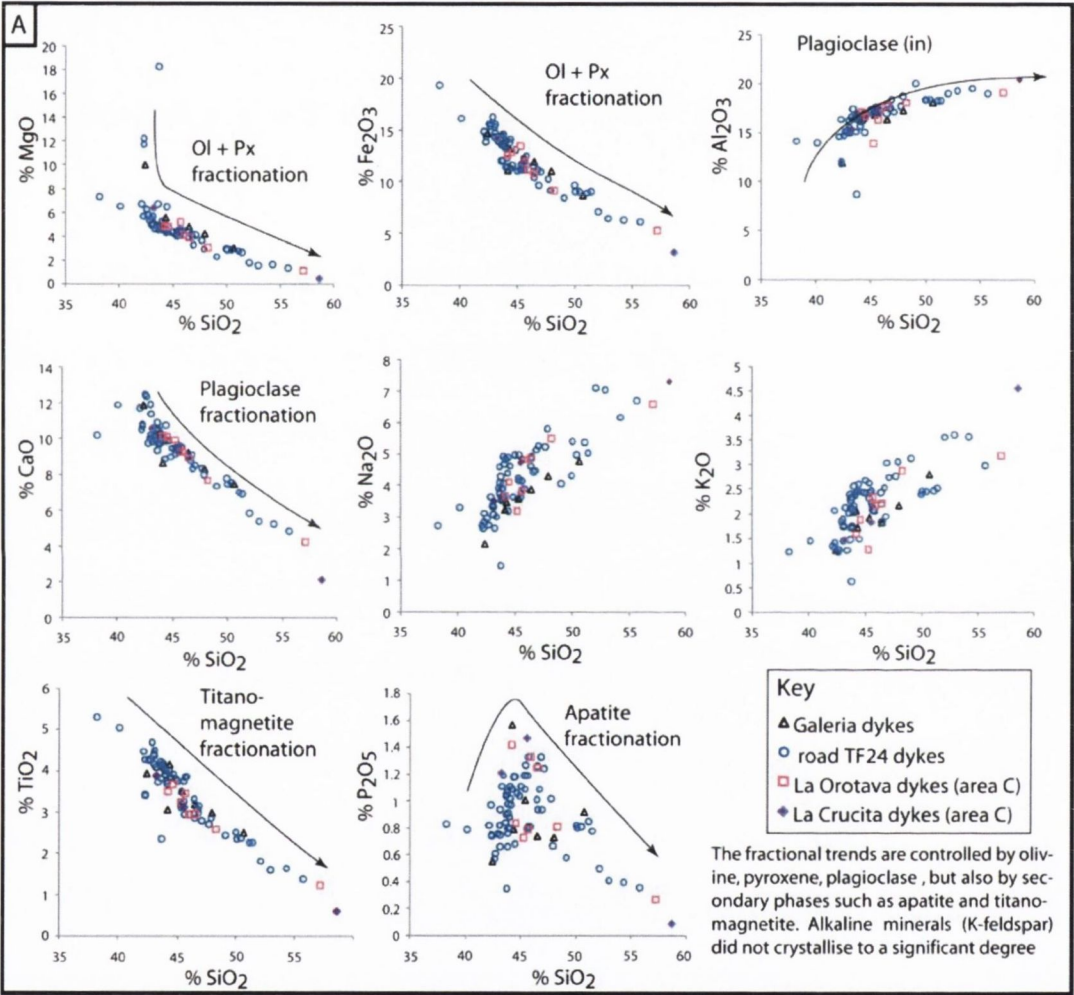


Figure 2: Pictures of representative thin sections of the different petrographic grouping. On the left, plane polarized light; on the right, cross polarized light. A: Feldspar-rich group (# 2). B: Feldspar + pyroxene group (#3). C: Ankaramite (#4). D: olivine + pyroxene group (#5). E: Pyroxene-rich group (#6). Fsp = feldspar; Px = pyroxene; Ol = olivine; Hbl= hornblende.

2.2. Geochemistry: major elements.

The samples were jointly prepared as powders at Trinity College by F.M. Deegan and myself before being sent and analyzed at IFM Geomar (Germany) by Dagmar Rau (see *Troll and Schmincke, 2002*, for analytical technique). The data obtained were then plotted

on standard SiO_2 vs. major element diagrams by F.M. Deegan (**Fig.3A**). In the TAS diagram, most of the dykes fall into the basanite field (**Fig.3B**). No clear grouping that matches the petrographic or the geographic groupings has been identified (**Fig.3**). Magma source processes and differentiation are not the topic of the present study and will be treated by F.M. Deegan independently.



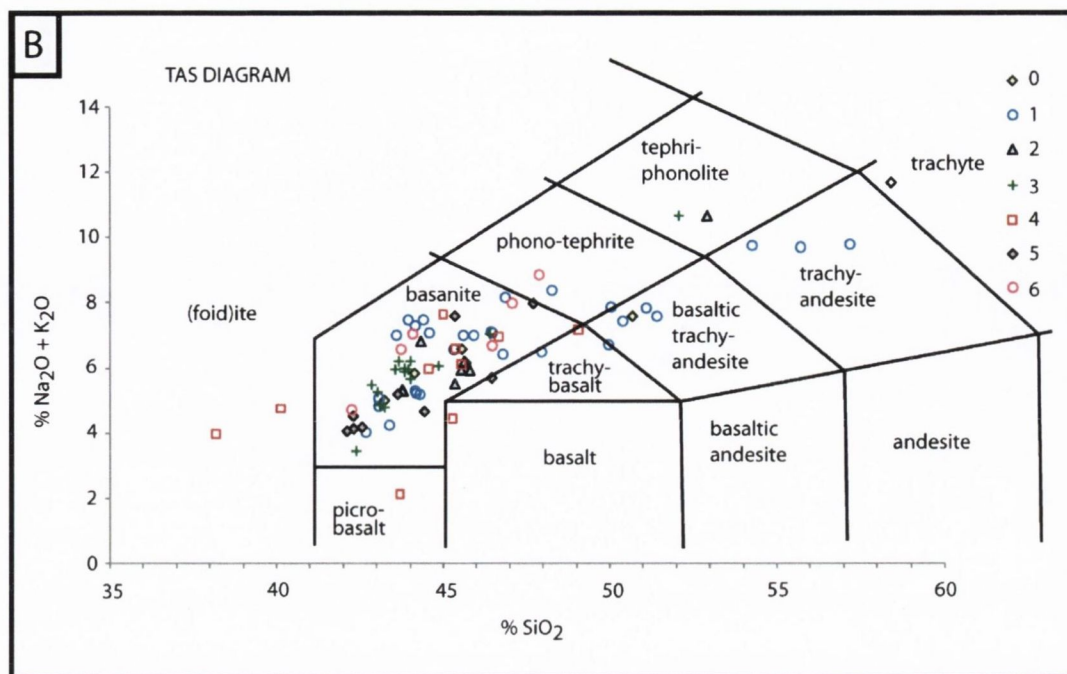


Figure 3: **A:** Major element plots of the *NERZ* analyzed rocks. **B:** TAS diagram of *NERZ* rocks. 0: Unassigned dykes; 1: Aphyric/subaphyric; 2: Feldspar-rich dykes; 3: Feldspar and clinopyroxene-rich dykes; 4: Ankaramitic dykes; 5: Olivine + clinopyroxene-rich dykes; 6: Pyroxene-rich dykes.

3. Dyke descriptions

3.1. Dyke diversity and morphology.

Most of the dykes are massive. Several dykes are strongly fractured usually parallel to their margins (**Fig.4**). Fracturing can affect the whole dyke, its margin only (with usually a massive centre), or its centre only. Also, some dykes display fractured zones that are randomly distributed. Only very few dykes are crushed, i.e. they are intensely cut by jigsaw fractures (**Fig.4**).



Figure 4: Pictures showing fractured and crushed dykes (jigsaw fractures). **A:** This dyke is fractured along the eastern margin only. **B:** Fracturing affects the whole dyke. **C:** Example of a crushed dyke (on the left).

The orientation of individual dykes is generally constant (at least at the outcrop scale), but local changes in strike, in dip or both have been sometimes noticed (**Fig.5**). Strike variations, sometimes of 10° only, are occasionally observed on dykes (for example a dyke at N040 slightly deviated to N050). Very few dykes followed more than 2 directions. In general, dip variation occurred within the same dip direction (for example, a dyke N020-80°W became N020-70°W). Only some dykes show a more extreme change in the dip direction (for example, a N000-80°N dyke became N000-80°S). Note that most of the outcrops are 2 or 3 meters above ground and display limited 3D information. Consequently, a dyke with an apparent constant orientation may have a variation of strike and/or dip above, below and/or further into the outcrop. Changes of orientation (strike, dip or both) can occur at the boundary between two layers of different competences (for example between a lava flow and its basal breccia), or within the same layer

(Gudmundsson *et al.*, 1999; Gudmundsson, 2002). Also, when two dykes are close to each other, the orientation of the younger one is deviated or propagates parallel to the older one (in that case, they may be contemporaneous). Change of strike and dip can also occur when the dykes are aligned en-échelons. Sometimes dykes are divided into two branches with different strikes.

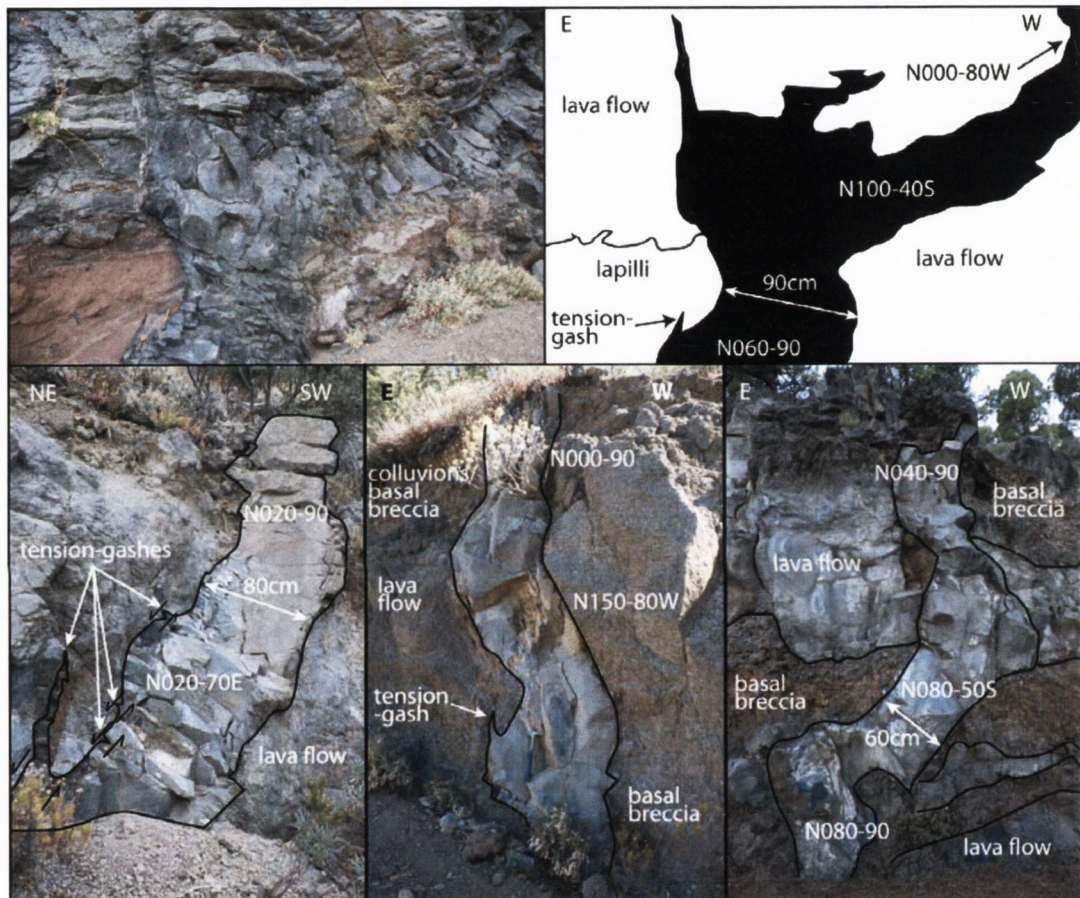


Figure 5: Pictures showing representative dykes that display variations in orientation (strike, dip or both). Such discrepancies can occur within the same layer (such as the dyke shown in left bottom corner and intruding a lava flow), or between two layers of different competence (rest of the pictures).

Two types of en-échelon dykes have been noted: “along-dip” and “along-strike” (as referred to by Clemente *et al.*, 2007; **Fig.6**). Some “along-dip” segmented dykes were

observed at the boundary between two different lithologies or within the same layer, i.e. the occurrence of “along-dip” segmented dykes does not follow a systematic pattern.

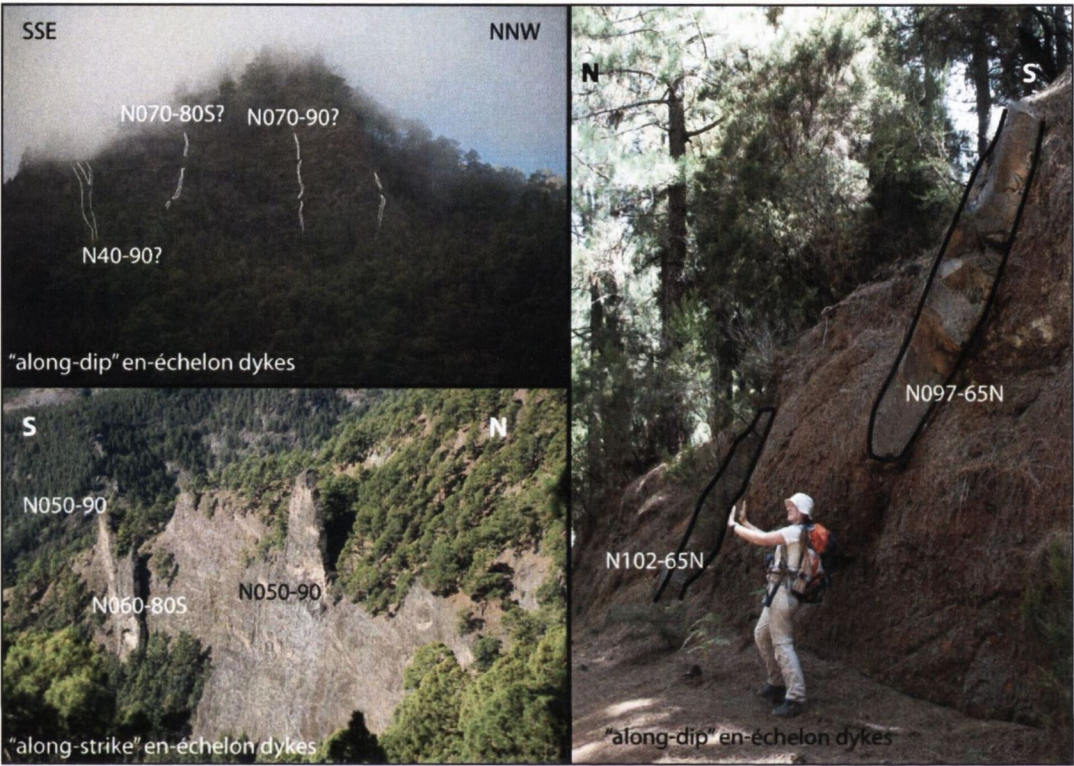


Figure 6: “Along-dip” and “along-strike” segmented dykes. “Along-dip” segmented dykes can occur between two layers of different competence or within the same layer.

The vesicle size is very variable and ranges from millimetre to centimetre-size (Fig.7). Vesicles are rounded, angular or elongate. Within one dyke, the size and shape can remain the same (for example, millimetre-sized rounded bubbles) or can change (for example, millimetre-sized angular and centimetre-sized elongate bubbles). The vesicles can be randomly distributed within the dyke, or can follow a gradation toward the centre (increase of vesicle size or content or both). Vesicles can be distributed following an upward bow, as frequently encountered in dykes and also in lava flow outcrops and recording the magma flow sense. Vesicles can concentrate along vertical bands and when their content and size becomes significant, they may even start to form fractures within the dykes. The vesicles in that case are very elongated, stretched and ruptured.

Another major feature contributing to dyke morphology is crystal content and distribution. Dykes with high crystal content (typically the ankaramites) are crumbly and weathered, due to the size and large amount of olivine and pyroxene, which are the least stable at ambient surface temperatures. As soon as the crystal content decreases, the rock is usually better preserved and more massive. Also, if the crystals are heterogeneously distributed within the dyke (i.e. crystal gradation towards the centre, towards one margin or vertical zones of high crystal content that alternate with sub-aphyric vertical zone), they become more massive and are thus more resistant due to lower crystal content.

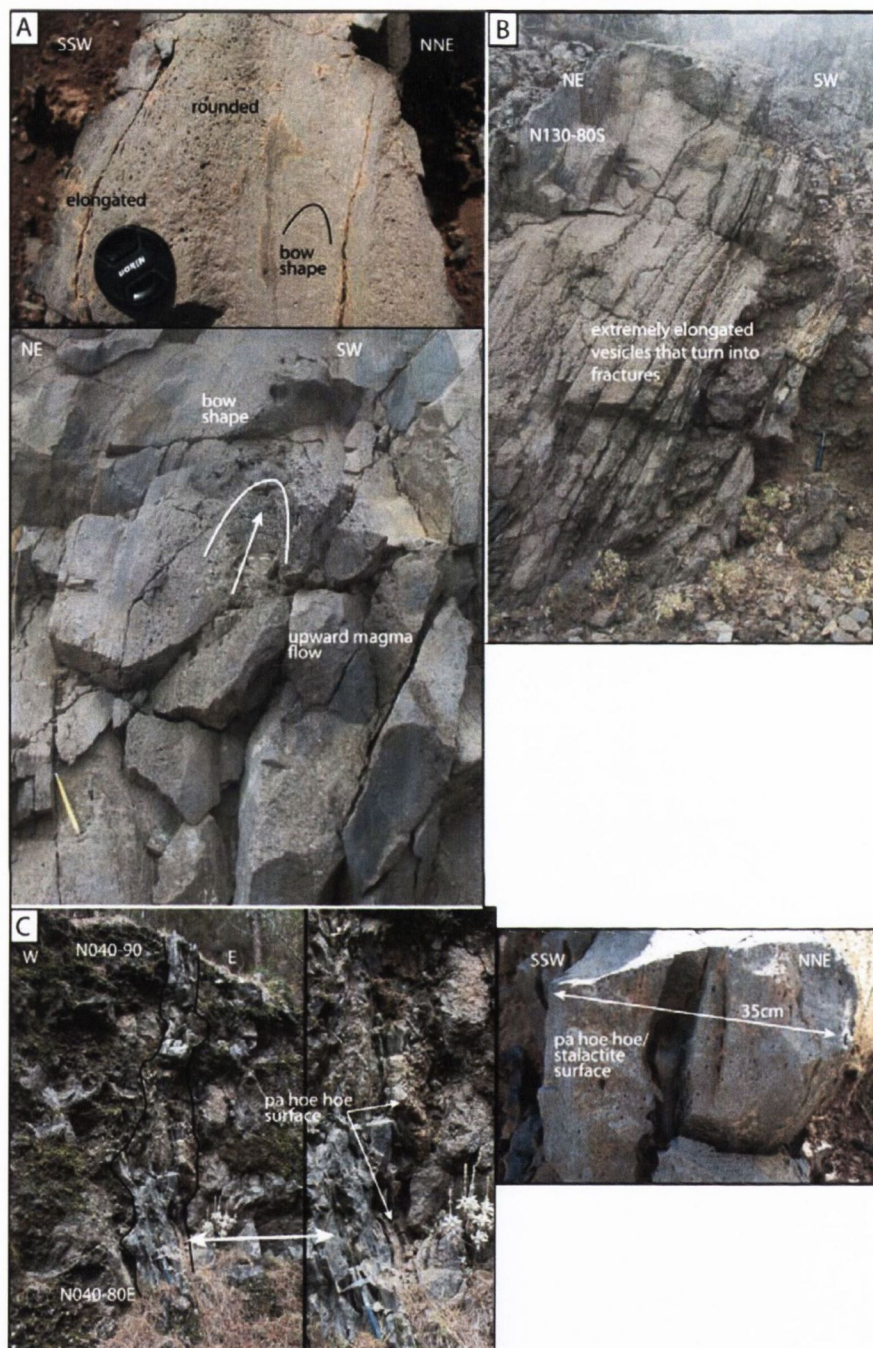


Figure 7: Examples of vesicle shapes and distributions within representative dykes. **A:** Two pictures of representative dykes showing upward bow-like vesicle distribution. **B:** Picture showing vesicle distribution along vertical bands; increase of bubble content and size allow fractures to form. **C:** Example of pahoehoe texture and “stalactite”-like feature.

A few dykes are highly vesiculated, especially in the centre (**Fig.8**). Sometimes, scoria can be found at the centre. Typical lava flow features, such as pahoehoe textures and “stalactite”-like features have been observed (**Fig.7**). Such textures have been previously observed and described e.g., in Lemptegy volcano, France (*Delcamp, 2005*). A few dykes contained centimetre-sized xenoliths of country rock. Tension gashes have been observed along several dyke margins, indicating shear movements along the walls (**Fig.5**). Note that the extrusive fingers that stick out from the dyke have been interpreted as tension gashes but could also be magma injection into the country rock.

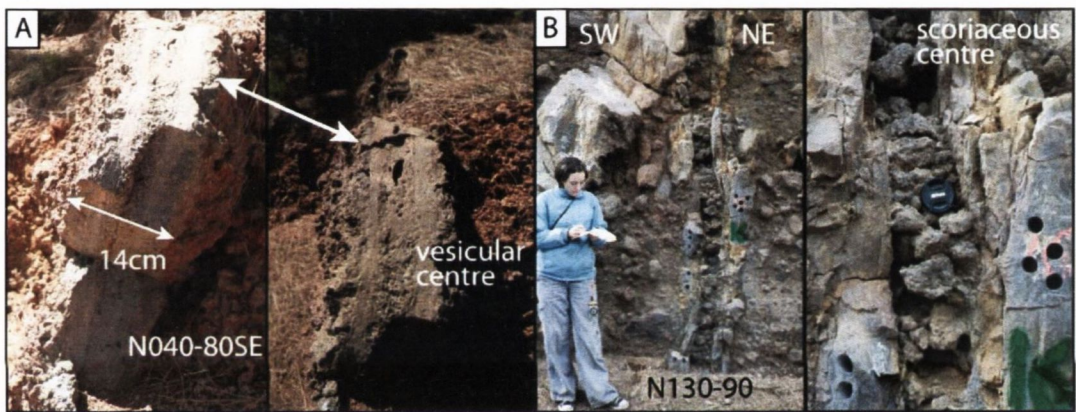


Figure 8: **A:** The two pictures show a representative dyke with a highly vesicular centre, that intrudes a cinder cone (sector F). **B:** The two pictures show a representative dyke with scoriaceous centre, observed along the road TF24.

3.2. Dyke thickness and density

The average dyke thickness for the whole studied sector (**Fig.1**) is 128cm (N=531 dykes). The averages for each sector (**Fig.1B**) are displayed in *table 1* and were calculated using only thickness values measurable with a ruler. In other word, the values of thicknesses estimated from panorama were not used for the average calculation. If the dykes were accessible and measurable with a ruler, then the thickest part of the dyke was considered. Note that, however, some of the thicknesses recorded along the trail G were estimates, but were still used for the calculations. This track cuts steep cliffs that cannot always be

safely approached, but view points were close enough to get reasonable estimates of the real thicknesses.

Table 1: Average of the dyke thicknesses for each sector (**Fig.1B**). * in the area LC two dykes of 10 meter thickness were found. If these are not included in the average calculation, the result changes from 265cm to 118cm.

Area	Road TF24	A/B/H	C	D	E	F	G	I	J	K	LC	RdI
Dyke number (N)	98	0	45	74	12	62	30	9	6	131	12	52
Avg. dyke thickness (cm)	105	/	188	91	72	82	120	149	93	116	265/118*	246

The chart in **figure 9** shows the distribution of dyke thickness. Most of the dykes are between 31 and 60cm thick. Note that the average dyke thickness (128cm) is shifted from the highest frequency peak (i.e. 31-60cm). We observe a progressive decrease of dyke number towards higher thickness values.

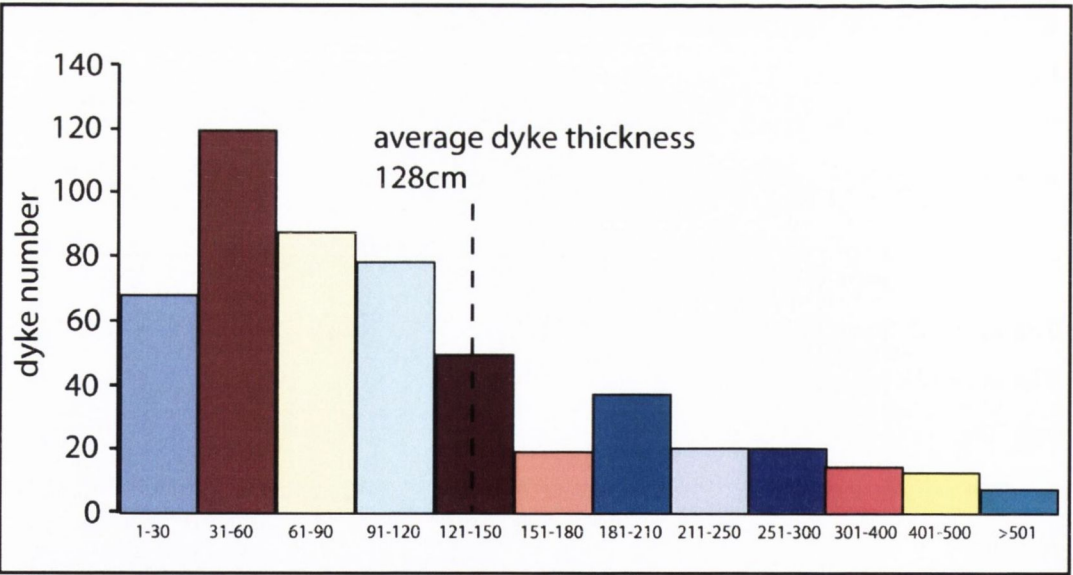


Figure 9: Histogram showing dyke thickness distribution (in cm). Number of dykes for each thickness category is displayed on the y-axis.

Within a dyke, thickness variations can occur, especially when the dyke passes from a low competence zone (e.g. scoria) into a more resistant stratum (e.g. a lava flow), where the dyke becomes thinner (**Fig.10**).

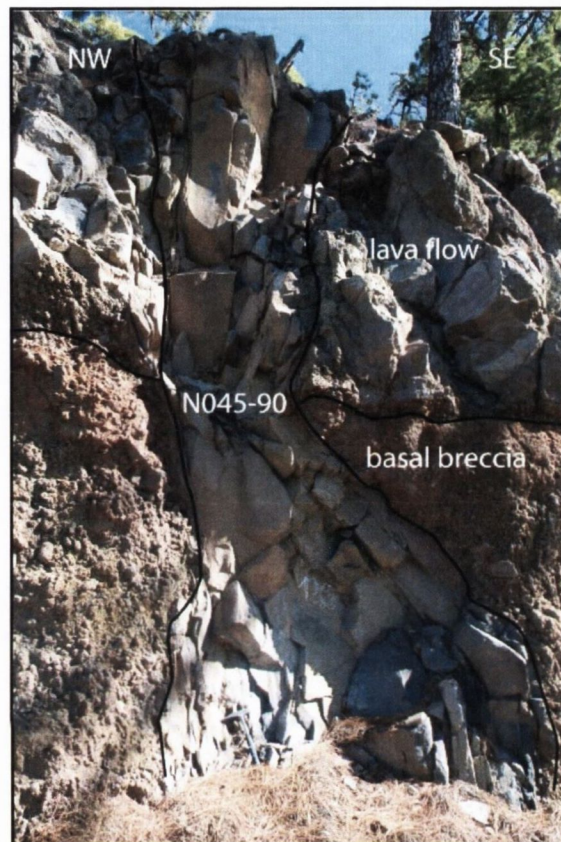


Figure 10: *Variation of dyke thickness due to change of host-rock lithology: the dyke is thicker within the basal breccia (lower part, low competence layer) than within the lava flow (upper part, high competence layer).*

Looking at table 1 and **figure 11**, we also note that dykes at higher altitude and at an intermediate altitude are overall thinner than the ones at the base of the ridge. Moreover, in the sector LC (La Crucita-belong to sector C), two dykes of 10 meters thickness were found. If these are not included in the average calculation, the result changes from 265cm to 118cm in this sector (*in **table 1**).

One of the trails in sector F leads down into the valley, where a dyke swarm can be observed. From afar, the dykes were estimated to be between 1 and 2 m thick.

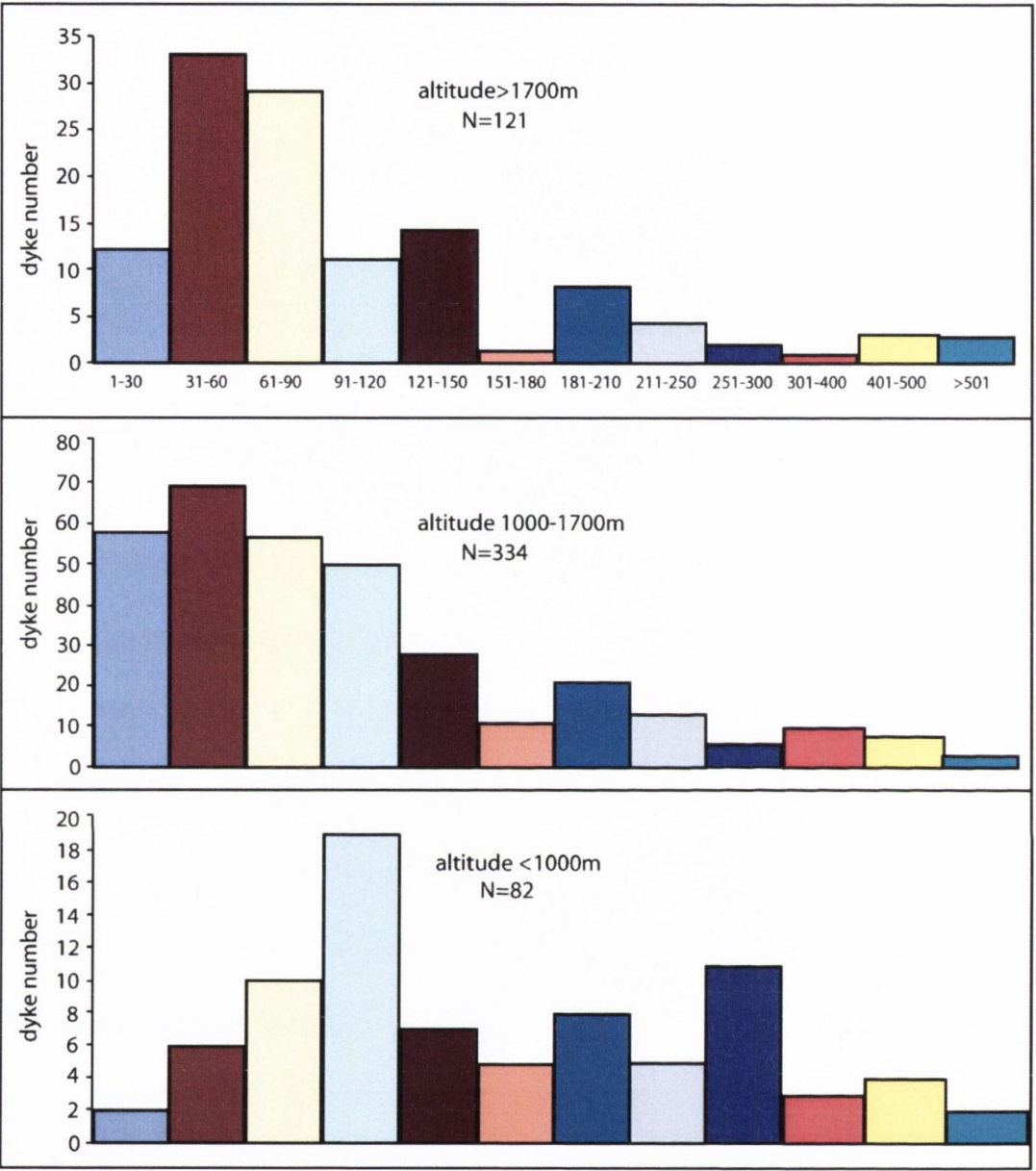


Figure 11: Histograms representing the number of dykes for each thickness category (in cm) for a specific altitude interval. From top to the bottom: altitude >1700m (highest); altitude between 1000 & 1700m (intermediate) and altitude <1000m (lowest). Note that dykes are overall thicker at lower altitudes.

The dyke density is higher towards the ridge and along the SW and NE side of the Güímar scar. No dyke has been found along the west side of La Orotava. The east side of

the La Orotava scar is mainly covered by vegetation and a thick pine needle cover that probably hides the dykes. The intrusive density along the 8km road section (TF24) is 1.3%. However, the intrusive densities calculated from three different smaller portions along the same road, respectively from NER 1 to NER 13 (437m), from NER24 to NER29 (119m) and from NER31A to NER45 (283m), give respectively slightly higher values of 2.6%, 3.6% and 9.9%. The intrusive density for these three horizontal logs is 5.2%. For the sector K, an intrusive density of 5.7% along a section of 1.7km, and of 6.6% along a 2.3km section has been calculated. Sector C corresponds to an intermediate altitude, compared to the galeria that represents the lower part of the rift and the road TF 24 that represents the upper part of the rift. In sector C, the intrusive density is 2.4%. In the galeria Rio de Igueste the intrusive density calculated along the 1.08km section is 11.8%. However, if the last 800m is considered in isolation, the intrusive density is ca.14%. For sector G, a value of 4.6% along a 775m section is derived (see **Table 2**).

Table 2: *Intrusion density calculated in several sectors (Fig.1B). See text for details.*

Sector	road TF 24			C	G	K	galeria Rdl
Section (m)	8000			3500	775	2300	1080
	437	119	283				
Intrusion density (%)	1.3			2.4	4.6	6.6	11.8
	2.6	4.2	9.9				

4. Country rock

The dykes intrude mainly lava flows and their associated basal breccias, as well as lapilli and scoria deposits. Up to the rift, along road TF24, several dykes intrude colluvial deposits, providing evidence for the existence of paleovalleys.

The galerias are a humid environment, and the rocks turned out to be very crumbly and seem to be altered. Clays can act as ductile layers (referred as Low Strength Layers or LSLs, *Oehler et al., 2005*) and help to destabilize the rift. To check for the presence of clays, several samples of country rocks (lava flows and scoria) have been collected from the two galerias visited and have been analysed using X-Ray Diffraction at TCD (see **appendix 1**). It appears that the sampled rocks contain only a low amount of clays, such as illite, or expandable clays, such as smectite. Nevertheless, the rocks in the galerias are very crumbly. Furthermore, large amount of scoriae has been noted across the areas. If

these rocks experience a high lithostatic pressure (i.e. when situated at a deep enough level), they can deform more easily than massive units and they can thus act as LSLs (strength of 20 to 40 MPa; *Rodríguez-Losada et al., 2009*). Those units can thus help to destabilize the rift. Also, in several places across the studied areas, crushed lava flows have been observed. Sometimes the lava flows are even turned into a fine whitish or greyish powder (**Fig.12**). This could be due to intense post-emplacement deformation or intense alteration. Consequently, those layers can act as LSLs.

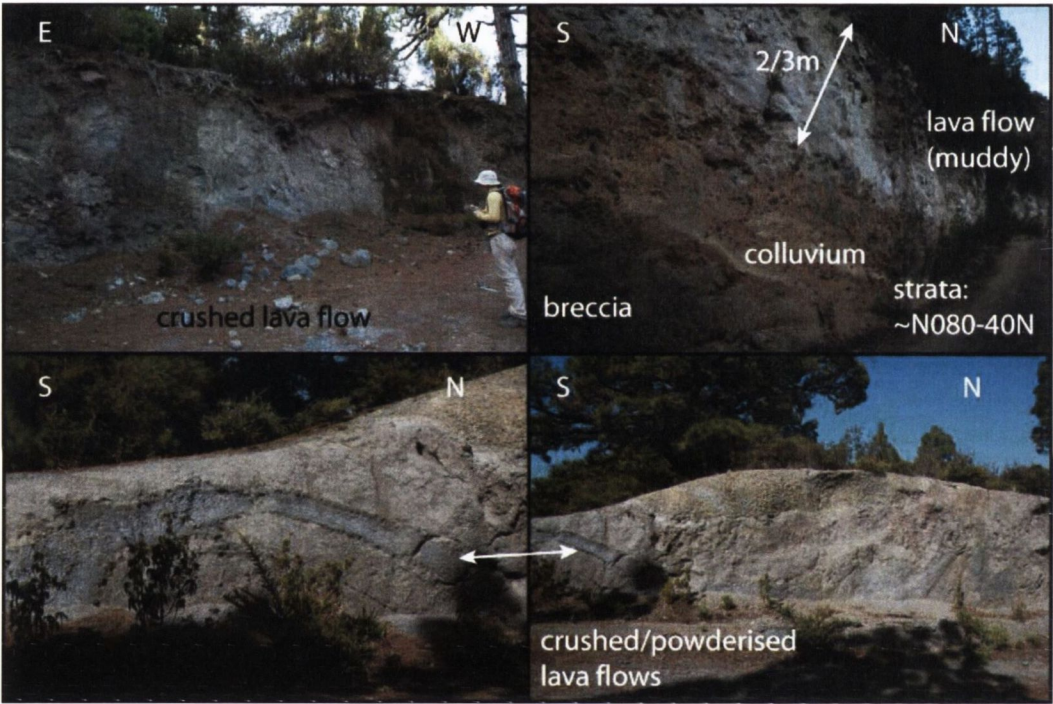


Figure 12: *Slightly and intensely crushed lava flows. Those crushed layers are of low competence and can help to destabilize the rift.*

5. Structural observations

Only a few cross-cutting relationships between dykes have been observed and are thus insufficient to infer any relative ages between the different dyke populations.

Rose diagrams for each area (**Fig.1B** and **Fig.13**) have been drawn, using the most reliable data, and dykes with variable strike were not taken into account. When dip variations occurred, we took the average into consideration.

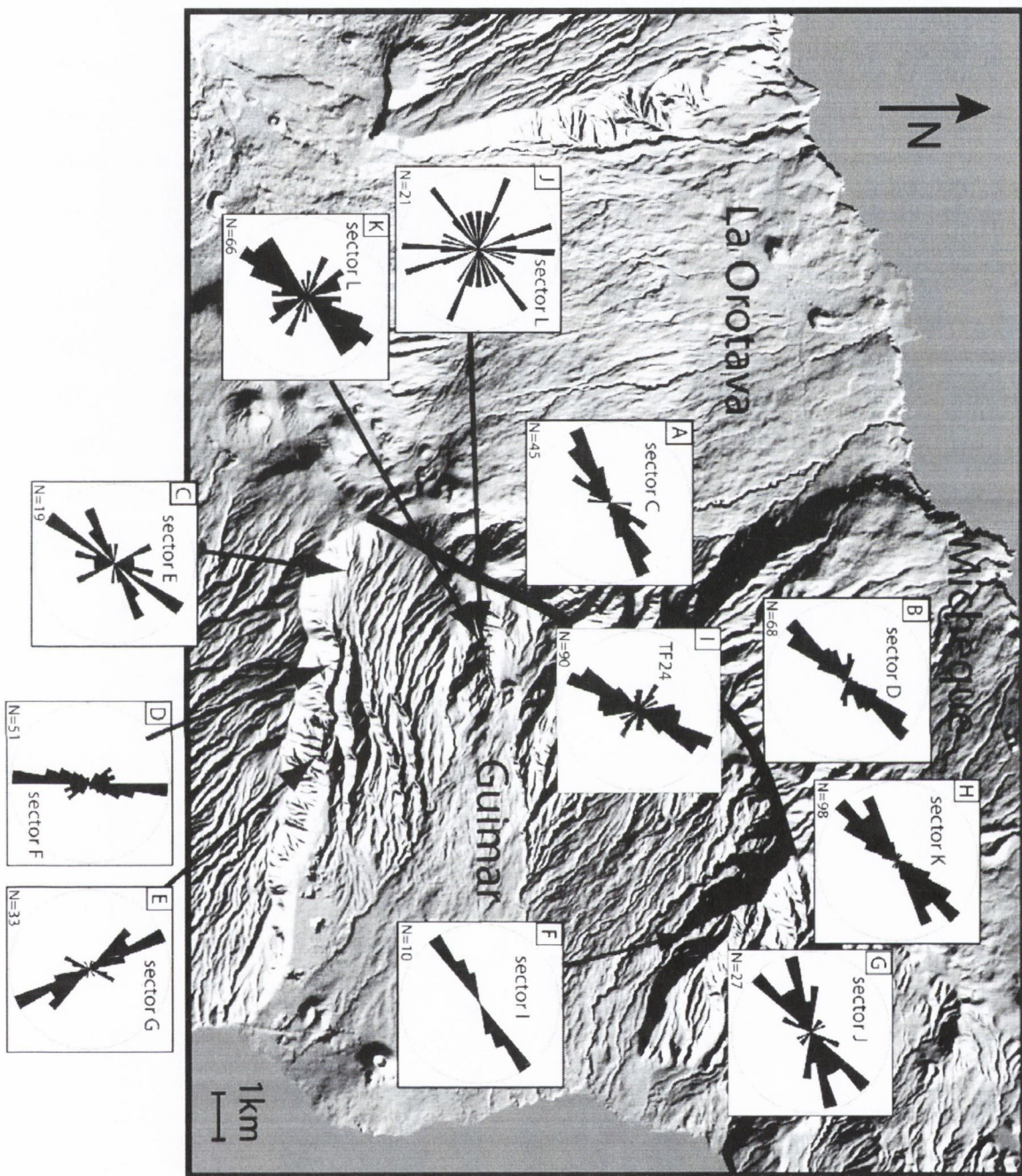


Figure 13: Rose diagrams of dyke orientations observed in the different studied sectors.

5.1. La Orotava valley

5.1.1. Sector A and B (West and East side of La Orotava scar).

Neither dykes nor fractures have been observed along the entire cliff of the sector A, which displays really good outcrops (almost no vegetation along the scar). The other side of the collapse scar (sector B) is mostly covered by vegetation, and the access is difficult.

5.1.2. Sector C (north scar of La Orotava).

The dykes recorded along sector C are situated at an altitude of ~1500m a.s.l. The area is densely fractured and faulted, following a constant direction of N060/070 (ENE-WSW; **Fig.14C**). The dykes also mainly follow the same direction at N060/070 (**Fig.13A**). The other observed directions, although less pronounced, are N030/N040 and N140 (from N140 to N170). Major fracture directions are observed on satellite images (from Google Earth) and are orientated N000, N030 and N060 which coincide with the fracture and dyke orientations measured in the field (**Fig.15**).

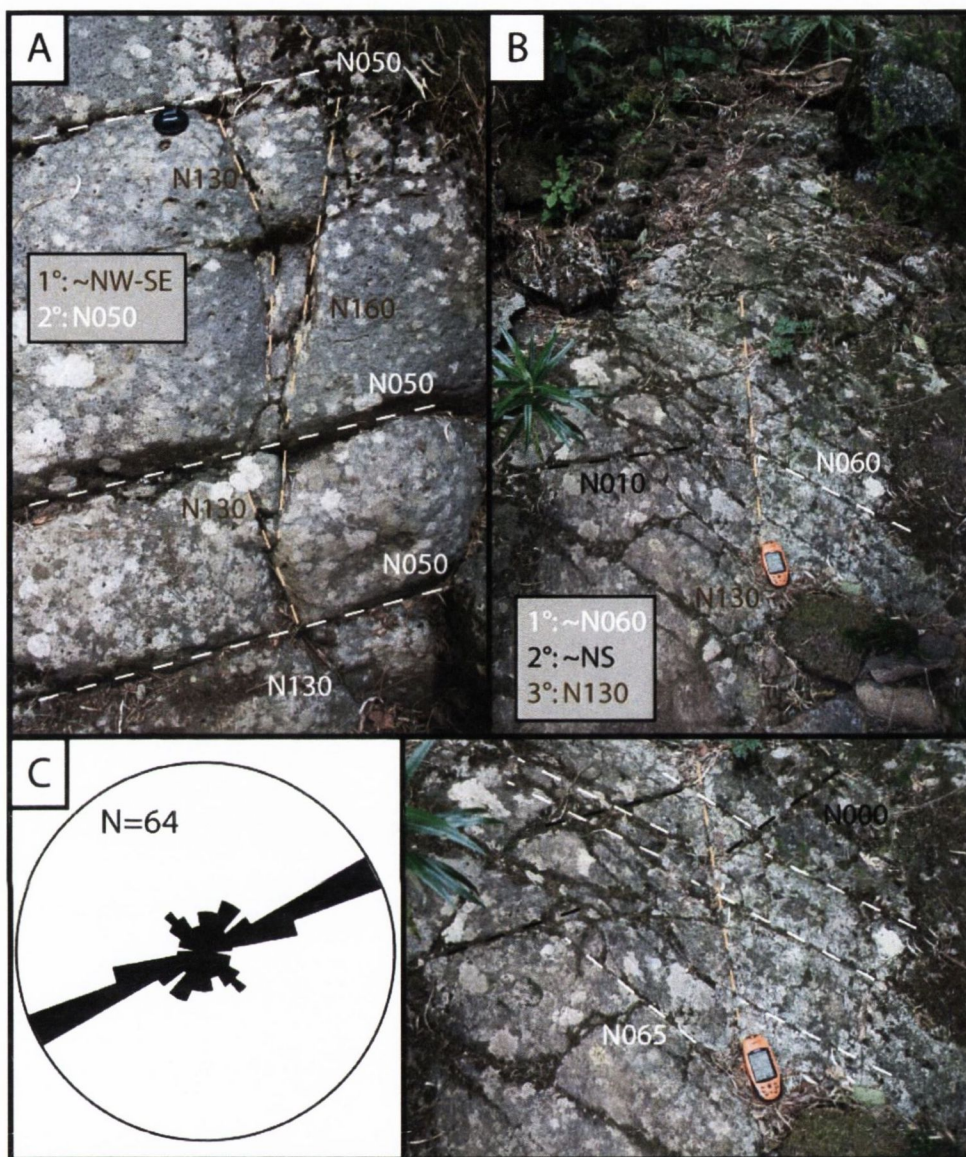


Figure 14: **A:** Outcrop in sector C showing fracture relationships. **B:** Another outcrop in sector C that displays fracture relationships; the photo below B is an enlarged photo of the outcrop. **C:** Rose diagram of the fractures recorded in sector C. Local relative age can be deduced from fracture relationships (see **A** and **B**), but they do not correspond to a general pattern across the area. For example, the NW-SE fractures in **A** are younger, while they are older in **B**. Thus, no general relative ages between fracture populations can be deduced for this sector. They may have been active at the same time.

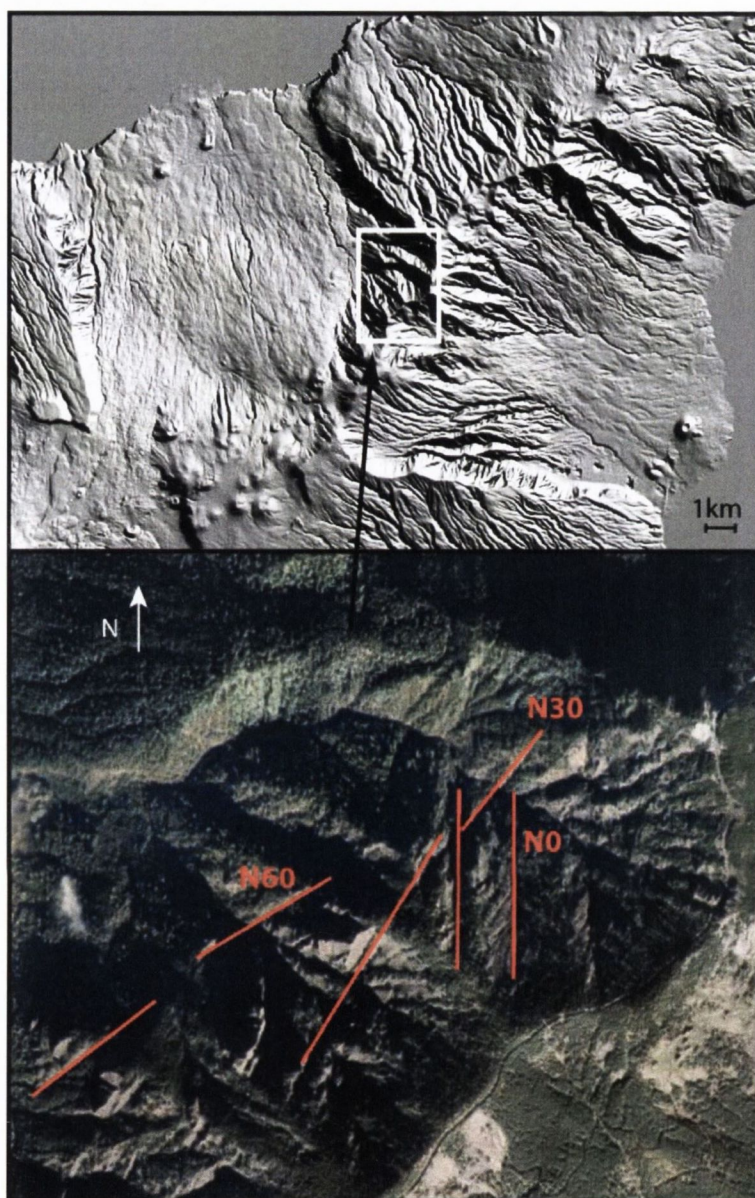


Figure 15: Satellite image (from Google Earth) of the north scar of La Orotava on the right (DEM from Grafcan©. Courtesy of R.Paris). The main structural trends are underlined in red.

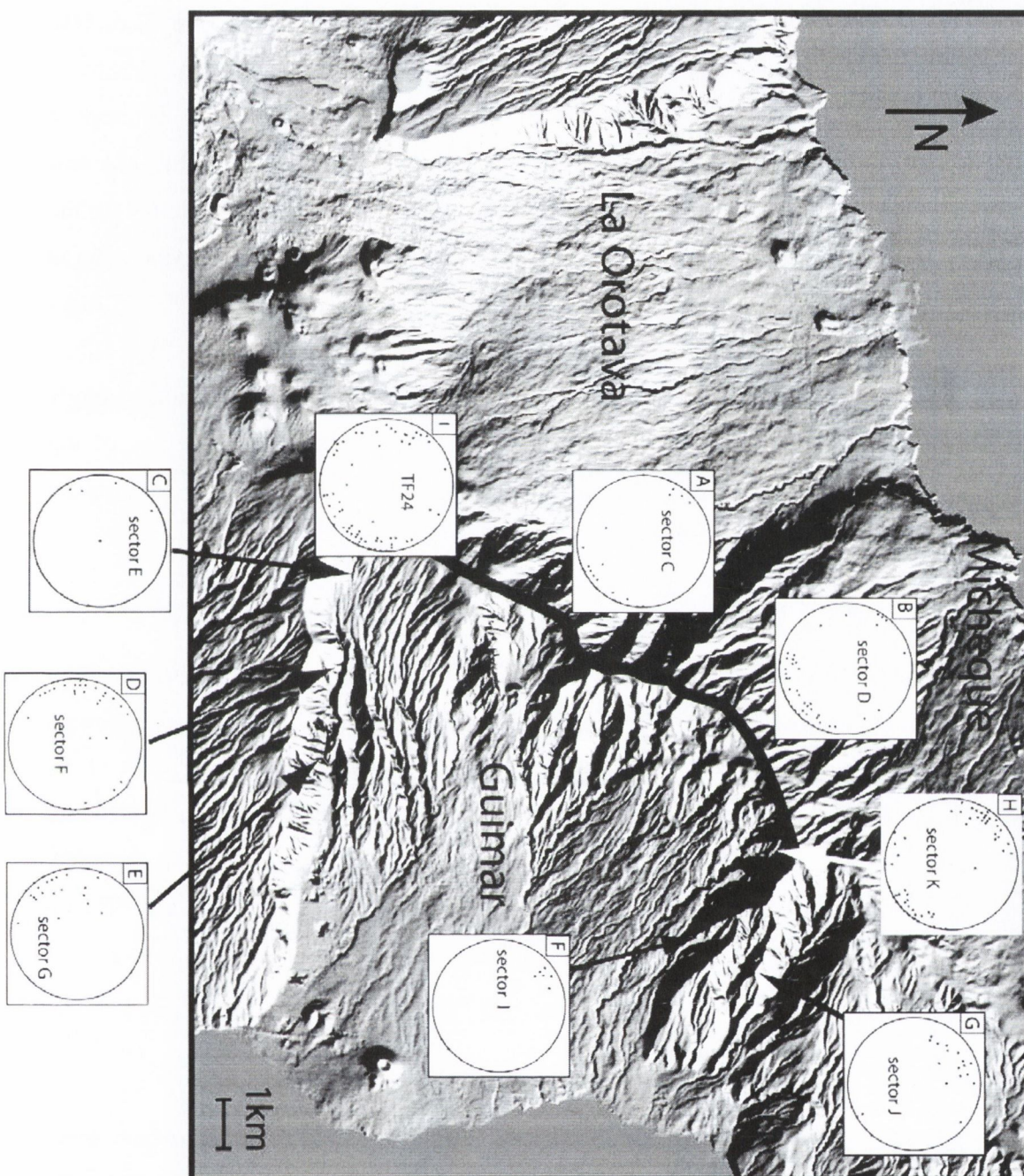


Figure 16: Stereoplots of dyke poles for each sector studied.

Most of the dykes are vertical (69%) and those that dip do so towards the W or NW (22%). Only 9% dip towards the opposite directions (E, NE and SE, **Fig.16A**).

This sector is densely fractured compared to the other zones and many small outcrops at the head of the collapse allow relationships between fracture sets to be seen (**Fig.14A and B**). However, no overall relative ages between fracture populations can be inferred for the entire area. They may have thus been active at the same time. For example, one outcrop displays N130 as the younger direction while another one shows N130 as the older direction (**Fig.14A and B**). The fractures display a clear trend at ENE-WSW (**Fig.14C**).

Dip variations have also been observed where dyke dips change from vertical to strongly inclined towards the east, north-east or south, or change from a high degree of dip (typically $<70^\circ$ towards the east, north-east or southwards) to a lower angle towards the same direction (E-NE or S).

5.1.3. Sector D (Micheque area)

This sector corresponds to the old Micheque collapse embayment described in detail by *Carracedo et al. (2009)*. The rose diagram of the dykes outcropping in this area shows a main trend of N030 to N050 (**Fig.13B**). Note that the dykes at the western part of the sector (i.e. the one closest to the east scar of La Orotava) show mainly an E-W trend. Very few fractures have been observed compared to sector C. Of all the dykes recorded, 51% are vertical, 39% dip towards the north, west, northwest, and the remaining 10% dip towards the south, east and south east (**Fig.16B**).

5.2. Güímar valley

5.2.1. Sector E (NW head of Güímar collapse).

The rose diagram of the dykes here shows a main trend at N030, then a subsidiary trend at N060 and a minor one at $\sim$ N150 (**Fig.13C**). The dykes are mostly vertical or dip gently towards the south or the east (**Fig.16C**). The exposures are mostly along a horizontal plane, allowing individual dykes to be followed for several meters.

5.2.2. Sector F (SW of the Güímar scar).

The main trend of the dykes is N000 (**Fig.13D**). As specified earlier, the rose diagrams were drawn using only reliable data. They thus do not include the strike of dykes recorded from distance. At the south of the sector, a view point allows a dyke swarm to be observed at N140/150/160. This NW-SE trend becomes more dominant from the north to the south of sector F.

Half of the dykes recorded in this sector (50%) dip towards the south, south-east, and east, while 46% are vertical. The remaining 4% dip towards the west or north (**Fig.16D**).

5.2.3. Sector G (SW of the Güímar scar).

The main orientation highlighted by the rose diagram (**Fig.13E**) is around N130-N150 i.e., NW-SE. A dyke swarm at N090/110 has been observed along a cliff (north part of area G). However, these orientations have been taken from afar (panorama view), so their reliability is probably lower.

The dykes mainly dip towards the east, north-east or south (86%), 12% are vertical and only one dyke dips northward (**Fig.16E**).

5.2.4. Sector H (SW of the Güímar scar).

No dykes or fractures have been found along road cuts; only lava flows and their associated top and bottom breccias.

5.2.5. Sector I (NE of the Güímar scar).

Dykes are found almost exclusively along the SW part of the scar (except at sector H) whereas here, dykes are only observed when the head of the collapse is approached. They trend mostly N050 and dip towards the south or south-east (**Figs.13F & 16F**). Dyke density is lower on this side of the scar than in the SW.

5.2.6. Sector J (NE of the Güímar scar).

The dykes trend mostly N040/N050 and N070 (**Fig.13G**). As in sector I, the dykes dip mainly towards the south or south-east (**Fig.16G**).

The galeria Rio de Igueste is situated in this sector (see **appendix 2**). Within the water tunnel, the compass may have been influenced by the surrounding rocks. However, the dykes recorded there show a uniform trend at N060 and are vertical or dip steeply southward.

5.3. Sector K (Boca Del Vallee).

The dykes show two main trends at N040 and N060 (**Fig.13H**). Dykes outside the collapse scar (east of the area) mostly trend N060/N070, while the dykes inside the scar (head of Güímar collapse, west of the area) trend N040 and N060 (as seen in the rose diagram for the whole area).

The dykes are mostly vertical (54%) (**Fig.16H**). Dykes dipping towards the south, east and south-east are found everywhere in the sector, while the dykes dipping towards the north, west and north-west are mostly confined to the west of the sector, corresponding to the head of Güímar collapse.

As observed in other sectors, many dykes that are fractured parallel to their margins have been observed. We noted that here, most of the fractured dykes show an orientation of ~N040 and also, to a lesser amount, N060.

5.4. Road TF 24 (black line on figure 1B).

Along the road, the dykes are mainly parallel to the rift (i.e. N030/N040), but follow also the other directions recorded elsewhere (**Fig.13I**). The dips are also very variable (**Fig.16I**). The dykes crop out chiefly around the Arafo volcano. The eastern ones are mainly dip to the east and south-east, while the others dip mainly towards the north, north-east and north-west to west. However, a ~26° vertical clockwise rotation of this area has been highlighted with paleomagnetic data (see **chapter 2**). The original dyke trend should then be around N000. However, it is not possible to calculate the original direction of each dyke using the paleomagnetic data as the vertical clockwise rotation was estimated from a group mean.

5.5. Sector L.

Sector L is situated in a small collapse scar partially filled with recent deposits from a cinder cone volcano (Arafo) within the scar. The dykes recorded have variable strikes (**Fig.13J**) and mostly dip towards the N and NW. The sector displays several dyke swarms that show one main trend NW-SE (**Fig.13K**). Note that the orientation may not be very precise as the directions were measured from a distance of several hundred meters. The thicknesses of the dykes could also be estimated only.

6. Field interpretations.

6.1. Dyke morphology and diversity

Two mechanisms can be envisaged to form the fractured dykes (**Fig.4**): the movement of magma during cooling, or local tectonic movement (or both). Some of the dykes contain vesicles, sometimes in high concentration, distributed along vertical bands. The vesicles are usually highly stretched and start to form fractures (**Fig.7**). It can thus be inferred that (1) while the magma is flowing within the dyke, the vesicles follow preferential paths and concentrate along vertical bands; (2) when the magma starts to cool, the shear movements become more important, stretching the pre-existing vesicles; (3) eventually, the deformation becomes so intense that the stretched vesicles start to form fractures; (4) the temperature is still decreasing and the deformation is now brittle with full fracture formation. The original distribution of the vesicles is probably due to various factors, such as the magma viscosity and the temperature gradient that will influence gas transfer within a dyke. Intense stretching during cooling is likely not the only process that allows dyke to fracture. In sector K (Boca Del Vallee), one direction of fracturing is prominent (N040) and likely represents a local tectonic trend. This implies that this preferential tectonic direction was active during and/or after dyke cooling. Unfortunately, it is not possible to conclusively distinguish between the two fracturing origins (cooling or tectonic) and it is probable that the two acted in concert. The crushed dykes and the crushed zones within some dykes are, however, interpreted as tectonic in origin, as cooling is not likely to be responsible for the formation of the jigsaw fractures (i.e. network of randomly orientated fractures).

Regarding the vesicle shape varieties (i.e rounded, angular and elongate) between dykes and within individual bodies, it can be inferred that the dynamic and cooling processes are very different from one dyke to another and are even different within individual dykes. The coexistence of elongate and angular bubbles within the same dyke indicates that some areas are affected by ductile deformation (elongate vesicles) while others by ductile/brittle (transition) shear movements (angular) or both (presence of both elongate and angular vesicles in the same zone). The regime's variability (brittle and ductile) and the vesicle distribution are likely due to temperature and viscosity gradients across a dyke. In addition, time may play a role in the vesicle diversity, as the margins and the centre of the dykes may reflect the early and the late cooling history respectively. Also, the rounded vesicles represent a second generation of bubbles that grew after, when the shear movements stopped (*Schmincke and Swanson, 1967*).

The magma flow direction can be approximated in some dykes, when vesicles are organized along an upward bow. This geometry is due to a magma velocity gradient across a dyke (velocity increases from the margins towards the centre). The fact that the vesicle shapes and that such bow-like features are preserved after cooling suggests that the magma has cooled sufficiently at that point to become so viscous that the bubbles cannot rise anymore.

Dykes with a strongly vesiculated or scoriaceous centre were found along road TF24, the Boca Del Vallee trail (sector K), the Las Lagunetas trail (sector D), and one was found in scoria up to the sector F, although this was interpreted as related to a cinder cone. These are all located relatively close to the present-day surface, which is also probably close to the ancient topographic surface. The scoriaceous areas probably formed because of gas concentration occurring when the dykes were approaching the surface. When the gas becomes volumetrically dominant, the magma starts to inflate and eventually to fragment allowing scoria to form. This kind of dyke morphology has been observed at Lemptegy volcano and seems characteristic of an eruptive vent (*Delcamp, 2005*). However, as the dykes are not entirely fragmented, it can be deduced that the vents may have widened during the eruption, allowing the pressure to progressively decrease and, consequently, only the centres were able to fragment.

A deviation of dyke orientation (strike, dip or both) at the transition between two layers of different competence is likely related to a mechanical effect. However, if the strike, dip, or both change within the same stratum, the variation can be due to the local tectonic regime. The deviation can also be due to the dyke propagation mode, especially if we are close to the tip. Indeed, several dyke tips are “wiggly”. In addition, a divergence in dyke orientation (strike, dip or both) can occur close to other dykes. In that case, it can be deduced that the deflected dyke is younger. Obviously, its new orientation is not indicative of the regional tectonic stress regime.

6.2. Dyke thickness and density

In general a slight thickness increase towards the base of the rift has been noticed. The variation of dyke thickness is not really significant between the shallow and intermediate altitude levels. However, if we compare the low and intermediate levels with the deepest level (corresponding to the galeria), the thickness variation is more pronounced (**Fig.11**). The dykes are mostly concentrated along the rift and along the Güimar valley scars. Consequently, the intrusive activity seems to have been concentrated in these areas, as opposed to La Orotava. Note that the dyke density reported depends on the amount of outcrop and that the real density is probably somewhat higher. For example, the low intrusive density of 1.3% calculated along the road section (TF24) is partly due to the activity of recent cinder cones that cover the western part (from the national park entrance) with scoria and lapili deposits. The feeder dykes of these cones would increase the intrusive density calculated in this chapter (**section 3.2**). The intrusive density calculated along three different logs is thus much higher. A density of 3.4% has been calculated along the section that is not covered by recent deposits (2.6 km; from ~ Mirador de Ortuno to the national park entrance). The intrusive density is yet higher at the base (e.g. galeria Río de Igueste) than at the top of the rift. Several studies have already shown that the dyke proportion is higher at deep level as many dykes become arrested before reaching the surface (*Hoek, 1995; Gudmundsson et al., 1999; Gudmundsson 2002*).

6.3. Country rocks

On the upper part of the rift, several dykes are surrounded by colluvial deposits. The question arising from this observation is whether the dykes are younger or older than the colluvium? Indeed, the dykes could have been exhumed, as it is largely observed around the young cinder cones (west part of the NERZ) where the lappili and scorias have been eroded away, allowing some dykes to stand proud. Then, the colluvial deposits surrounded the dykes. However, the colluvium bounding the dyke margins are often slightly cooked. It is thus more likely that the dykes have intruded the colluvium and are thus younger. Consequently, the upper part of the rift has experienced period(s) of erosion that may have been concurrent with intrusive activity.

6.4. Structural interpretation (Fig.17)

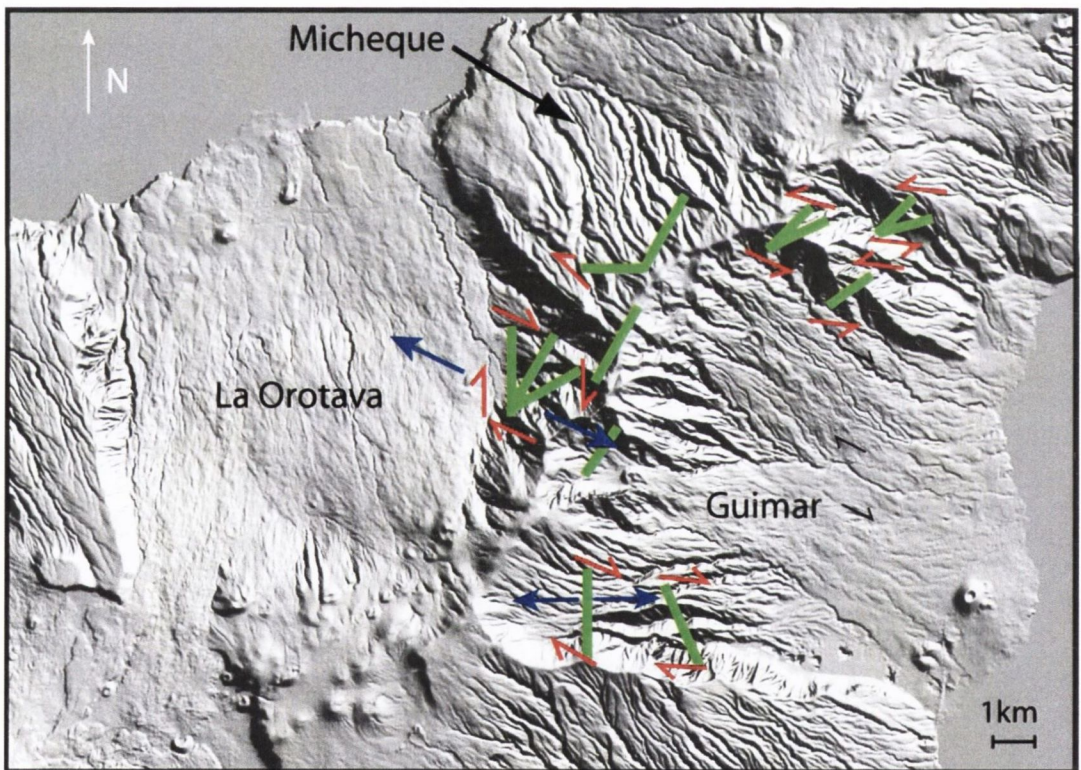


Figure 17: Interpretative structural sketch of the different areas. The green lines represent the main structural directions, the red arrows symbolize the strike-slip movements inferred from dyke orientations and the blue arrows indicate the possible direction of extension deduced from dyke orientation patterns.

6.4.1. Micheque

At the east of La Orotava valley (sector D) is the fill of the Micheque collapse, described by *Carracedo et al. (2009)*. The 14 dykes recorded directly at the western side of Micheque collapse are orientated E-W. It is possible to explain this E-W direction as the result of head strike-slip movements along the scar that accommodated dyke emplacement. Further east, the dykes are orientated along a N030 to N050 trend that represents the main rift direction. Several dykes dip towards the north, west and north-west, suggesting that the stress field (i.e. σ_1 , σ_2 and σ_3) is tilted (**Fig.18**). Such a configuration can occur if the ridge flank slowly spreads towards the sea and the progressive displacement is accommodated by the opening of northward, westward and northwestward dipping fractures, parallel to the rift direction. It is thus proposed that Micheque mainly formed by flank creeping. Note that this mechanism has also been suggested to have occurred in the Anaga massif, NE of Tenerife (*Walter et al., 2005*). However, *Walter et al. (2005)* observed that the older dykes dip landwards, which is to be expected if the creeping persists long after dyke emplacement, allowing the dykes, initially dipping seaward, to be rotated. Another possibility to explain the differences between the dips of the dykes is that in Anaga, we are below the décollement level while we are above it in the NERZ. The stress field in these locations may be thus very different and for example, the stress conditions maybe be more influenced by the intrusive complex below the décollement. One can envisage that the dips of the NERZ dykes below the décollement are different, and maybe landward, as in Anaga.

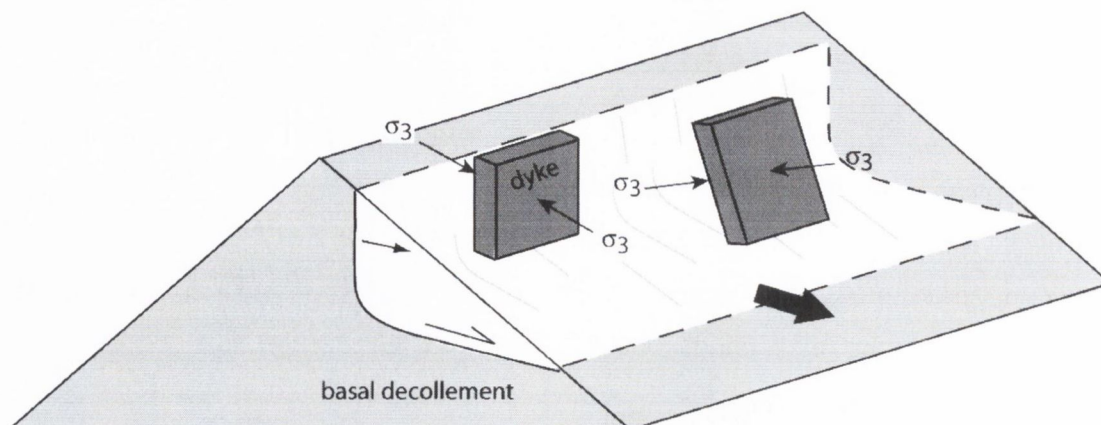


Figure 18: Orientation of the stress field thought to have prevailed in the different collapse areas.

The slow motion of the flank is likely related to intrusive activity (*Dieterich, 1988*). Many vertical dykes also have been recorded at the head of the collapse, suggesting a horizontal minimum stress component (σ_3). We thus have two different minimum stress components: one tilted and one horizontal (**Fig.18**). Unfortunately, it is not possible to determinate their relative age. It can be suggested that deep basal slips accommodate the rift extension, allowing for the coexistence of vertical and tilted dykes (**Fig.18**). Also the effect of the topography needs to be considered, as the stress field will be tilted according to the external ridge slope.

The supposed strike-slip and basal-slip displacements together with a tilted stress field strongly suggest that flank creeping, due to intense intrusion emplacement, lead to the Micheque collapse. It is also conceivable that flank spreading was helped by basal low strength layers, such as powdered and crushed lava flows and associated basal breccias.

6.4.2. Güímar

The N030 and N060 strikes of dykes recorded on the SW side of the head of Güímar collapse represent probably the rift orientation. However, theses dykes intrude the Izaña cinder cone and are located at high altitude, which is probably close to the ancient topographic surface. They are thus interpreted as the feeders of the recent edifice of Izaña and hence their orientations likely represent a local topographic effect (cf. *Delaney and Pollard, 1981; McGuire and Pullen, 1989; Gudmundsson, 2000; Tibaldi, 2003; Klügel et*

al., 2005). Some “along- strike” en-échelon dykes (*Clemente et al.*, 2007) have been observed here. As previously mentioned, those dykes are close to the surface. Consequently, the magma preferentially propagated laterally, generating segmented dykes (*Baer*, 1995).

Note that the dykes in sector F mainly trend N000, but strike directions N140-160 become dominant toward the south. In sector G, the dykes mostly trend N130-150. The majority of the dykes dip towards the E or SE, implying that the minimum stress component σ_3 is tilted. (1) The consistency of the dyke dips (E and SE) together with (2) the progressive change of strike from the north to the south (area F to G) and (3) the fact that the dykes are not always straight are consistent with flank creep towards the sea. The N000 direction found almost at the head of the collapse corresponds to extensional movements, while the NW-SE direction along the side of the collapse corresponds to strike-slip movements (**Fig.19**). The N000 direction is also found in one galeria to the north of area F (*Carracedo et al.*, 2009).

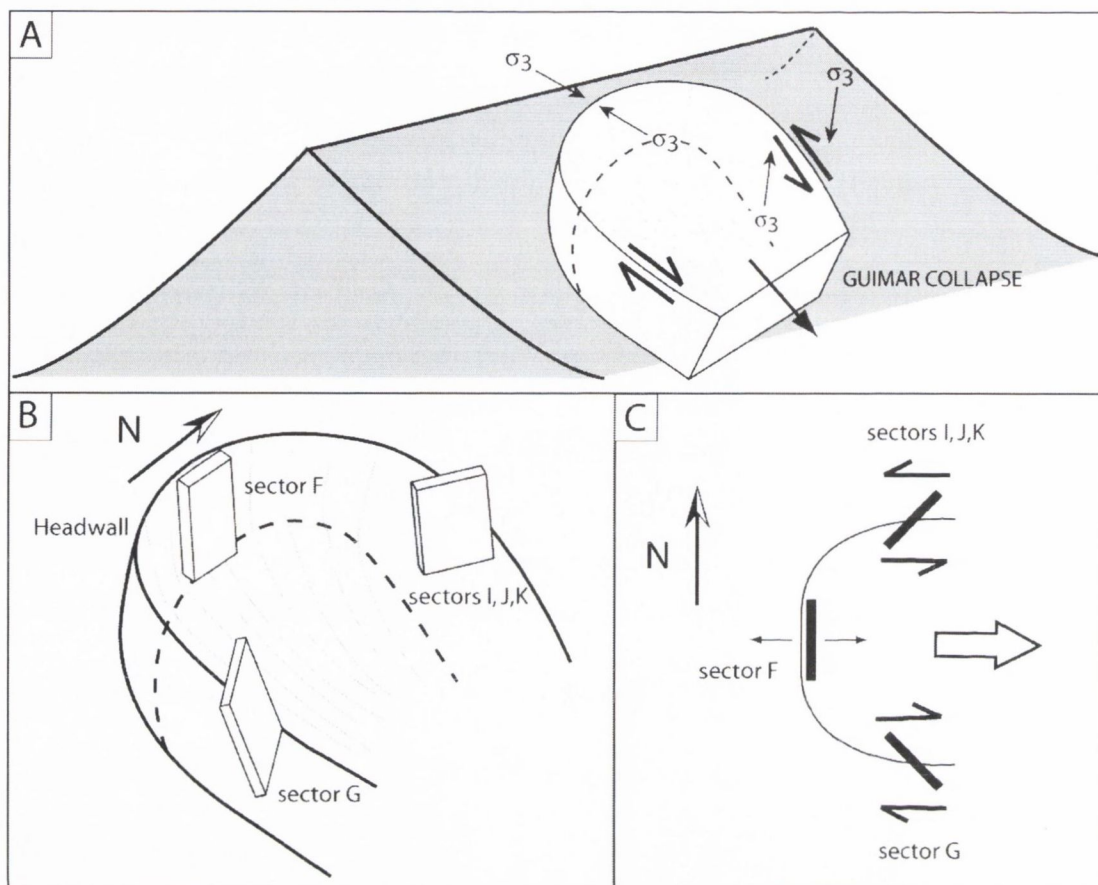


Figure 19: Interpretative sketch of the dyke orientations displayed in Güimar valley. **A:** Stress field orientation and resulting displacements. **B:** 3D view of dyke orientations resulting from the stress field displayed in A. **C:** Same as B but in a 2D plan view.

The NE side of Güimar valley is more complicated than the SW. It seems that there is still more evidence for creep as the dykes dip mainly towards the east, south and southeast. The N050 direction (sector I and J) can either represent the rift trend or the strike-slip movements related to the creeping sector edge (**Fig.13F, 13G, 13H; Fig.19**). However, sector J is outside the scar and shows another trend at N060-N070. This direction could represent a deviation of the rift. Note that the direction N070-N090 is also found in sector D (Micheque). Large deep vertical canyons are observed in both sectors I and J. They are more or less parallel to the scar (i.e. NW-SE with some kinks) and may correspond to major faults. One of them is interpreted as a fault mirror plane at N134, partially filled by a lava flow that also could be interpreted as a vertical paleovalley.

The dykes recorded in sector L do not show any specific trend. This sector is almost at the upper part of the rift, so it is probable that the absence of consistent trends is due to the fact that we are close to the surface and the orientations are due to topographic effects (e.g. Walter and Troll, 2003; Tibaldi, 2003). However, the dyke swarms observed in panorama follow mainly the direction of the rift, with only few dykes offset from the trend.

6.4.3. La Orotava

The dykes at the north-east corner of the head of La Orotava valley (sector C) are mainly orientated along a N060 trend. A few major sets of fractures are observed, demonstrating the intense deformation at the head of the collapse (**Fig.14**). The NE-SW trend may correspond to the rift direction. The N-S together with the ENE-WSW trends form a set of conjugate faults accommodating intrusion emplacement by NW-SE extension (**Fig.20A**). Another way of interpreting the N-S and ENE-WSW trends is to consider them as the expression of dextral or sinistral strike-slip movements that accommodate continuous dyke emplacement (**Fig.20B**). Furthermore, a set of conjugate fractures at N000 and N060 related to extension are expected to dip 60° towards the east and towards the NW respectively. However, the faults are subvertical and dip mainly towards the W and NW. It is thus likely that the N000 and N060 fractures are strike-slip faults rather than conjugate faults. The dextral regime seems to have been dominant, considering that we are at the head of the landslide.

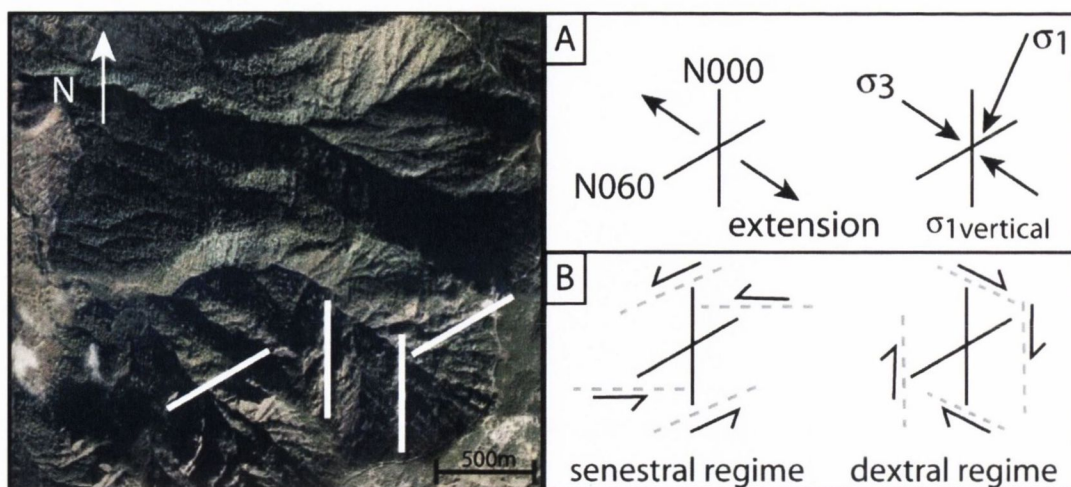


Figure 20: Interpretation of the structural orientations observed in sector C, La Orotava valley. The white lines on the Google Earth image represent the main structural lineaments. The two sketches on the right represent the two possibilities to explain these main directions: **A:** a NW-SE extension or **B:** a strike-slip regime (dextral or sinistral). The field data are more in agreement with a strike-slip regime.

Similarly to the head of Micheque collapse, most of the dykes are vertical or dip towards the north, west and north-west, suggesting a horizontal and a tilted stress field (for possible explanation see **section 6.4.1** and **Fig.18**). La Orotava valley was thus probably partially formed by flank creep. The movements were perhaps less coherent at the base of the valley (keep in mind that the base of the valley does not correspond to the base of the collapse) as the dykes record inconsistent dips, but note that only 10 dykes have been found.

No dykes are found along the two opposite sides of the collapse scar. Sector spreading may have thus been favoured by the presence of basal LSLs in addition to dyke emplacement (Oehler *et al.*, 2005). Several crushed, powdered and weathered lava flows have been found in the vicinity of La Orotava collapse scar and could have acted as LSLs.

6.4.4. The Ridge

The N040 direction recorded along the Boca Del Vallee trail (area K) may represent the rift orientation. No preferential direction has been observed for the dip of the dykes. Towards the west of this trail, we recorded vertical dykes or ones dipping mainly towards the south and south-east, as observed in area I and J. The dip directions can be interpreted as the result of creep. Still at the west of the track, two main directions at N040 and N060 were recorded, N060 is considered as a deviation of the rift. The dykes observed along the road TF24 display a main direction towards the NE and are thus parallel to the rift. However, a large variability in strike and dip is observed. Here, we are here at the upper part of the rift, probably close to the old topographic surface. Thus, the variation of dyke orientation does not reflect any regional tectonic process and can be simply explained by surface effects or by local structural variations (*Tibaldi, 2003; Acocella and Neri, 2009*).

7. Discussion

7.1. Flank creeping

The fractured and crushed dykes found frequently in the studied area indicate that they were still tectonic movements, parallel to their strike, during and/or after their cooling. Two types of movements were observed: (1) basal detachments responsible for a tilt σ_3 and the creation of seaward-dipping fractures and (2) lateral strike-slips responsible of dyke strike orientation (**Fig.21**). These two points are in favour of a slow, progressive and continuous motion of the rift, i.e an episode of prolonged flank creep. This process implies a slow motion (i.e. spreading) of one sector of the edifice (volcano or rift) due to intense intrusive activity, i.e. through spreading of the basal intrusive complex (*Dieterich, 1988, Delcamp et al., in review*), probably along pre-existing low strength layers (*Merle and Lénat, 2003; Oehler et al., 2005; Delcamp et al., 2008*). A few dykes were dated using the Ar-Ar dating technique (**see chapter 2**) and they all are contemporaneous with the several collapse events that affected the NERZ (Guïmar, Micheque and La Orotava). It is thus possible that rotation of dyke occurs before creeping and collapse events or simultaneously as few crushed dykes, probably syn-tectonic, have been observed.

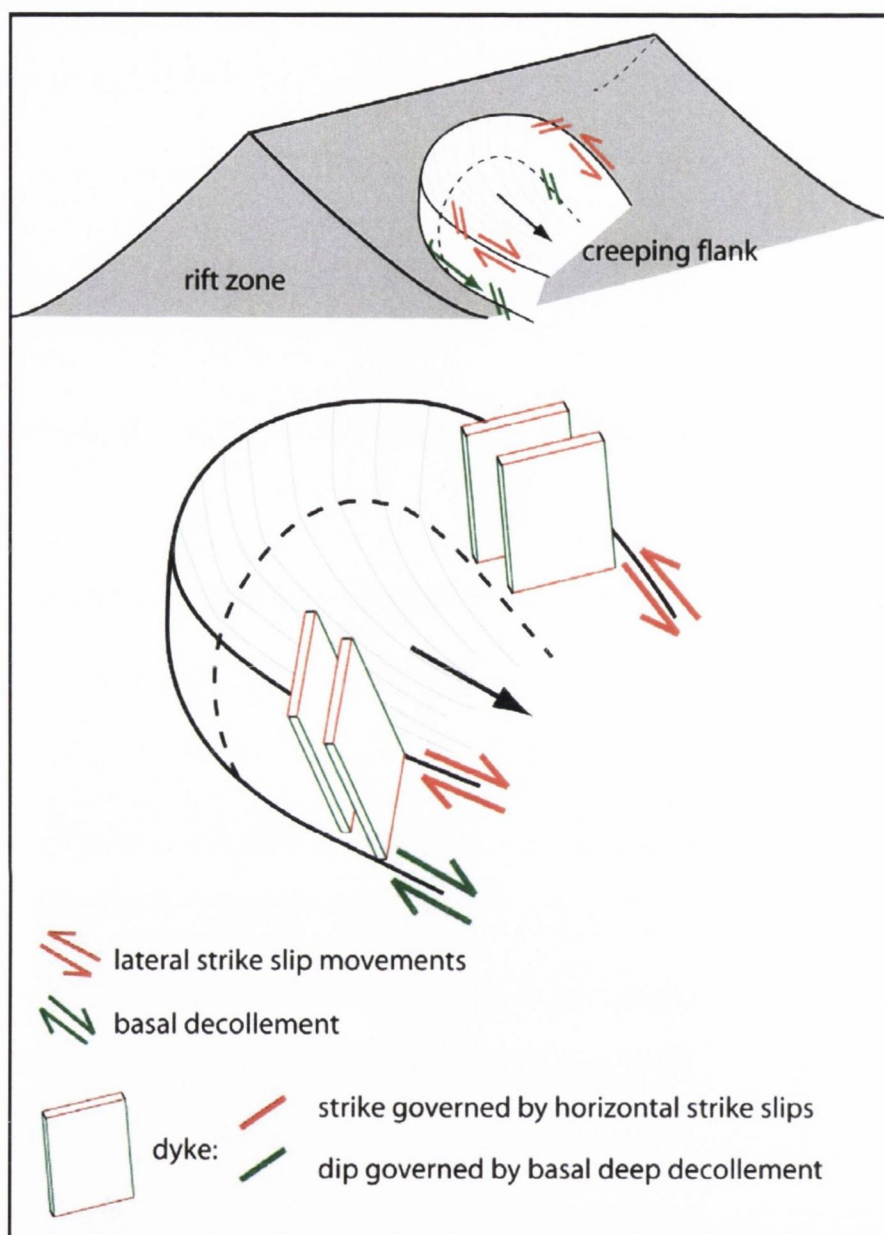


Figure 21: Interpretative sketch showing the relation between basal- and strike-slip fault zones and the resulting dyke orientations.

The dykes are slightly thicker and their density higher at the deepest part of the rift (especially in the galeria Rio de Igueste). The force initiated by the intrusions is thus higher at the base and favours a deep décollement level. Furthermore, dyke intrusion is

accompanied by an increase of pore fluid pressure that decreases the friction coefficient (frictional resistance), and thus favours failure (*Elsworth and Voight, 1996*).

The directions (strike and dip) of the intrusions are more consistent at the deeper compared to shallower levels. This observation reflects a consistent stress field at the base that contrasts with a variable stress field at the top of the rift, likely controlled by the topography (cf. *Tibaldi, 2003*).

The rift was built during three main stages of intense and short activity during the Miocene, the Upper Pliocene and the Pleistocene (*Carracedo et al., 2009*). During these peaks of growth, the intrusive activity was high and almost continuous (*Carracedo et al., 2009*). As the different periods of activity were short (few hundred ka), the rift was able to progressively accommodate the intrusive core and associated dyke emplacement by flank creep, reaching a stabilized state. The last period of activity during the Pleistocene was more intense and the rift was not able to accommodate intrusive core growth and further dyke emplacement by flank creep anymore. Despite the rift being able to reach temporarily a semi-stable state by flank creeping, its height was increasing and its slopes were steepening. Therefore, with each new intrusive event, the rift progressively evolved towards an unstable configuration and eventually a major lateral landslide occurred. Consequently, the initial stage of rift construction may be controlled by regional stress field or pre-existing structures, but progressively, the growing intrusive complex(es) and ridge take control of the near-surface stress field.

7.2. Comparison with other oceanic volcanoes

The creep involves an extension at the head and strike-slip movements on the side of the flank. This process has been also evoked on Piton de La Fournaise, La Réunion Island, by *Carter et al. (2007)*. They observed strike-slip fractures around the crater rim that they interpreted as the expression of a progressive displacement of the area related to flank spreading (*Merle and Lénat, 2003; Carter et al., 2007; Michon and Saint-Ange, 2008*). Several earthquakes in Hawaii are associated with dyke emplacement accommodated by slow flank displacements (*Swanson et al., 1976; Dieterich, 1988; Cayol et al., 2000; Brooks et al., 2008*).

Following the major collapse events on the NERZ, the La Orotava and Güímar embayments were only partially filled, as the intrusive and eruptive rates were relatively low (Carracedo, 1999; **Fig.22A**). On the contrary, Piton de la Fournaise on La Réunion Island is considerably more active (**Fig.22B**). There, the intense intrusive activity is accommodated by flank spreading of the Enclos Fouqué (Merle and Lénat, 2003), but in the meantime, the embayment has been regularly filled by recurrent igneous activity. In Hawaii, intrusive activity is also accommodated by flank displacement (Swanson *et al.*, 1976; Dieterich, 1988; Cayol *et al.*, 2000; Brooks *et al.*, 2008). As well as the Enclos Fouqué, collapse embayments are regularly filled by the high eruptive activity (**Fig.22C**). In Tenerife, the deep collapse scars of Güímar and La Orotava are preserved as the eruptive rate was low (eg. Carracedo, 1999). In Hawaii, the eruptive rate is very high, and the collapses are quickly covered by more recent eruptive products. Micheque can be regarded as an analogue to the Hawaiian case, as the collapse was rapidly filled by subsequent volcanism. La Réunion, in turn, represents an intermediate case.

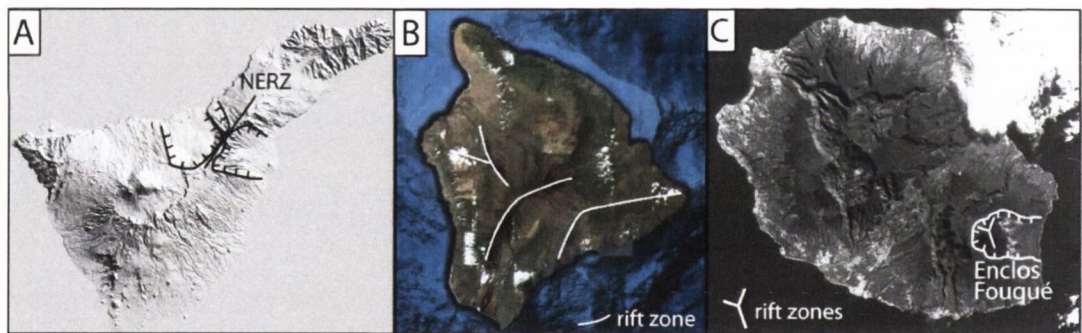


Figure 22: A: Tenerife Island (courtesy R.Paris, source: Grafcan©) B: Hawaii Island (Google Earth image) C: La Réunion Island (Landsat image), and associated rift zones and collapses.

7.3. Dyke propagation

Evidence of dyke propagation through pre-existing fractures but also by hydro-fracturing has been found. The country rock surrounding the dyke is not deformed and open fractures have been observed in La Orotava collapse headwall (area C) and in the visited galerias. A lot of dykes become thicker while propagating through weak layers and are

thinner in competent layers (mainly lava flows). All these observations support a mode of propagation using pre-existing fractures. The magma will dilate an older fracture more easily when it propagates through the low competence layers. Note that the apparent absence of deformation within the surrounding country rock might be due to strain dissipation within the low competence layers (lava flow basal breccias, lapilis or colluvium). This strain dissipation is essential for the accommodation of the deformation and thus the stabilization of the rift. On the other hand, incorporation of country rock has been sometimes observed, as well as tension gashes along several dyke margins involving shear movements. These two last observations favour the hypothesis of dyke propagation by hydrofracturing. As previously mentioned, thickness variation have been noticed when the dykes intrude layers of different competence. This can also imply dyke propagation by hydrofracturing. The two mechanisms of emplacement are thus inferred from field data.

The variation of dyke thickness between low and high altitude might be related to their capacity to propagate. At low altitude and thus deep levels, the dykes will have more difficulty to propagate considering the confining pressure. Being close to the source, the driving pressure is higher and the dykes are thus thicker. When they approach the surface, their driving pressure is lower, and dykes are thus thinner.

7.4. Chronology of the collapse events

The relative chronology between the different collapses deduced from stratigraphy and geochronology data by *Carracedo et al. (2009)* is consistent with our structural data. The rift was accommodating the growing intrusive core and dyke emplacement by flank creeping that, in turn, favoured more dyke intrusion (positive feedback). The rift was becoming steeper and eventually was not able to accommodate further intrusions. Consequently, a sector of the rift collapsed, forming the Micheque embayment. Following this mass-wasting event, the scar was rapidly filled by eruptive products. The other side of the ridge, already in an unstable state, was weakened by this major event and responded by another lateral collapse, creating the Güímar embayment. Activity of nested volcanoes continues to fill the Micheque embayment and has started to fill Güímar. No dykes have been found along the scar of La Orotava collapse (except at the

head). Therefore, this last collapse is probably not intrusion-driven, but probably largely gravity controlled, as this part of the NERZ was weakened by previous intrusion-collapse episodes. The destabilization events could have been helped by LSLs such as scoriae or partially deformed lava flows, or even by the land-sea interface. Note that the field data do not show if one scar corresponded to one or several collapse events. However, it can be suggested that different blocks are sliding at different rates, leading to a set of strike-slip faults and collapsing sectors. The difference between the various sliding rates would be directly linked to the intrusive density for the cases of the Micheque and Güímar collapses.

8. Conclusion

The rift zones are widely considered as narrow and well-defined structures, with a uniform dyke and fracture orientation pattern. Our field data from Tenerife show that a rift can be rather diffuse, with several intrusive batch events separated by periods of quiescence, deformation, erosion and collapse.

Forceful dyke emplacement favours sector spreading of the ridge, resulting in flank creeping. This mechanism involves progressive displacements accommodated by extension and strike-slip movements at the head and along the walls of the creeping sector, respectively. Flank creeping involves basal detachment that controls the dyke dips and lateral strike-slip zones that control dyke strike. Accommodation of dyke emplacement by creeping results in flank steepening. Further intrusive events will continue to steepen and eventually destabilize the flank, triggering collapse event(s). Consequently, flank creeping constitutes a main mechanism for initial rift stabilization, but also for destabilization and recurrence. In the North-East Rift Zone, the two first collapses are thought to be essentially intrusion-driven (Micheque and Guímar), while the last one (La Orotava) seems mainly gravity-driven.

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APPENDIX 1: X-RAY DIFFRACTION (XRD)

Several samples of country rocks (lava flows and their basal breccia) have been collected in the two galerias explored during the field campaign in February 2007. They were analysed in Trinity College Dublin using X-Ray Diffraction to check for the presence of clays. It appears that the sampled rocks contain only a low amount of clays, such as illite, or expandable clays, such as smectite. Below, a few graphs are shown.

For each sample, two graphs of XRD analyses are displayed, corresponding to two different methods of sample preparation. The x-axis corresponds to the angle in degrees 2θ (cf. Bragg equation) and the y-axis to the diffracted X-ray intensity (termed “counts”). For the first graph, a classic clay separation has been performed, while for the second graph, clay separation was done using glycol to separate better the expandable clays. The peak obtained after clay separation by glycol should be offset from the one obtained after classic clay separation.

The peak omnipresent at 33° corresponds to contamination.

SAMPLE TF-G1-CR-D4

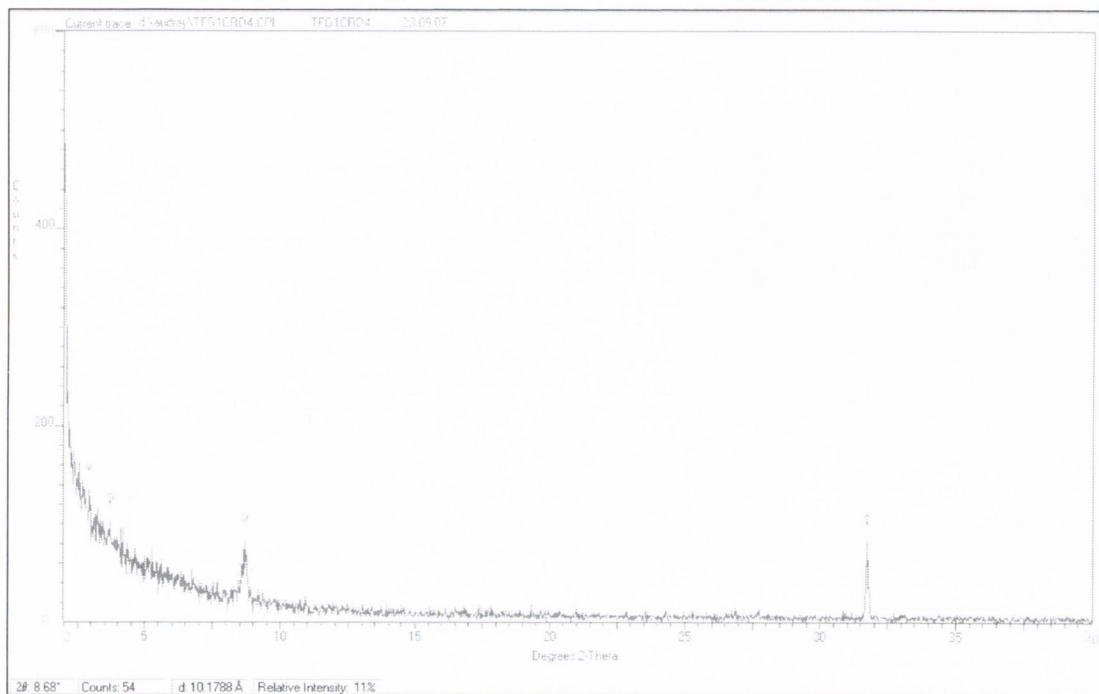


Figure 1A: XRD graph for the sample TF-G1-CR-D4; classic clay separation.

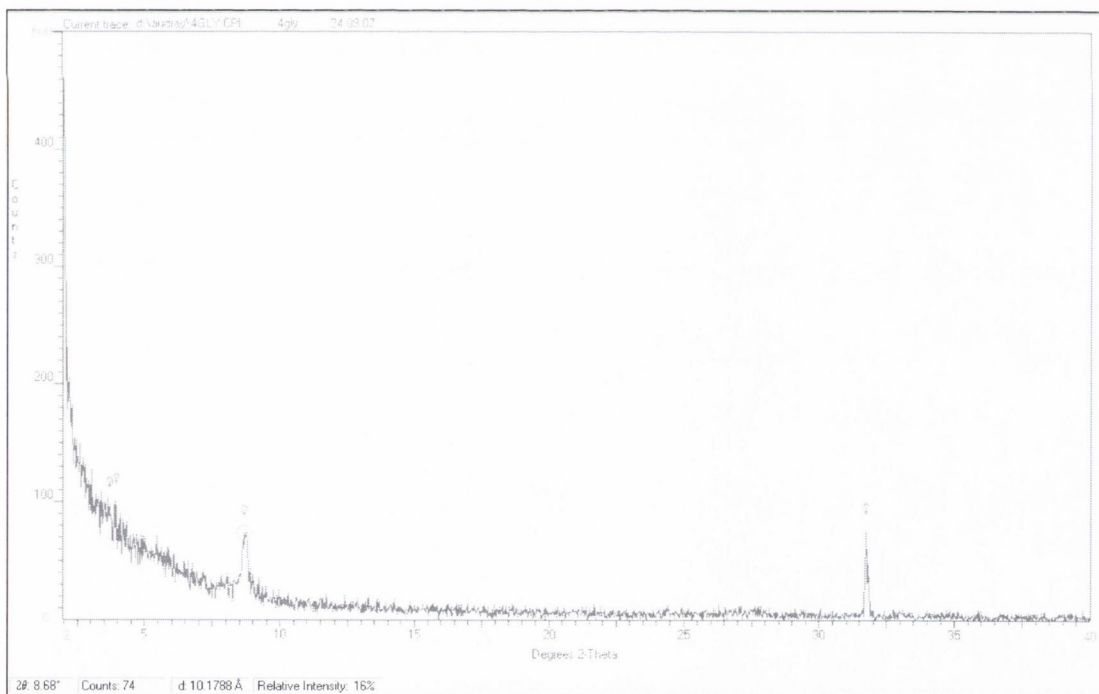


Figure 1B: XRD graph for the sample TF-G1-CR-D4; clay separation using glycol.

The peaks obtained may correspond to illite or montmorillonite, the last one being an expandable clay.

SAMPLE TF-G1-CR-D5

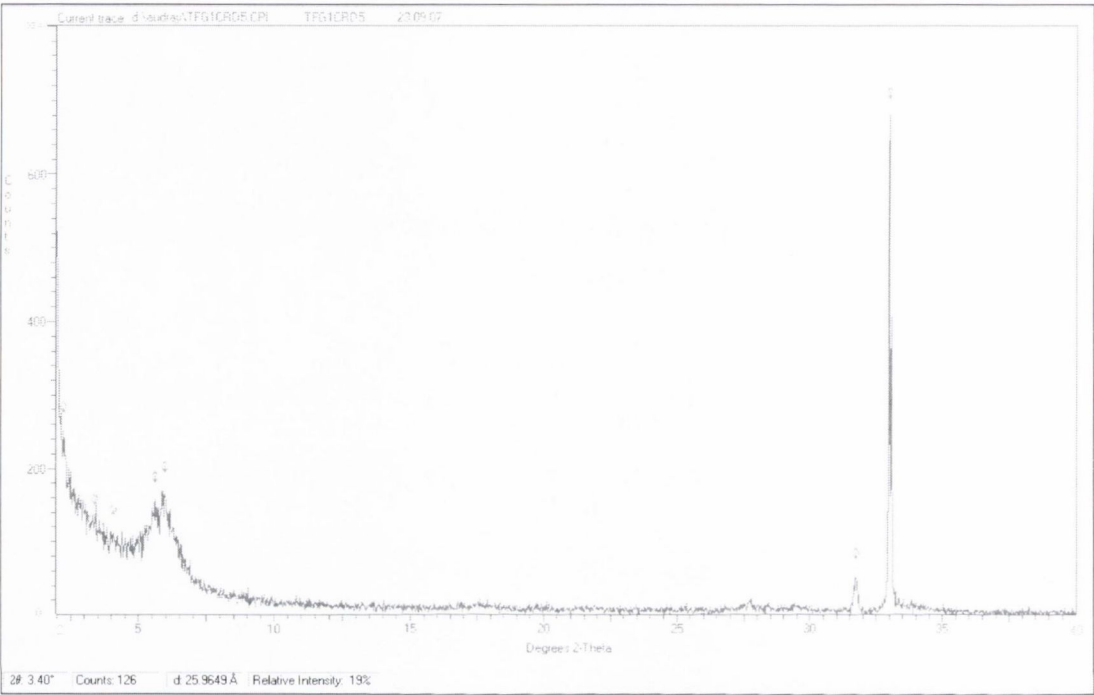


Figure 2A: XRD graph for the sample TF-G1-CR-D5; classic clay separation.

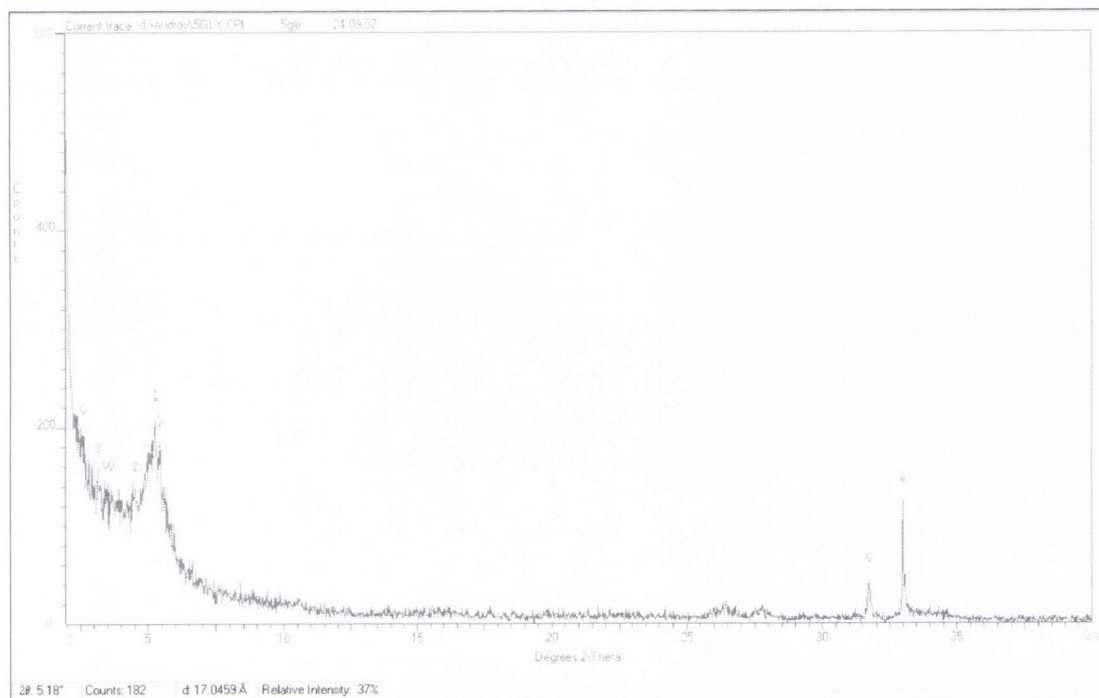


Figure 2B: XRD graph for the sample TF-G1-CR-D5; clay separation using glycol.

The peaks may correspond to montmorillonite.

SAMPLE TF-G2-CR-D4

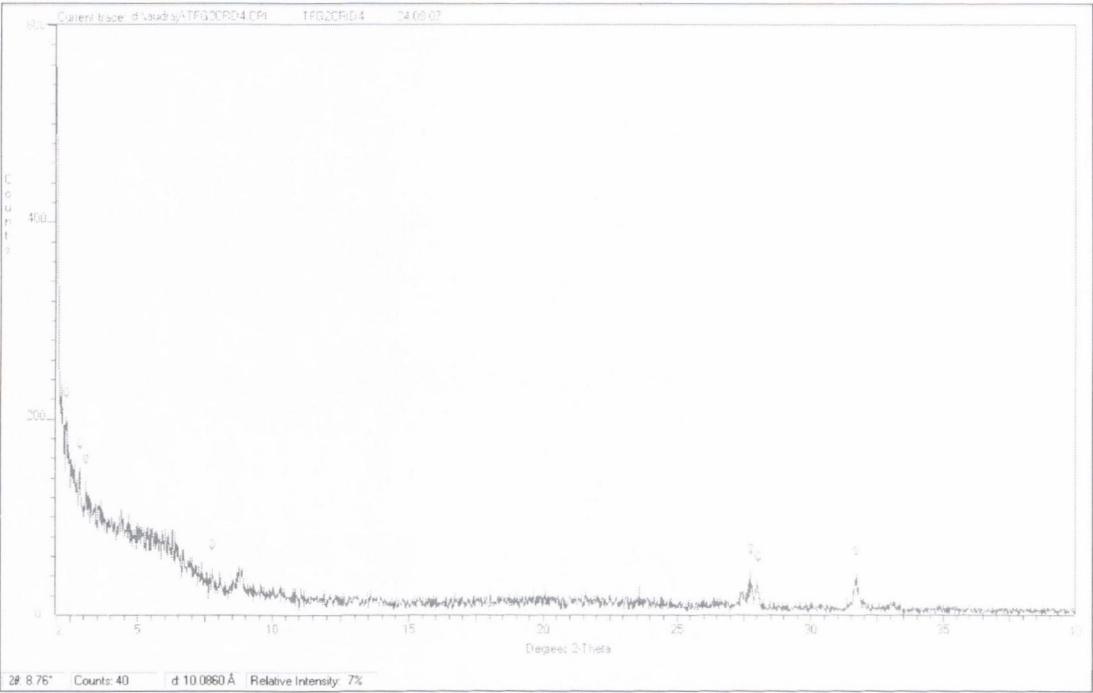


Figure 3A: XRD graph for the sample TF-G2-CR-D4; classic clay separation.

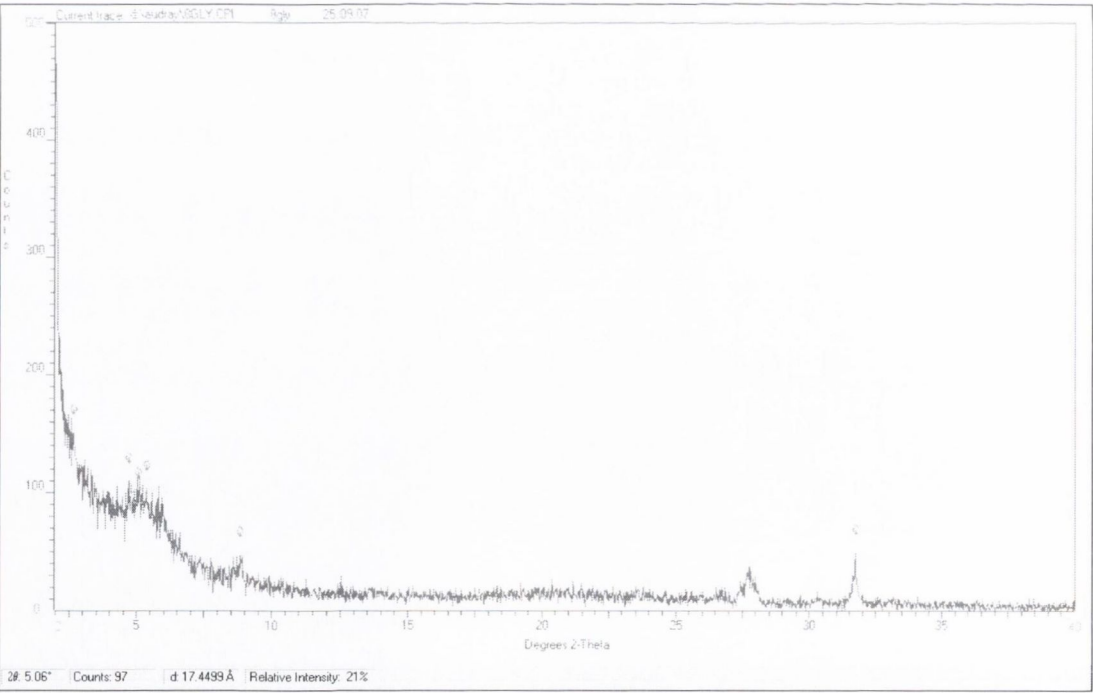


Figure 3B: XRD graph for the sample TF-G2-CR-D4; clay separation using glycol.

The wide peak observed on the second graph may correspond to expandable clay(s) such as montmorillonite and smectite.

SAMPLE TF-G2-CR-X2

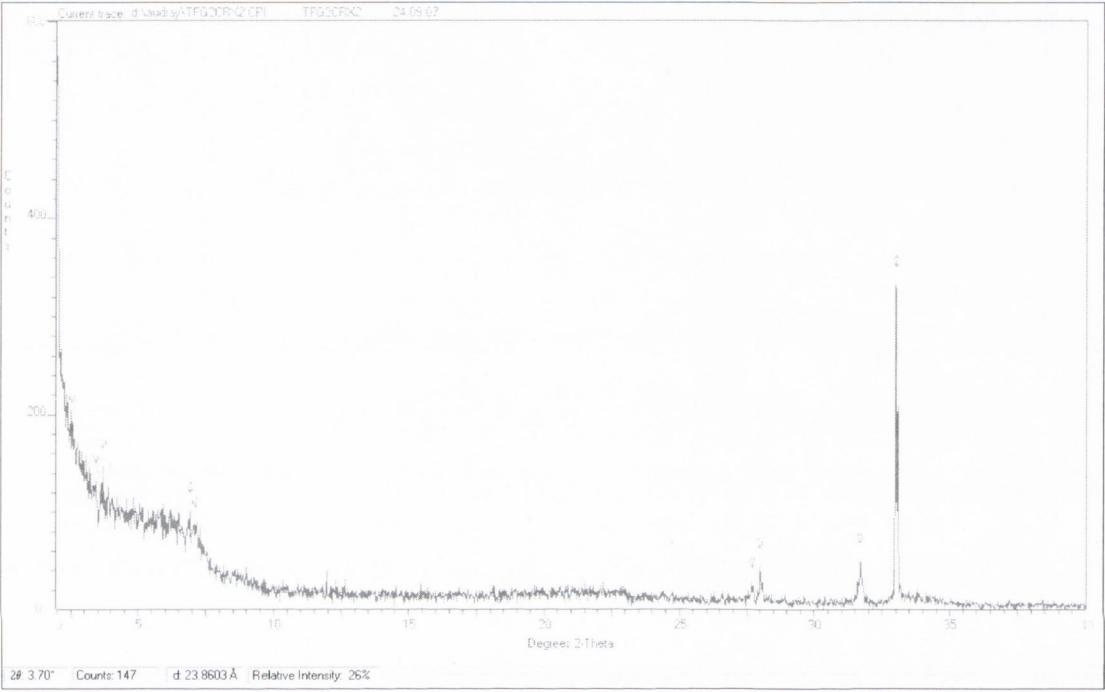


Figure 4A: XRD graph for the sample TF-G2-CR-X2; classic clay separation.

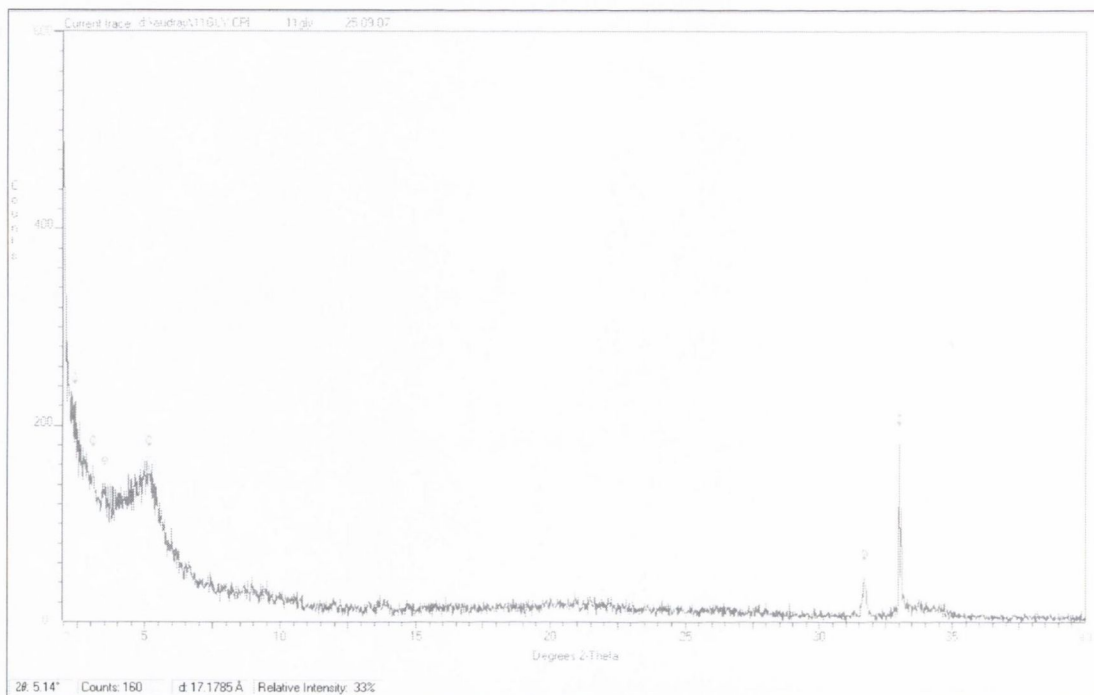


Figure 4B: XRD graph for the sample TF-G2-CR-X2; clay separation using glycol.

The peak observed on the second graph could correspond to expandable clay(s) such as montmorillonite.

SAMPLE TF-G2-CR-AK

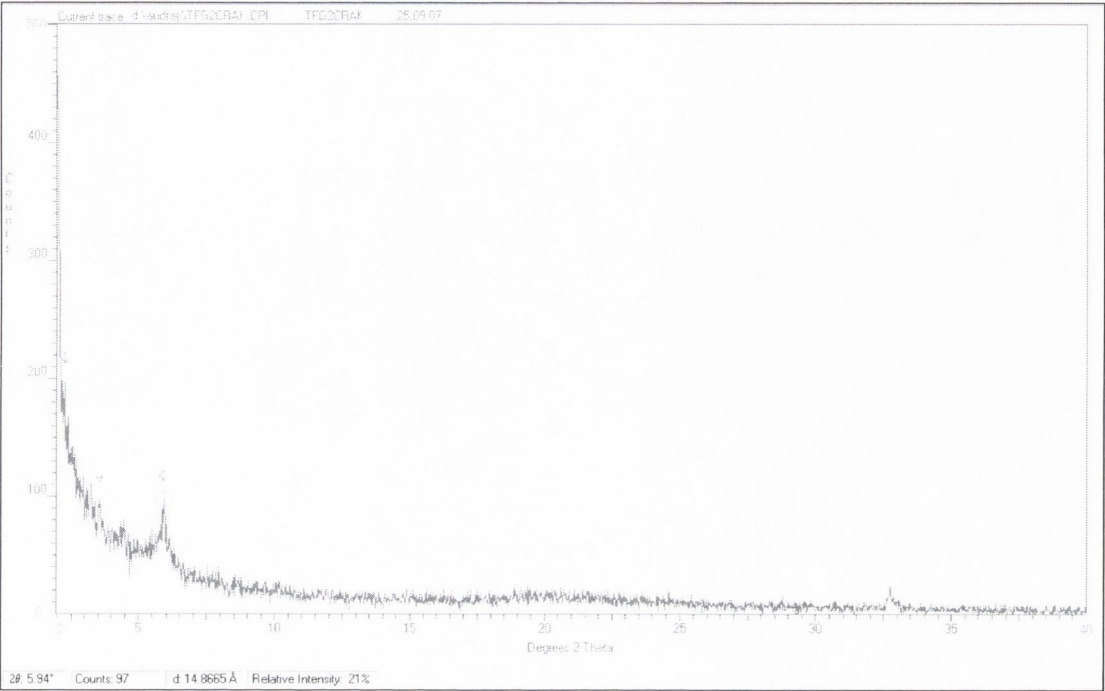


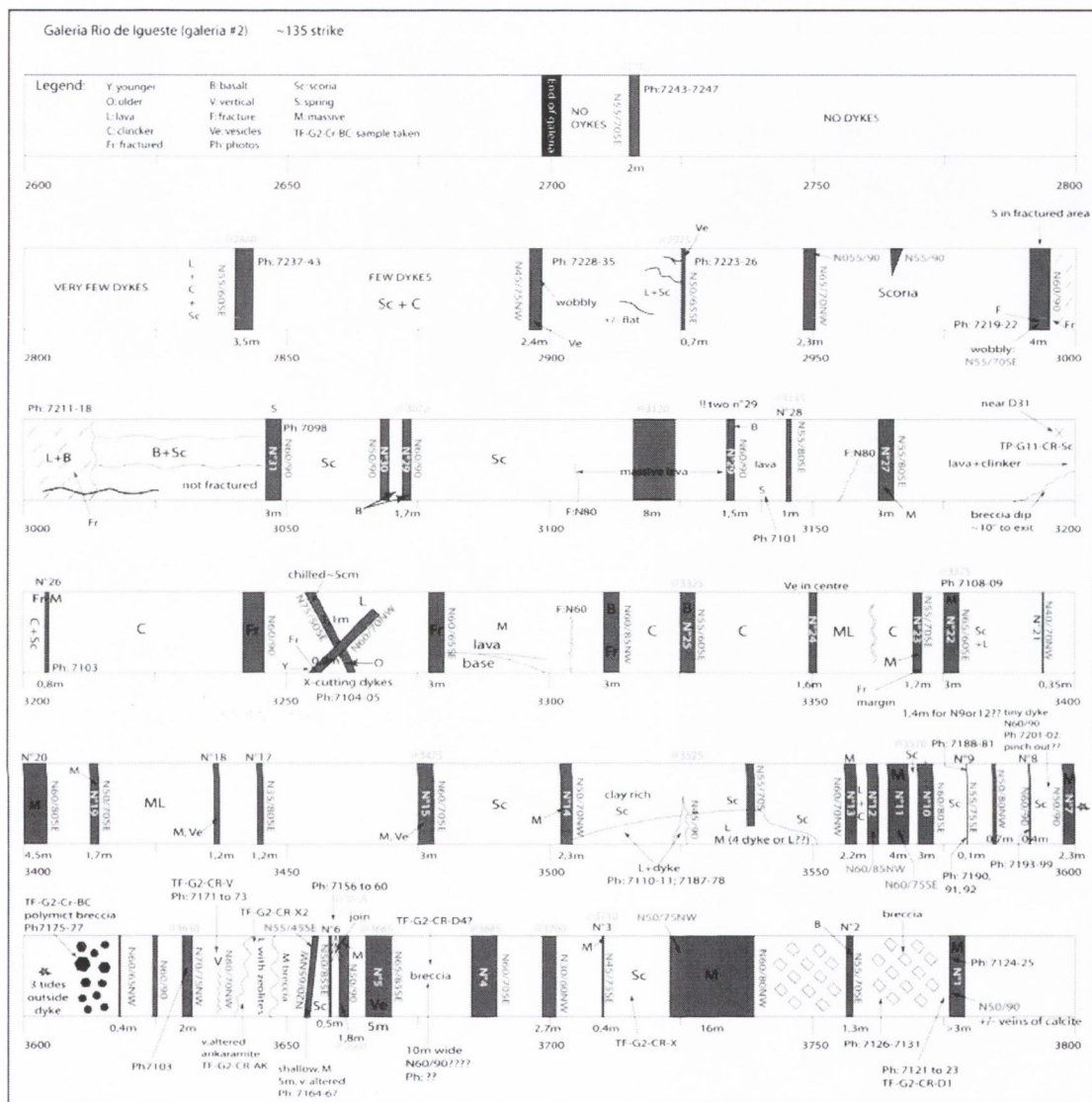
Figure 5A: XRD graph for the sample TF-G2-CR-AK; classic clay separation.



Figure 5B: XRD graph for the sample TF-G2-CR-AK; clay separation using glycol.

The peak at $\sim 7^\circ$ on the first graph is offset to $\sim 5^\circ$ on the second graph and corresponds to montmorillonite.

APPENDIX 2: CROSS SECTION OF THE GALERIA “RIO DE IGUESTE”



CHAPTER 2: PALEOMAGNETISM

Abstract

Paleomagnetic sampling sites were established on 82 dykes along a 8km long section of Tenerife NE rift-zone (NERZ). Of the 70 interpretable sites, 16 are of normal polarity and 54 of reversed polarity. For the grand mean calculation, 19 sites were excluded due to scattered data or because they represent transitional fields. After inverting the reverse polarity sites through the origin, the in-situ grand mean yields a $D=23.8^\circ$, $I=42.3^\circ$, $\alpha_{95}=3.2^\circ$, $K=39$, $N=51$. These values are discordant to the expected late Miocene to Pleistocene magnetic field ($D=357.6$, $I=38.8$, $A_{95}=4.7$). This discordance can be explained as either a 26° clockwise vertical axis rotation or a 28° WNW-side-down tilt about an average N009 horizontal tilt axis. The sampled section is composed of numerous semi-vertical dykes cutting mainly lava flow units that are sub-horizontal and cross-cut by steeply dipping faults (70° - 90°). Field evidence is more consistent with a vertical-axis rotation rather than a tilt of the drilled dykes. Clockwise vertical-axis rotation is likely related to lateral movements that accommodate the emplacement of the underlying intrusive core and associated dykes. This study therefore reveals a long-lived, multi-faceted history that shaped the NERZ. The intrusive activity in the sampled section of the rift has been almost continuous, with several intrusion pulses probably related to flank destabilisation event(s) during the late Pliocene to early Pleistocene.

1. Introduction

1.1. Brief introduction to paleomagnetism concepts (adapted from *Butler, 1992; Dunlop and Özdemir, 1997*).

A magnetic field exerts a force on a positive magnetic charge. When a material is exposed to the present-day magnetic field (H), it acquires an induced magnetisation (J_i) with an intensity value that depends on the magnetic susceptibility of the material (K):

$$J_i = KH \quad (\text{eq. 1})$$

In addition, some materials can also possess a remanent magnetisation (J_r) resulting from the acquisition of the past magnetic field and this is of major interest for paleomagnetic studies.

The Earth's magnetic field (H) can be divided into vertical and horizontal components (**Fig.1**). Two components are defined as followed:

-Inclination (I): corresponds to the angle between the horizontal component (H_h) and the magnetic field H .

-Declination (D): corresponds to the angle between the horizontal component (H_h) and the geographic North.

The aim of this paleomagnetic study is to access the declinations and inclinations that are preserved in the dykes of the North-East rift-zone, to assess the timing of emplacement and any post-emplacement tectonic deformation of the rift.

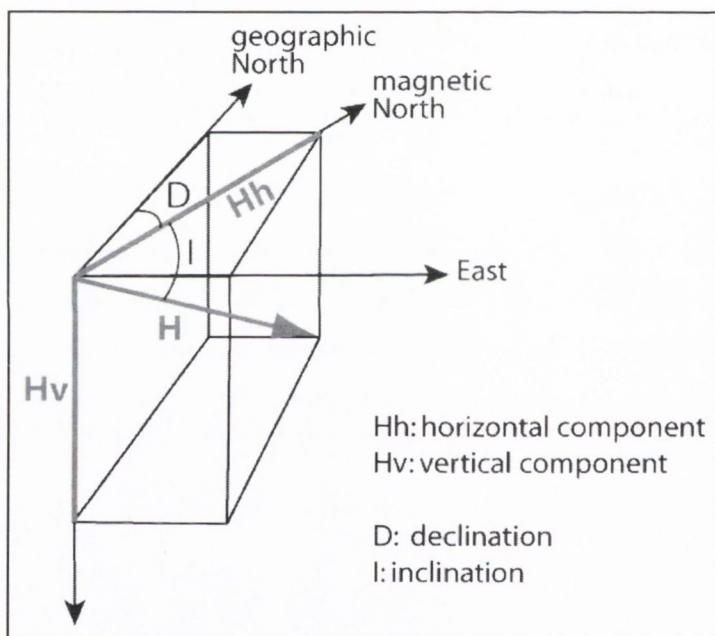


Figure 1: Separation of the magnetic field (H) into horizontal (H_h) and vertical (H_v) components (modified from Butler, 1992).

The Earth's magnetic field is modelled as a geocentric axial dipole. In this model, the magnetic field is produced by a single magnetic dipole placed at the centre of the Earth and aligned with the Earth's rotation axis. The present geomagnetic field is more complex than this model. An inclined geocentric dipole is thus preferentially used (**Fig.2**). This model is inclined to the rotation axis with an angle of $\sim 11.5^\circ$. The actual configuration of this dipole, i.e. positive inclination in the northern hemisphere, is

referred to as normal polarity. A reverse polarity configuration corresponds to a negative inclination in the northern hemisphere.

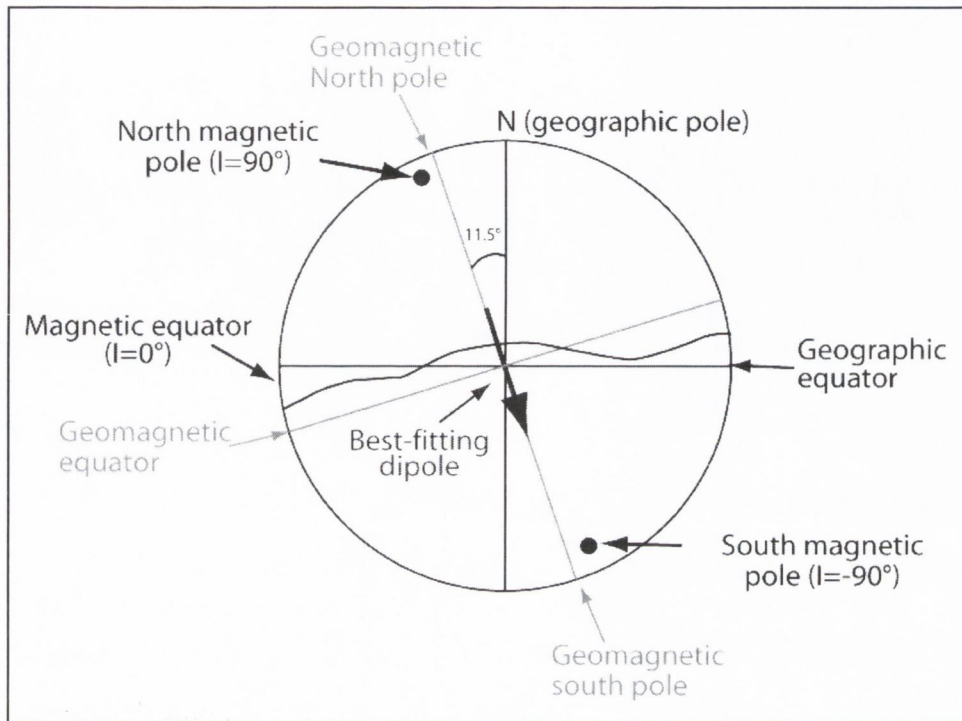


Figure 2: Inclined geocentric dipole model; in this model, the geomagnetic poles and equator are inclined from the geographic poles and equator by 11.5°. The grey arrow represents the best-fit-dipole model (from Butler, 1992).

Most solids exhibit different behaviour when they are subject to a magnetic field:

- a diamagnetic solid will acquire a small induced magnetisation J_i opposite to the applied magnetic field. As soon as the magnetic field is removed, J_i reduces to zero. Example: quartz.

- a paramagnetic solid will acquire a small induced magnetisation parallel to the magnetic field but similar to a diamagnetic material, J_i decreases to zero when the magnetic field is removed. Examples: olivine, biotite.

- a ferromagnetic solid will acquire an induced magnetisation parallel to the magnetic field. These solids are of major interest for paleomagnetic studies as after removal of the magnetic field, they do not reduce to zero but keep a record of the applied magnetic field. Examples: hematite, magnetite (**Fig.3**).

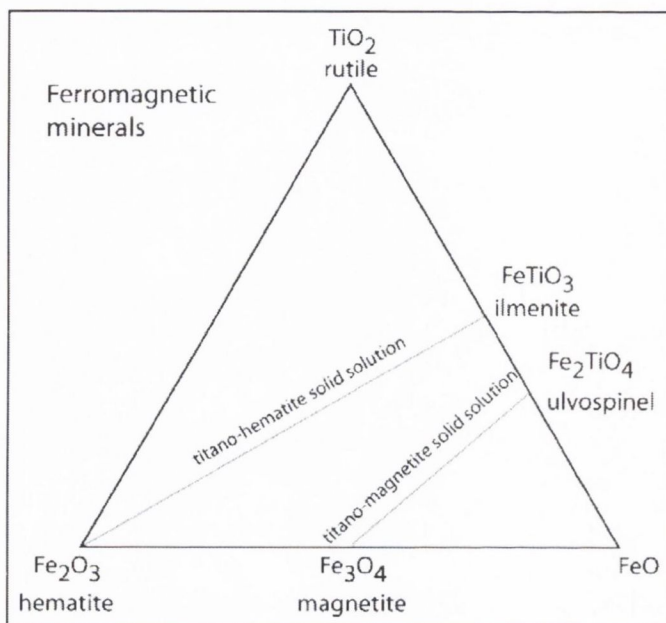


Figure 3: Tertiary diagram of ferromagnetic minerals (modified from Dunlop and Özdemir, 1997).

Figure 4A represents a spherical ferromagnetic grain with a uniform magnetisation. The magnetic charges at the surface induce magnetostatic energy due to the repulsion between the adjacent charges of similar sign. This type of grain is referred as Single-Domain (SD). To decrease the magnetostatic energy, the grain is subdivided into several magnetic domains (**Fig.4B**). Indeed, this allows to have a smaller surface covered by the magnetic charges and to bring charges of opposite sign closer. These grains, with a diameter > 10µm, are referred as Multi-Domains (MD). An intermediate type of grains, named Pseudo-Single Domains (PSD), exists but the physical properties are not fully understood. They can be important carriers of remanent magnetisation.

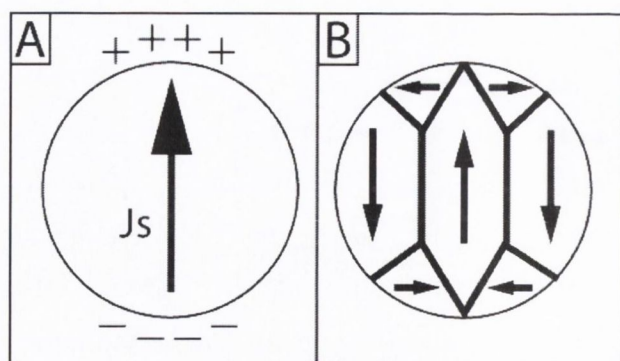


Figure 4: Representation of magnetic domains: the direction of saturation magnetisation (J_s) is represented by the arrow. **A:** Single Domain grain (SD); the magnetic charges are shown by plus and minus signs. **B:** Multi Domain grain (MD). After Butler (1992).

Most geologic materials, prior to laboratory treatment, contain a Natural Remanent Magnetisation (NRM). The NRM corresponds to a record of the geomagnetic field during rock formation and its following history and is thus composed of potentially multiple magnetisation components. The primary NRM is the component acquired during rock formation. In most cases, it is the component of interest for paleomagnetic studies. Secondary NRM components are acquired post rock formation and overprint the primary NRM. The NRM can be thus seen through the following equation:

$$\text{NRM} = \text{primary NRM} + \text{secondary NRM} \quad (\text{eq.2})$$

Three forms of primary NRM are recognized:

- Thermoremanent magnetisation that is acquired during rock cooling from above the Curie point. Above the Curie temperature any material has a paramagnetic behaviour and below this temperature, the magnetic moments are aligned with the ambient magnetic field.

- Chemical remanent magnetisation that is acquired during growth of ferromagnetic grains below the blocking temperature. The blocking temperature (T_B) corresponds to the temperature where a ferromagnetic grain is not stable anymore and loses its remanence.

- Detrital remanent magnetisation, acquired during deposition of sedimentary rocks containing detrital ferrimagnetic minerals.

In this paleomagnetic study, we want to access the primary NRM acquired by the dykes during cooling. By partial demagnetisation procedures, such as alternating-field demagnetisation and thermal demagnetisation (discussed below), components of the NRM are removed. The components easily removed are referred to as low-stability components, and usually correspond to a secondary NRM. By removing these low-stability components, we slowly isolate a stronger signal referred to as a high-stability component, usually interpreted to be the primary NRM. However, it can occur that the high-stability component does not correspond to the primary NRM. We thus generally refer to the high-stability component as the characteristic remanent magnetisation component of NRM (ChRM). Additional information is then necessary to determine if the ChRM corresponds to the primary NRM.

In the present study, we used two methods to isolate the ChRM: alternating-field (AF) demagnetisation and thermal demagnetisation. For AF-demagnetisation, the sample is exposed, in a stepwise fashion, to an alternating magnetic field (H_{AF}) (**Fig.5**). As an example, we define a magnetic field at point 1 with a magnitude of 10mT and we arbitrarily define this direction as “up” (**Fig.5B**). Magnetic moments of all grains in the sample with a microscopic coercive force $h_c \leq 10\text{mT}$ will be forced to point in the up direction. The microscopic coercive force (h_c) is a measure of the energy that resists the rotation of j , magnetisation of individual ferromagnetic particles, to an external field. If the magnetic field increases by 5mT every half period, the field at point 2 will be 15mT in the “down” direction. Thus all the grains with $h_c \leq 15\text{mT}$ will be forced to point in the down direction. Then, the magnetic field increases until 20mT in the up direction, where all the magnetic moments of the grains with $h_c \leq 20\text{mT}$ will be forced to point upward (point 3). Consequently, from stage 1 to 3, the magnetic moments are progressively randomized. Finally, after AF-demagnetisation, only the ChRM carried by the grains with $h_c \geq H_{AF}$ will remain.

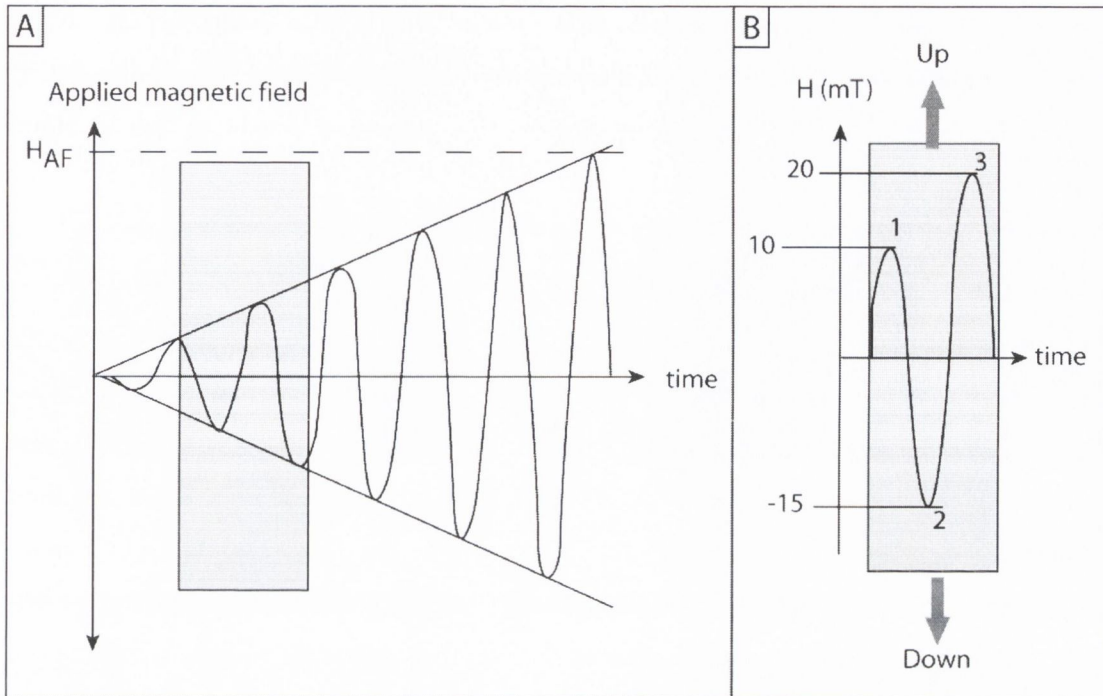


Figure 5: Representation of alternating-field (AF) demagnetisation. **A:** The applied magnetic field (in mT) is a sinusoidal signal that increases through time, from a maximum amplitude at H_{AF} . **B:** Zoom of the grey box in 5A. The peak field at point 1 is 10mT, and as the half period of the signal is 5mT, the peak field is 15mT at point 2 and is 20mT at point 3. Modified from Butler, 1992.

During thermal demagnetisation, a sample is heated to a temperature below the Curie temperature of the ferromagnetic minerals that constitute the sample. Then, the sample is cooled until room temperature in a zero magnetic field. All the grains with blocking temperature $T_B \leq T_{demag}$ acquire a random magnetisation in a zero field, effectively removing the remanence carried by these grains.

There are several rock magnetic experiments that allow the characterisation of the magnetic mineralogy, such as isothermal remanent magnetisation (IRM) and the investigation of continuous susceptibility vs. temperature. During an IRM experiment, we measure the remanent magnetism after a short exposure of a sample to an increasingly strong magnetic field generated by an impulse electromagnet. We thus proceed with IRM on samples that were previously fully demagnetised by AF-demagnetisation. Another

method to determine the magnetic phase present within a sample, consists of continuously measuring the magnetic susceptibility while the sample is heated to 700°C and then cooled to room temperature. These experiments are conducted in an argon atmosphere to reduce the oxidation effects.

1.2. Field area.

The present study uses paleomagnetic data to provide new constraints on magma emplacement, rift formation and rift evolution related to ancient and recent collapse events. Our attention is focused on the dykes of the North East Rift Zone (NERZ), also referred as the Dorsal de La Esperanza (**Fig.6**). The dykes intrude lava flows and their associated basal breccias and colluvium. The lava flows on the Dorsal de La Esperanza mostly dip gently between 0 to 20° towards the NW but some occasionally dip 20-40° NW. These upper flow units are related to the old Arafo volcano and their orientations are consistent with this old volcano flank. Three major collapses are on each side of the NERZ. At the north, the Micheque landslide has been recognized as the oldest and has been estimated at 830ka (*Carracedo et al., 2009*). At the south of the NERZ, the Güímar collapse event was probably contemporaneous to the Micheque event as *Carracedo et al. (2009)* estimated this event to have taken place at 830ka. The La Orotava collapse is to the north of the NERZ and is bracketed between 690 and 566ka. The dykes sampled in this study cover two polarity intervals. As the last transition occurred 780ka ago (Matuyama-Bruhnes), the reverse dykes are at least older than 780ka. The normal polarity dykes are likely younger than 780ka, although it is technically possible that the normal polarity dykes belong to the Jaramillo (1072ka-988ka) or even Gauss period (3.4Ma- 2.48Ma). Six dykes were dated using Ar-Ar method allowing the age of a few intrusive events to be constrained. They all seem to be contemporaneous or predate the several collapse events. They thus belong to Matuyama and Bruhnes periods.

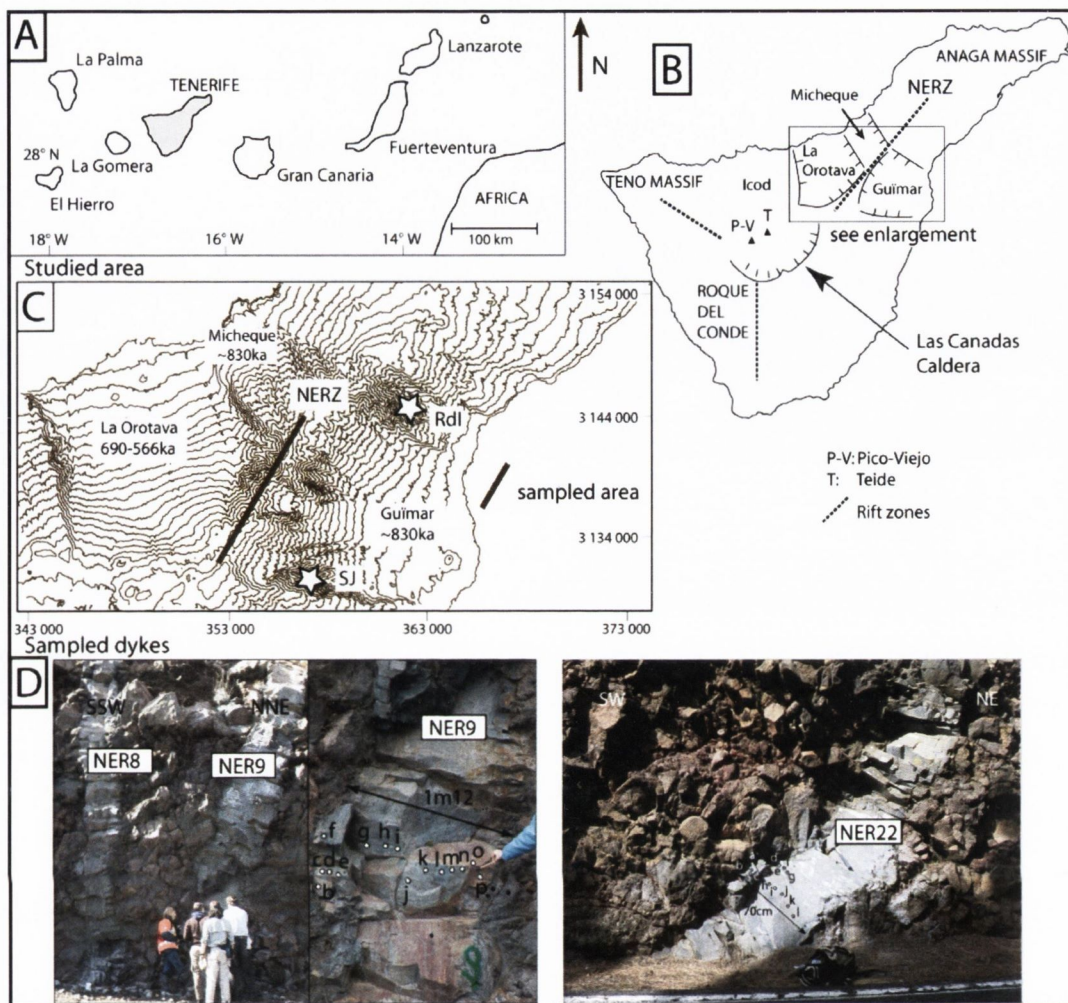


Figure 6: **A:** Tenerife location in the Canary Archipelago. **B:** Simplified structural map of Tenerife with its characteristic rift zones arranged in a triple-armed pattern. Major collapse embayment(s) bound each side of the rift. **C:** Enlarge map of the studied area that focuses on the North East Rift Zone (NERZ) where two major collapses have been recognized, La Orotava valley to the North and Guimar valley to the South. Note that a recent study (Carracedo et al., 2009) recognized a third collapse event -Micheque- to the east of La Orotava. Rdl: galeria Rio de Igueste (400m a.s.l); SJ: galeria Aguas de San José (1130m a.s.l) **D:** Field pictures of three sampled dykes intruding rift zone lava flows.

Road cuts along the NERZ display fresh outcrops and a near parallel section along the dorsal ridge intrusive suite. In total, 82 dykes, corresponding to a total of 943 samples, were drilled along the TF24 road. In addition, 9 dykes (105 samples) were drilled and 19 orientated blocks were sampled within the two galerias (i.e. water tunnels) San José and Río de Igüeste (see **introduction and chapter 1**). Unfortunately, all the samples collected in the galerias did not give interpretable results due to strike orientation errors associated with the surrounding volcanic rocks. To complement these data, 9 lava flows and 2 dykes were sampled on the eastern part of the island in the old Anaga massif, a supposedly stable area unaffected by recent collapse events (<1Ma).

The mineral content of the mafic dykes is variable and we distinguished six petrographic groups, using 82 hand samples and 47 thin sections (see **chapter 1**). All the thin sections have a cryptocrystalline groundmass composed by glass, feldspar (sometimes in needles), olivine and/or augite in different proportion. All the thin sections are characterized by relatively high oxide content with up to 40% phenocryst content. Magnetite was the main opaque phase identified using reflective light microscopy. The six groups are:

- aphyric/subaphyric with little to no phenocrysts (the largest one).
- feldspar rich with rare olivine and/or pyroxene in some samples.
- feldspar and clinopyroxene (augite) group (roughly the same amount of the two minerals).
- ankaramitic dykes, i.e. the amount of clinopyroxene is higher than the amount of olivine or feldspar.
- olivine and pyroxene group.
- pyroxene group.

2. Method

2.1. Field and Analytical Methods

In total, 1048 orientated samples from 91 sites (see **Fig.6C** and **appendix 1**) –one site corresponding to one dyke- were collected using a portable gasoline-powered drill fitted with a non-magnetic diamond drill bit, and 19 orientated blocks were sampled in the galerias. At most sites, between 9 to 11 samples were collected from each dyke to ensure robust statistics and thus the reliability of our results (see **Fig.6D**). All samples were

orientated using a magnetic and, when possible, a sun compass. For most of the sites, the values obtained from the two orientation methods yielded similar results. However, for some strongly magnetised sites (typically ankaramitic dykes; eg. NER 7, 37, 37A), the sun correction was used. All the specimens were then cut into cylinders of $2.2\text{cm} \times 2.5\text{cm}$ using a non-magnetic saw. At least 9 specimens (from different samples) per site were fully demagnetised at the paleomagnetism laboratory of the University of New Mexico (USA) using a static, three-axis 2G Enterprises superconducting rock magnetometer with an integrated alternating field (AF) demagnetised unit to obtain the characteristic remanent magnetizations (ChRM) of samples. The specimens were progressively AF-demagnetised (in 2 to 10mT steps) to a maximum field of 120mT. However, it was not possible to measure some samples due to their high natural remanent magnetisation (NRM) intensities ($>10^{-1}$ A/m). For these samples (mostly ankaramitic dykes), the specimens were cut in half to decrease the signal strength. Several samples were of high-coercivity, with generally 50% of the NRM remaining after treatment to 100mT field. These were then thermally demagnetised up to 575°C. The best fit lines were determined using principal component analysis (PCA) following the method of *Kirschvink (1980)* with the *PaleoMag* software (*Jones, 2002*). Sample maximum angular deviation (MAD) values for linear data ranged from $<1^\circ$ to 5° . It was never necessary to connect the magnetisation vector to the origin. When the angular distance between the direction of an individual sample within a site and the estimated site mean direction was greater than 15° , the sample was rejected from the site mean calculation. The individual specimens for all sites sampled within the galeria yielded highly scattered results between samples, likely due to orientation errors associated with the strong magnetic field within the galerias.

Of the 82 surface sites collected, 77 were fully demagnetised (5 sites were unusable because of post-sampling damage) with 70 sites yielding interpretable results. The seven rejected sites were excluded due to high dispersion between samples from the site.

2.2. Rock Magnetic Experiments

To fully characterise the magnetic mineralogy, standard rock magnetic experiments were conducted on representative specimens. Experiments include AF-demagnetisation of the

NRM, acquisition of isothermal remanent magnetisation (IRM), thermal demagnetisation experiments and continuous susceptibility vs. temperature experiments (see section 1.1). The AF-demagnetisation experiments were conducted by applying an increasing magnetic field up to 120mT. The IRM acquisition curves constitute a record of the increase of the induced magnetisation intensity when the sample is exposed to regularly increasing magnetic fields until saturation is reached (*Dunlop, 1973*). The thermal demagnetisation experiments involves a progressive heating of the specimen in a near zero magnetic field, followed by progressive cooling till room temperature. This process removes the NRM and isolates the characteristic remanent magnetisation (ChRM), allows demagnetisation of high coercivity phases (e.g., fine-grained maghemite, hematite, and very fine-grained magnetite), and also allows one to infer the blocking temperatures. The investigation of continuous susceptibility vs. temperature experiments is a powerful method to constrain the composition of the magnetic phase. The sample is first heated continuously from 25°C to 700°C and then cooled to 25°C while the susceptibility is measured continuously (*Hrouda, 1982*). These experiments were conducted in an argon atmosphere to reduce the oxidation effects.

3. Results

3.1. Rock magnetism

The behaviour during AF-demagnetisation allows the determination of the magnetic grain size (Single Domain, Pseudo Single Domain, Multi Domain; *Dunlop, 1981*) that carries the NRM (**Fig.7**). Most of the curves show a regular drop of intensity, with a slight variation (increase or plateau of the intensity value) at the first steps of the AF-demagnetisation (**Fig.7B, C, D, E, F, G, I**). Some samples were characterised by rapid decrease of intensity at the very first steps of demagnetisation followed by a plateau with a low intensity (**Fig.7H**).

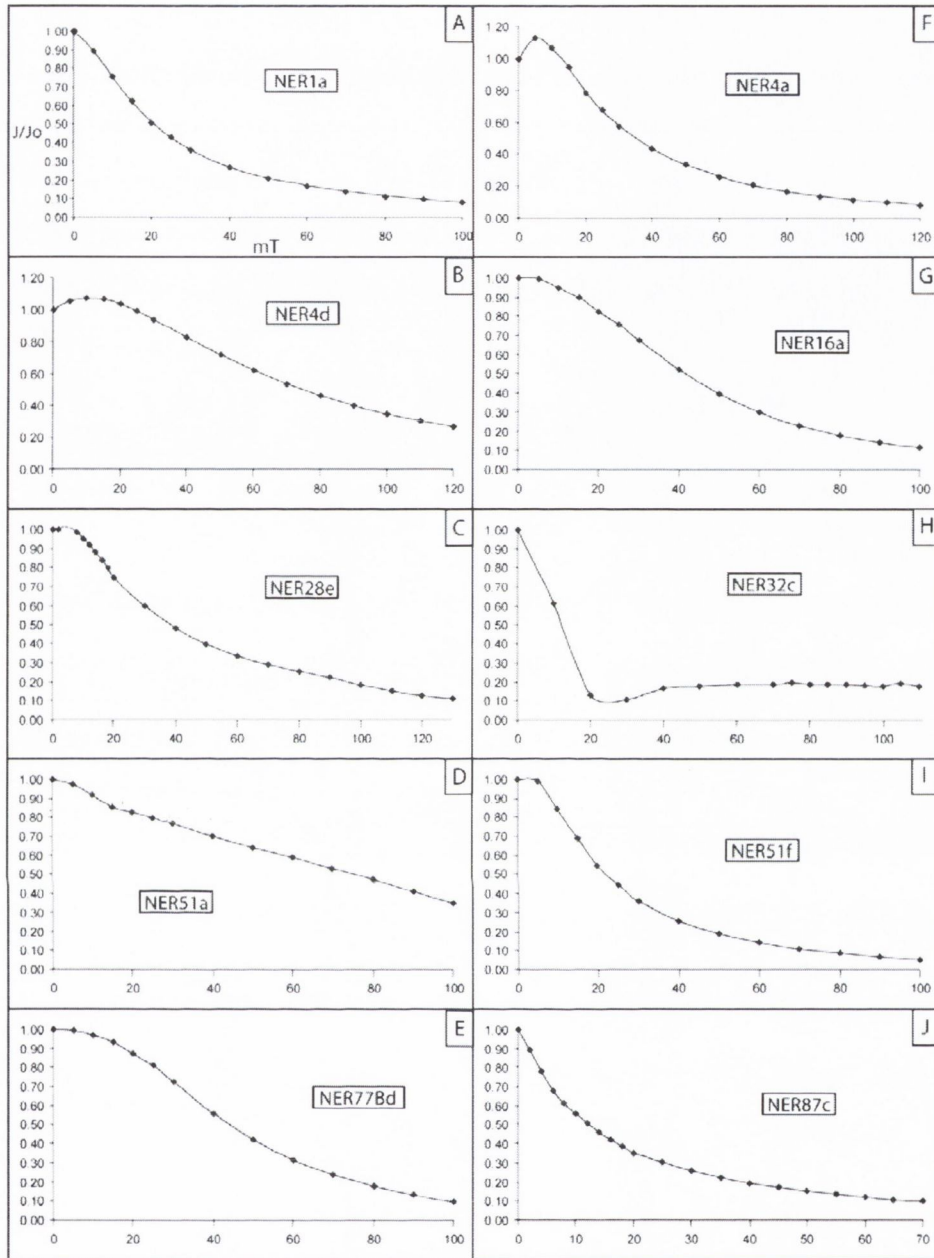


Figure 7: AF-demagnetisation of the natural remanent magnetisation (NRM) for several representative sites. The y-axis corresponds to the intensity of magnetisation normalised to the NRM and the x-axis corresponds to the strength of the applied magnetic field in mT. Note the difference of response between the specimens within the same site (NER4, specimens a & d; NER51, specimens a & f). Diagrams are designated by a site number (NER1, NER2 etc) and a specimen letter (a, b etc). Most of the curves display a concave-upward decrease of intensity.

The shapes of the IRM curves (Fig.8) provide information on the magnetic grain size and composition. In general, the IRM acquisition experiments showed steep acquisition of the magnetic intensity with increasing applied field followed by a saturation plateau (Fig.8), usually reached after 0.15/0.20T (an exception to this general pattern are from the ankaramitic part of NER18 that reaches the plateau after 0.05T, see Fig.8).

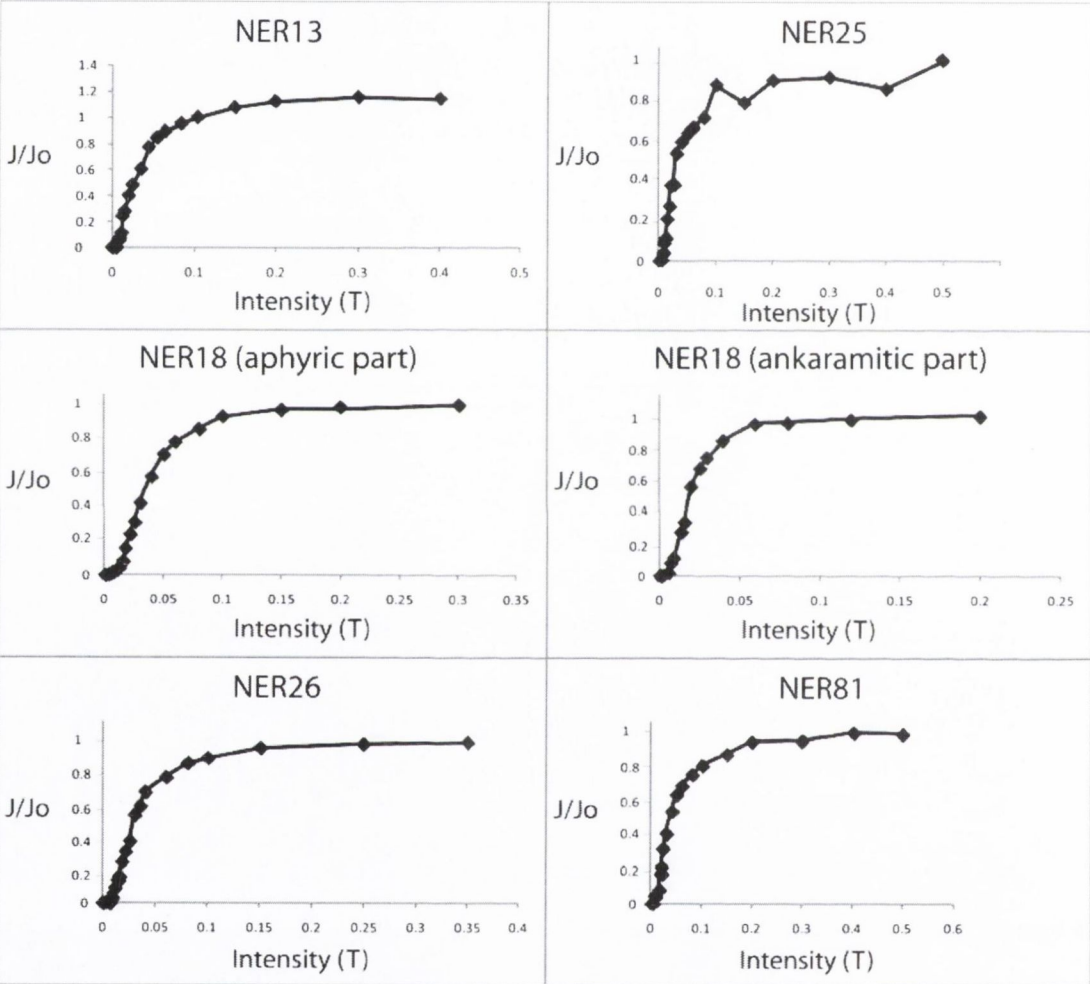


Figure 8: Representative isothermal remanent magnetisation acquisition (IRM) experiments. The ordinate represents the normalised intensity of magnetisation and the abscissa corresponds to the intensity of the applied magnetic field (in Tesla). Most samples are characterised by a steep magnetic intensity acquisition with increasing applied magnetic field followed by a saturation plateau.

Temperature variation versus bulk susceptibility was investigated to characterize the type and composition of the magnetic phase that carries the NRM. The experiments showed a complex behaviour with various Curie points during the heating phase followed by an increase of the susceptibility on the cooling curve on several samples (**Fig.9**). For example, NER77 (**Fig.9A**) displayed a Curie point at about 560 to 580°C where the susceptibility dropped dramatically until the end of the heating phase. During the cooling phase, we observed a rapid increase of the susceptibility from 600 to 480°C. In contrast, NER35 (**Fig.9B**) shows one Curie point at about 120°C with an inflexion point at 480°C, while NER14 (**Fig.9C**) displays a Curie point near 95°C. For both samples, an increase of the magnetic susceptibility was again observed during the cooling phase. NER76B (**Fig.9D**) shows a Curie point at about 100°C followed by a gradual decrease in susceptibility during the heating phase until reaching a second Curie point at about 480°C. Susceptibility then decreased regularly until 700°C. The cooling phase was characterised by a rapid increase of susceptibility from 700 to 500°C. Then the susceptibility decreased slightly with an inflexion point at about 100°C.

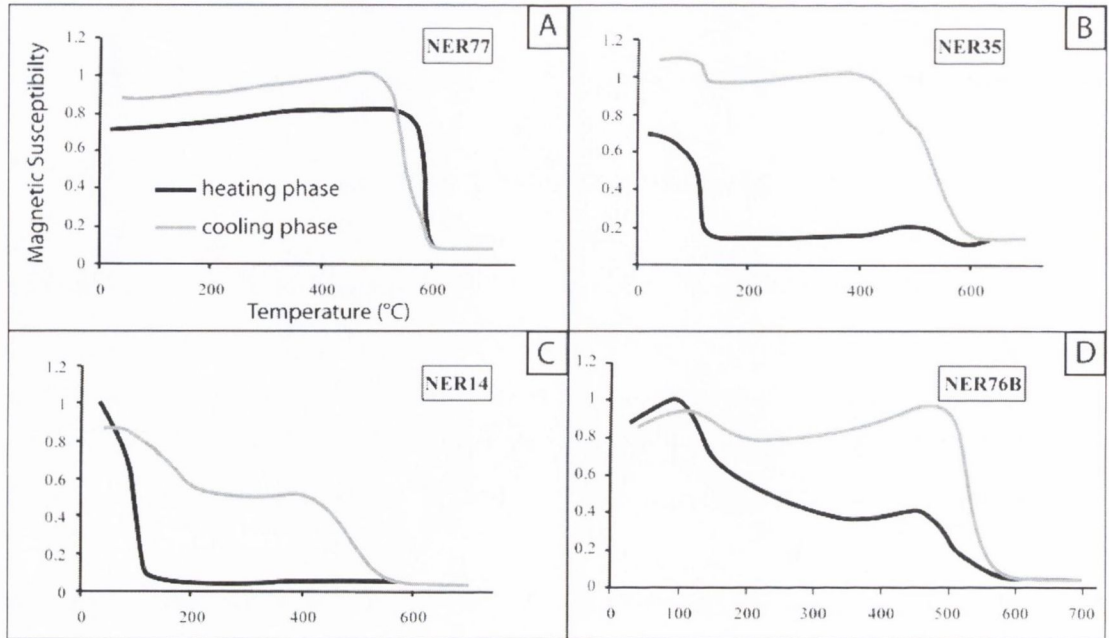
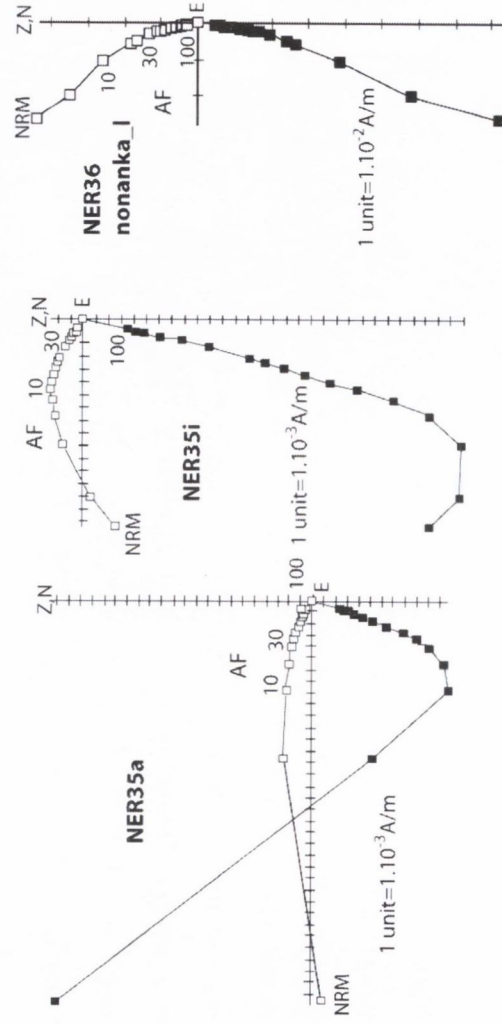
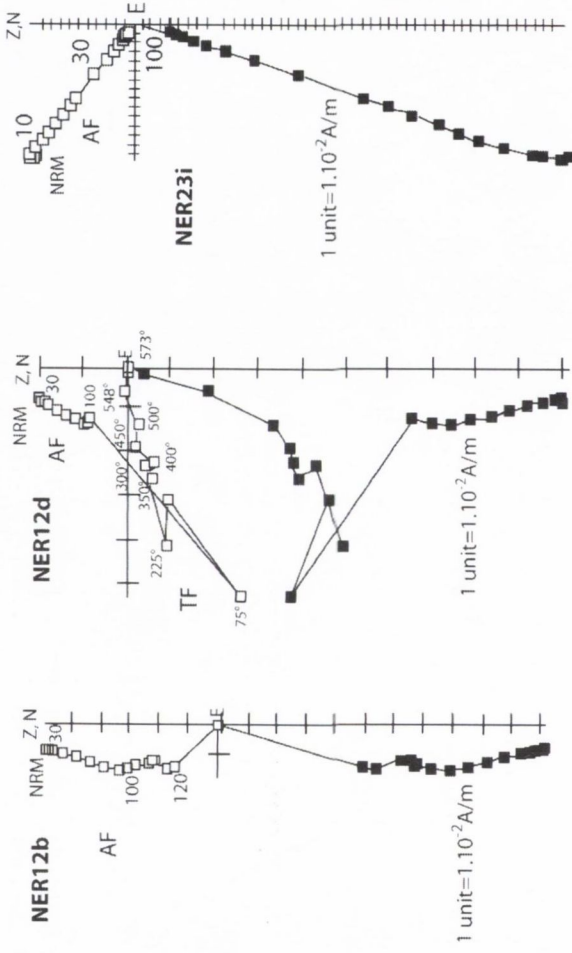


Figure 9: Continuous bulk susceptibility versus temperature experiments. Measurements were acquired during a heating phase (black line) up to 700°C followed by a cooling phase (grey line) down to ~25°C. All experiments were conducted in an Ar atmosphere. In general, the experiments reveal a complex behaviour during the heating phase with several Curie points and by an increase of the magnetic susceptibility during the cooling phase.

For most sites, the maximum applied AF-demagnetisation field was 120mT and most samples were fully demagnetised by this field with less than 10% of NRM remaining after this treatment. However, some samples required thermal demagnetisation to fully isolate the ChRM. Upon thermal treatment, the demagnetisation steps were highly scattered and did not yield a linear trend to the origin (**Fig.10** Vector Diagrams NER6_c & NER12_d). The thermal demagnetisation curves provide unblocking temperature spectra (**Fig.11**; see **section 1.1**; Dunlop and Özdemir, 1997). Blocking temperature and unblocking temperature are the same but the first one refers to the cooling phase, while the second to the heating phase. All samples analysed reveal low unblocking temperature spectra (between 100 and 200°C) and high unblocking temperatures around 500/600°C.



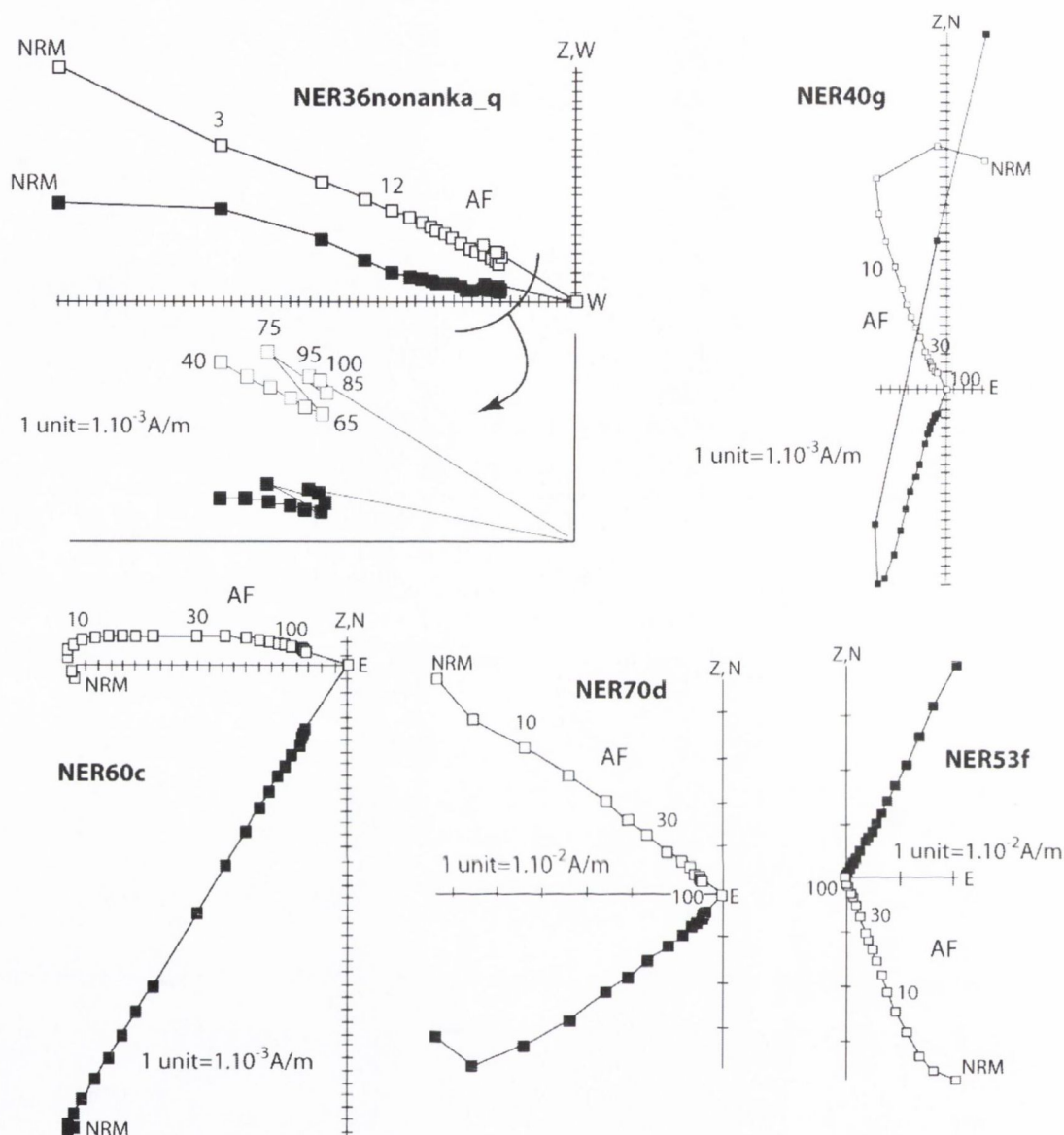


Figure 10: Representative *in situ* modified demagnetisation diagrams (cf. Zijdeveld, 1967; Roy and Park, 1974) for samples collected along the NERZ. Solid (open) symbols represent the projection onto the horizontal (true vertical) plane. Typically, AF demagnetisation steps are given in milliTesla (mT) and thermal demagnetisation steps in degrees centigrade °C. AF and thermal demagnetisation results from two specimens of the same sample are shown for comparison for some samples. Diagrams are designated by a Site number (e.g., NER2), method of treatment (AF or TH). Intensity (A/m) is shown along one axis for each sample, each unit mark equals indicated intensity. NRM = natural remanent magnetisation.

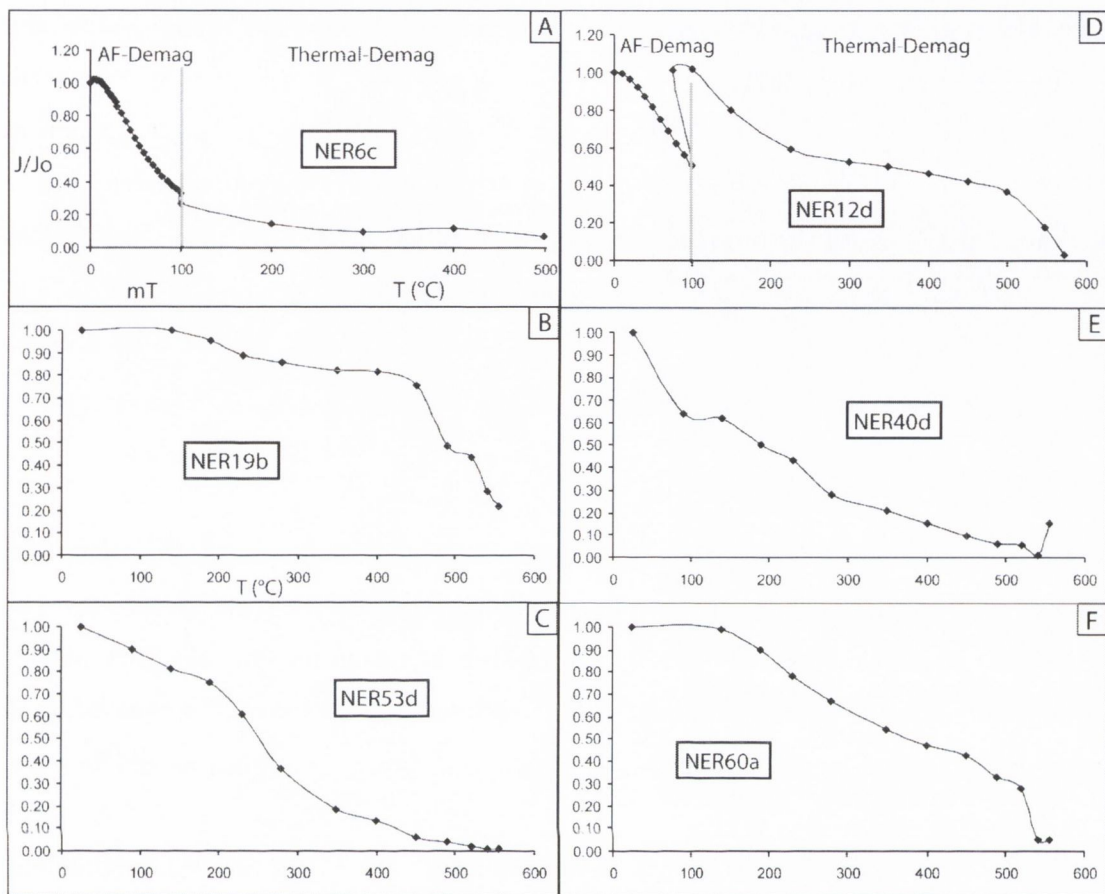


Figure 11: Thermal demagnetisation of the natural remanent magnetisation (NRM). The y-axis corresponds to the intensity of magnetisation normalised to the NRM and the x-axis corresponds to the strength of the applied magnetic field in mT for **11A** and **11D** or to the temperature (in °C). The two specimens **NER6_c** (**11A**) and **NER12_d** (**11D**) were first AF-demagnetised up to 100mT. Low unblocking temperatures (100/200°C) and high unblocking temperatures (500/600°C) are observed.

3.2. Paleomagnetism

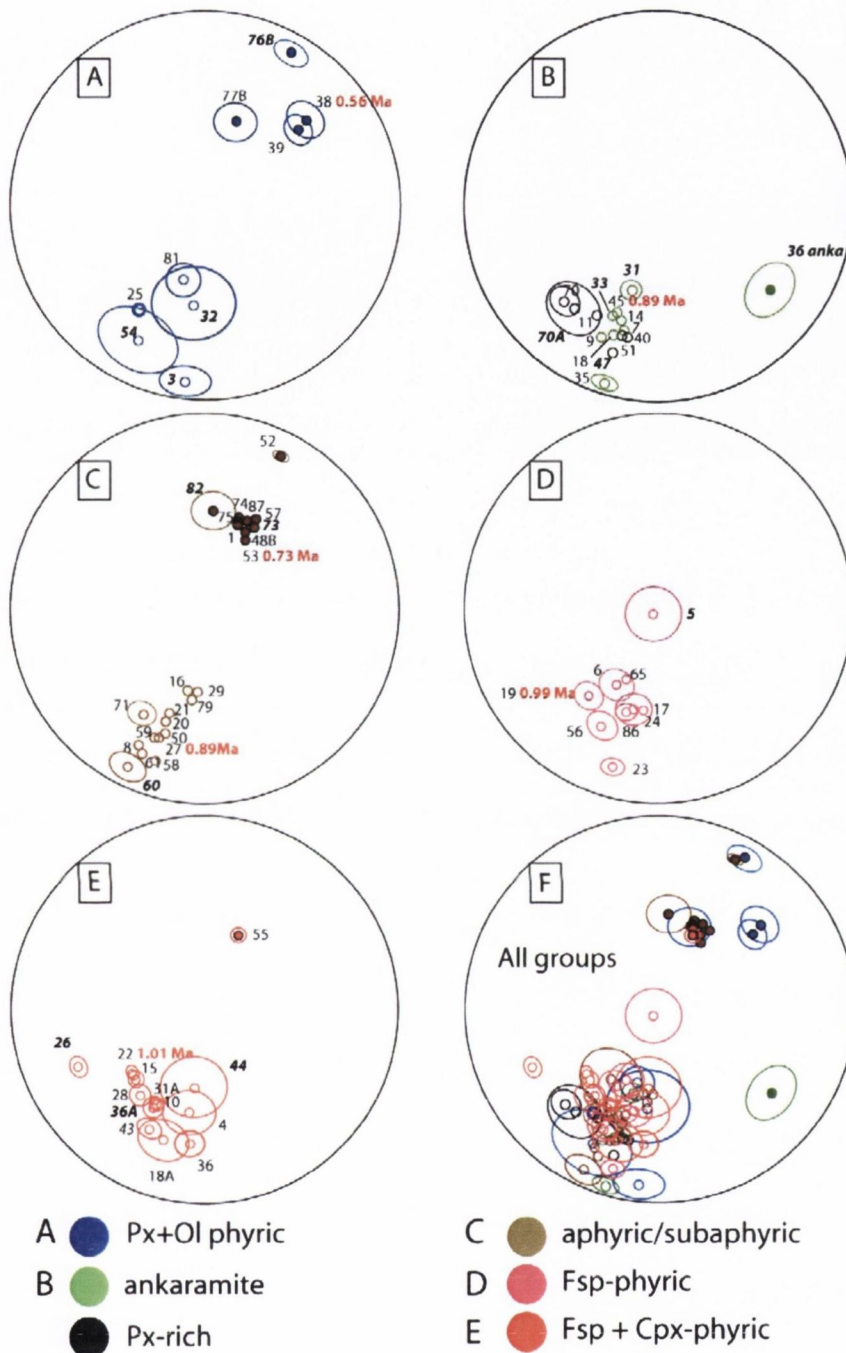
3.2.1. General Demagnetization Behaviour

The ChRM was successfully determined from 70 sites with most sites yielding a reverse polarity magnetisation (77%) and the remaining sites displaying a normal polarity magnetisation (23%).

As previously mentioned, thermal demagnetisation of replicate specimens resulted in highly scattered results. Fortunately, AF-demagnetisation provided high quality and readily interpretable results for most sites. Most specimens showed a single component magnetisation with a small superimposed signal. This secondary component corresponded to a low coercivity overprint on the NRM and was generally removed at around 10-15mT (**Fig.10**). Usually, after the removal of the overprint, the ChRM typically decays along a near univectorial path towards the origin with less than 10% of the NRM intensity left after AF-demagnetisation to 120mT. A few samples showed highly scattered data, likely due to lightning strikes (*Cox, 1961*).

3.2.2. Equal area projection and dyke groups

From the 77 analyzed dykes it was possible to determine 70 site mean directions. The data of the 7 remaining sites were too scattered to estimate the ChRM of these sites. The Fisher statistical procedure was used to average the data from the 70 sites. Each plot on **figure 12** corresponds to one equal area projection where the site means of each petrographic group have been plotted in order to check if petrographic variability is reflected within the site mean paleomagnetic directions. Note that the pyroxene-rich group and the ankaramitic group are displayed on the same diagram. Where possible, confidence ellipses were added (depending on readability). The last plot includes all the data from the 6 petrographic groups. The normal polarity sites (solid circles) plot in the NE quadrant with inclinations varying from shallow to intermediate values (i.e. around 50°). The reverse polarity sites (open circles) plot in the SW sector with inclination angles that also vary from shallow to intermediate values. The distinctive petrographic groups on **figure 12** are not clustered on the equal area projection, however, the groups are statistically indistinguishable meaning that the confidence intervals overlap for each petrographic group. Note that the large groups of aphyric and Px+Ol dykes are of both normal and reverse polarity (**Fig.12A&C**), whereas the other petrographic groups are of single polarity.



in bold-italic: sites that have been rejected for the grand mean calculation (table 1)

Figure 12: Equal area projection of site mean directions (lower hemisphere). Open symbols represent a site that belongs to a reverse polarity. Solid symbols represent a site that belongs to a normal polarity. Confidence ellipses are also represented. However, for an esthetical and clarity reasons, some ellipses are not displayed in the sketch (see $\alpha 95$

in table 1 if needed). Each petrographic group is represented by one colour and all the sites (irrespective of petrography) are plotted in the last diagram (F). The sites rejected for the group mean calculations are in italics. The normal polarity sites (solid circles) are in the NE quadrant with shallow to intermediate inclination values (i.e. around 50°). The reverse polarity sites (open circles) are in the SW quadrant with shallow to intermediate inclination values. The distinctive petrographic groups are statistically indistinguishable (F), i.e. the confidence intervals of each groups overlap. Ages in Ma obtained using the Ar-Ar technique are added in red for each dated dyke.

3.2.3. Grand mean results

The expected paleomagnetic pole for the Canary Islands was averaged from published literature values (**table 2**) (Watkins, 1973; Storevedt et al., 1979; Najiid et al., 1981; Carracedo and Soler, 1995; Quidelleur and Valet, 1996). The expected declination is $D=358^\circ$ and the expected inclination is $I=40.7^\circ$ with an $A95=5.6$ (see **Fig.13B**). The group mean of the reverse polarity of our data set is $D=202.7^\circ$, $I=-40.1^\circ$, $\alpha95=3.7^\circ$ and the group mean of the normal polarity data yields $D=27.7^\circ$, $I=49.3^\circ$, $\alpha95=6^\circ$ (**Fig.13A**). After inverting the reversed polarity sites through the origin, we estimate a grand mean of $D=23.8$, $I=42.3^\circ$, $\alpha95=3.2^\circ$, $K=39$, $N=51$, which is discordant to the expected paleomagnetic direction (**Fig.13B & C**). Note that only 51 sites were used from the 70 interpretable sites for the grand mean calculation, as 19 sites were excluded due to scattered data or because they represent transitional fields

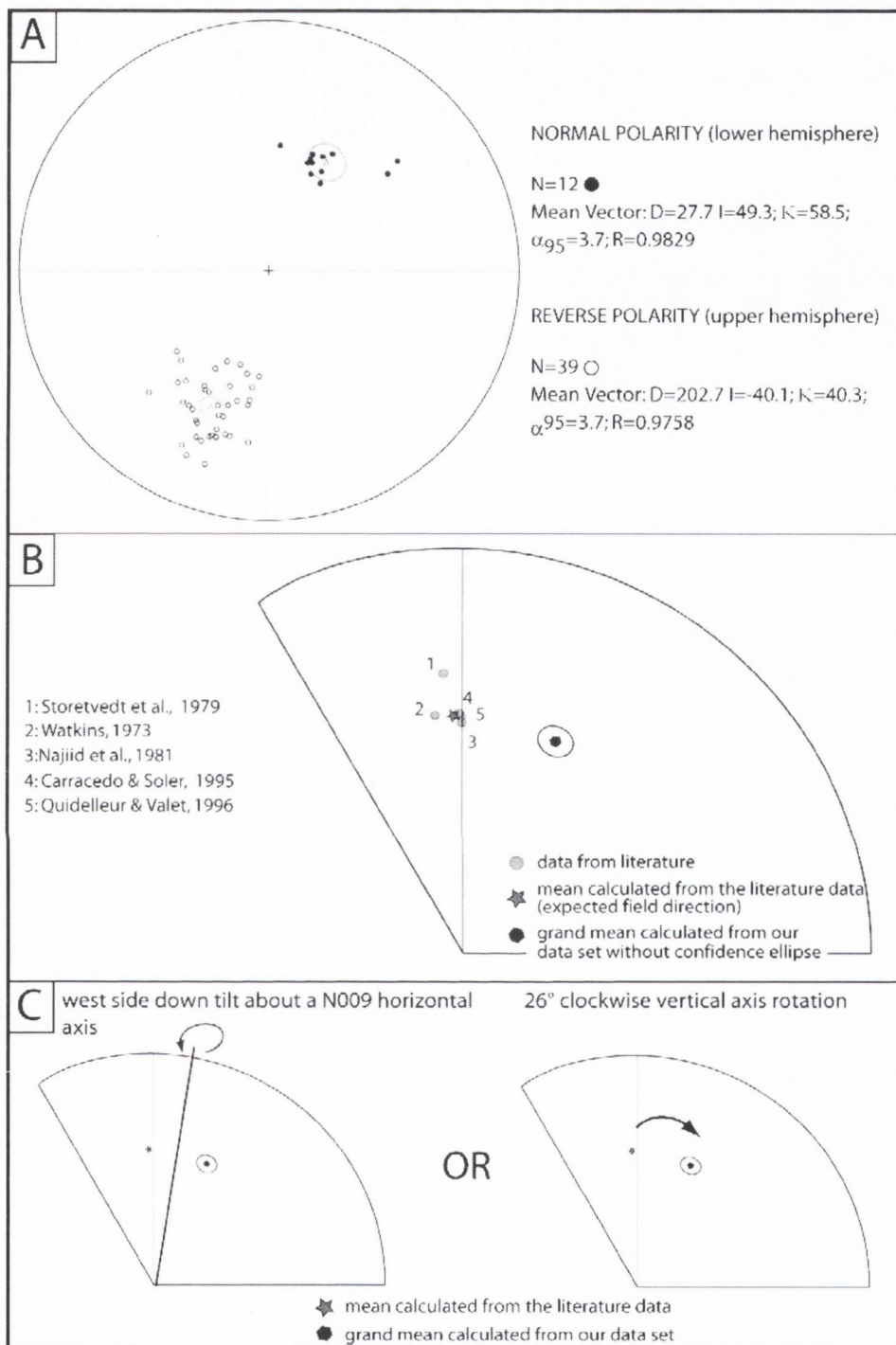


Figure 13: A: Equal area projection of the site mean data that were used to calculate the overall group mean of the NERZ dykes. Open (solid) symbols represent a site that belongs to a reverse (normal) polarity. The means calculated for the normal and reverse polarity data sets are shown on the right. **B:** Equal area projection of the grand mean

result calculated from our data set (after inverting the reverse polarity sites through the origin) compared to the expected direction for a Miocene to Pleistocene field (see table 2). The grand mean is discordant to the expected late Miocene to Pleistocene magnetic field. **C:** Schematic diagram of the two possibilities that can explain the offset between our grand mean direction and the expected field direction. West side down tilt about a N009 horizontal axis on the left; 26° clockwise vertical axis rotation on the right.

3.2.4. Field and laboratory tests

3.2.4.1. Baked country rock test

The baked country rock test is a complementary study to ensure paleomagnetic stability and antiquity of the magnetisation (Butler, 1992). The principle of this test is that the baked country rock should have the same paleomagnetic direction than the igneous rock. If this test is successful, the ChRM may be a primary NRM. In the current study, five baked country rock tests were conducted. The country rocks are lava flows cut by the dykes. The site NER19A (adjacent to the dyke NER19) yields highly scattered results and the ChRM of the unbaked host rock cannot be determined. It is not possible to calculate the current fits for NER47A, the demagnetisation step being highly scattered (adjacent to the dyke NER47). For the sites NER44A and 50A (adjacent to the dykes NER44 & 50 respectively) the ChRM of the baked country rock, the country rock and the dykes are different. The samples within the host rock do not show a coherent and progressive remagnetisation towards the dykes (see **Fig.14**). We also observed a change in polarity for the country rock 44A (reverse polarity), with cores k and l exhibiting a normal polarity. These two specimens (k, l) are, moreover, the closest to the dyke NER44 which belong to a reverse period. Given the high scatter between samples within the country rock, we conclude that the nature of the host rock is, unfortunately, not amenable to a contact test.

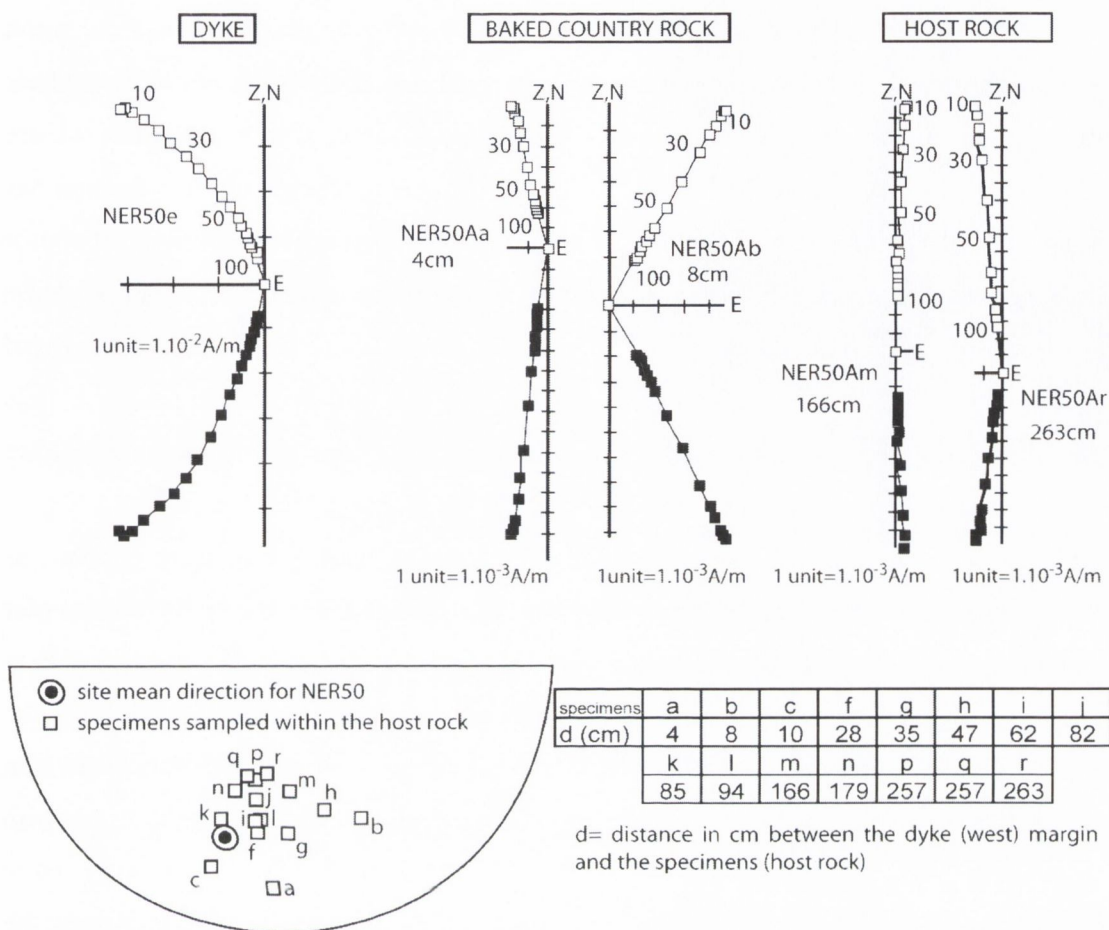


Figure 14: Baked country rock test for the site NER50. Demagnetisation diagrams for the dyke, the baked country rock and the host rock. Solid (open) symbols represent the declination (the inclination). The distance (in cm) between the samples in the host rock and the dyke margin are shown along on the diagrams and in the table (bottom right). The mean direction for the dyke NER50 is plotted on the equal area projection (bottom left) together with the individual directions of the host rock specimens. We note that the host-rock samples do not show a progressive remagnetisation towards the dyke.

3.2.4.2. Reversal test

The reversal test is based on the fact that, if secular variations have been adequately averaged, the mean direction of the reversed polarity data set should be anti parallel to that of the normal polarity data set. However, the presence of an un-removed secondary NRM may cause a deviation from anti parallelism between the two data sets.

Consequently, if the mean calculated from normal data set on one hand, and the mean calculated from reversed data set on the other hand plot at 180° from each other, and are statistically indistinguishable, the data set passes the reversal test. In that case, we are sure that any secondary NRM has been successfully removed and secular variation has been adequately time-averaged. On the contrary, if the dataset fails the reversal test, it may indicate that secondary NRM was not totally removed and/or the secular variation has not been effectively averaged. The reversal test is a statistical test. Merrill and McElhinny (1983) established a statistical method to check if the dataset passes or fails the reversal test. This method gives a number, referred as gamma, whose value indicate whether or not the dataset passes the test.

The time interval recorded by the sampled dyke spans at least two polarity intervals as both normal and reverse periods are observed. Four sites (NER5, 26, 31 & 36anka) plot more than 45° from the expected field direction and likely represent transitional field directions (Merrill and McElhinny, 1983). These sites were thus excluded from the calculation of the group mean. Sites without a mean direction or with high data dispersion (**table 1**) were also excluded from the group mean calculation. In addition, the sites with shallow inclination values (typically about $10/20^\circ$) and the sites with very large confidence ellipses were not taken into account. The 51 accepted sites did not pass the reversal test at the 95% level with an observed gamma = 9.86 and a critical gamma of 7.21 (Merrill and McElhinny, 1983; **Fig.13B**).

Another statistical method developed by McFadden and McElhinny (1990) consists in calculating the probability that the dataset passes the test. Under the null hypothesis (ie the datasets have mean directions that lay 180° apart and pass the reversal test), the probability of obtaining a test value greater than observed is 22%. In other word, the probability to pass the test is 22%. Thus the null hypothesis is accepted. The estimated value for common kappa is 42.6458 (Kappa corresponds to a measure of the dataset dispersion), and yields a critical angle of 7.3° . The angle between the mean of the normal polarity set and the mean of the reversal polarity set is 9.9° , therefore the data do not statistically overlap at 95% confidence. Under the null hypothesis that the two samples are drawn from populations with means that are 180° apart, the probability of exceeding

this angle is 0.468E-02. Consequently, the data did not pass the reversal test at 95% confidence.

3.2.4.3. Dispersion estimates.

A Virtual Geomagnetic Pole (VGP) is a paleomagnetic pole-position calculated from a single site-mean direction. The VGP represents a single measure of the position of the magnetic field relative to the site location. When VGPs are calculated from several sites, those VGPs should be scattered around the geomagnetic paleopole. The angular dispersion of these VGPs around the geomagnetic paleopole depends on the latitude (*Merrill and McElhinny, 1983*). If the angular dispersion is within the predicted value, it is probable that secular variation has been adequately average for the population of site mean directions.

An estimate of the dispersion of the VGPs was calculated for the 51 accepted sites (**table 1**). The calculated dispersion estimate yielded a value of 12.84, which coincides with the predicted VGP dispersion of 13.5 for a paleolatitude of $\sim 28^\circ$ (*Merrill and McElhinny, 1983*).

4. Discussion

4.1. Rock magnetism:

Most of the AF-demagnetisation experiments (**Fig.7**) indicate that the magnetic grains are dominantly pseudo-single-domain (PSD) with very few single domain (SD) grains. Rare samples reveal the presence of multi-domain (MD) grains, which are mostly confined to the coarse-grain ankaramite dykes.

The shapes of the IRM acquisition curves (**Fig.8**) are typical of SD to PSD grains, consistent with the AF-demagnetisation results. Consequently, the NRM is carried by a fined-grained, high-coercivity, cubic, magnetic mineral phase, likely low to moderate Ti-magnetite. Furthermore, rapid saturation is typical of the behaviour of a low-Ti Fe-oxide phase, likely magnetite.

Magnetite (Fe_3O_4) and ilvospinel (Fe_2TiO_4) constitute the end-members of the titanomagnetite solid solution (*Dunlop and Özdemir, 1997*). The temperature variation versus bulk susceptibility experiments allows determination of the magnetic phase and an

estimate of the relative Ti content of the magnetite. The increase in susceptibility during the heating phase reflects a mineralogical change (**Fig.9**) likely corresponding to the growth of a new magnetic phase during the heating step (*McElhinny and McFadden, 1999*). As the experiments were all conducted in an inert Ar atmosphere, it is thus likely that enough oxygen was trapped between the grains to allow the growth of a new magnetic phase, probably low-Ti magnetite, during the heating step (*Pariso and Johnson, 1991; Dunlop and Özdemir, 1997*). This is consistent with the significant increase of the magnetic susceptibility on the cooling curve (**Fig.9**).

Alternatively, a high Ti-Fe oxide phase may be the primary magnetic phase present in these rocks (*Appel and Soffel, 1984*). During the heating experiment, the high Ti-Fe oxide unmixed to produce separate *ülvospinel* and magnetite phases within the samples. The exsolution of the magnetite during the heating experiment is reflected by the dramatic increase in the magnetic susceptibility during some of the cooling experiments. The fact that the demagnetisation steps are highly scattered after thermal demagnetisation (**Fig.10**) reveals that oxidation of the magnetic phase occurred during the heating phase. It is indeed similar to the results obtained for susceptibility vs. temperature variation experiments. Geochemical data also show a high TiO_2 content that is consistent with these results (*Rodríguez Badiola et al., 2006*).

The thermal demagnetisation curves (**Fig.11**) show low unblocking temperatures (T_{UB}) and high unblocking temperatures. The low T_{UB} may correspond to maghemite, while high T_{UB} may correspond to magnetite or titano-magnetite, which is consistent with the IRM curves (**Fig.8**) and the graphs of temperature variation versus bulk susceptibility experiments (**Fig.9**).

4.2. Paleomagnetism

4.2.1. General demagnetisation behaviour

The rocks are typified by high quality and readily interpretable results during progressive alternating field demagnetisation (**Fig.7; Fig.10**). In general, most samples contain a single component magnetisation that is well-grouped at the site level, but most of the samples also contain additional magnetisation that is easily removed by 20mT. We interpret these low-coercivity magnetisation components as viscous remanent

magnetisation overprints (VRM). After removing the VRM, the ChRM, which we interpret as the primary thermal remanent magnetisation (TRM), decays along a roughly univectorial path to the origin with less than 10% of the NRM intensity remaining after treatment in 120mT fields.

The high scattered results using thermal demagnetisation on duplicate specimens are likely due the oxidation and exsolution of titanomagnetite grains during the heating experiment (cf. **section 4.1**).

4.2.2. Equal area projection: implication for the rift zone activity.

The paleomagnetic data plotted with the petrographic data (**Fig.12** and **section 3.2.2**) reveal the absence of distinctive groups that would record different intrusive batch events. The results instead overlap and form an almost continuous trend. This means that there has probably not been any prolonged cessation of intrusive activity, and the rift can be considered to be semi-continuously active during the Pleistocene. It is possible that magma has been stored in the rift for longer periods. Hence, we have sampled at a resolution that reflect several magma pulses during the late evolution of the rift. We exclude the hypothesis that the data record a transitional or high amplitude secular variation, as the dispersion of the data average the secular variation. One possible explanation for this dispersion is that the rift zone has had continuous activity over a long period of time as the dykes record at least two polarity intervals corresponding respectively to Matuyama (2.48 My – 780ky) and Bruhnes (began 780ky ago). The Ar-Ar dates and the paleomagnetic data suggest that the rift was relatively frequently active with multiple and ongoing minor intrusive events during the Pleistocene (**Fig.12**). This has also been confirmed by the field data collected during several campaigns (**chapter 1**). Indeed, the upper part of the rift is composed of lava flow units but also colluvium that provide evidence for the existence of paleo-valleys. Furthermore, these sedimentary units are cut by dykes, suggesting that the upper part of the rift is affected by limited erosion, concurrent with semi-continuous emplacement of minor intrusions. The deep core of the rift is Miocene and Middle Pliocene with a last stage being Middle Pleistocene in age (*Carracedo et al., 2009*). Considering the Ar-Ar ages obtained (**Fig.12**), this last period most probably corresponds to the stage that we sampled in our paleomagnetic study.

Consequently, the rift activity does not constitute a single event but a more dynamic continually active, long-lived system.

4.2.3. Baked country rock test

Three possibilities can explain the dispersion of the baked country rock test results. The first hypothesis consists of a large remagnetisation episode. However such event is unlikely to have occurred as the tests were performed at different locations across the rift. Furthermore, to our knowledge, no published study exists that documents a widespread remagnetisation event in the NERZ. The few samples with high NRM intensity (apart from the ankaramitic dykes) are probably due to lightning strike (*Cox, 1961*), but the different country rock tests were not conducted near these sites. Another line of evidence is that there are multiple remanence directions between the dykes across the ridge. If a regional remagnetisation event occurred, it is probable that all the dykes would have preserved this secondary magnetisation. The second explanation is that the dykes are more or less contemporaneous to some of the lava flows. This second possibility is also not likely as the ChRM recorded by the different dykes is different from the ChRM observed in the lava flows. We can thus infer that the sampled host rocks have been locally affected by the dyke emplacement but at a larger scale than the sampled zone (**Fig.14**). A reheating of the host rock may have caused changes in the magnetic properties (by oxidation and/or growth of a new magnetic phase for example) responsible of the ChRM lost. This process likely explains the random nature of the country rock orthographic projections on **figure 14**. The site means of the dykes are well-constrained, while the samples drilled and analysed in the host rock led to scattered data. Fluid circulation within the country rock may have also changed the magnetic properties.

4.2.4. Reversal and VGP dispersion tests.

The value of the VGP dispersion (12.84) is close to the predicted value (13.5) for the paleolatitude $\sim 28^\circ$ (*Merrill and McElhinny, 1983*). Consequently, the results from the rift zone intrusions collectively average secular variation and represent a time-field direction. The data are thus expected to provide reliable results concerning the vertical axis rotation

or the horizontal axis tilt of the rift. The reversal test failed at 95% confidence, with an observed $\gamma = 9.86$ and a critical γ of 7.21 (*Merrill and McElhinny, 1983*). Failure of the test can be due to either an unremoved secondary NRM component or to an insufficient sampling of the geomagnetic secular variation (*Butler, 1992*). However, the estimation of the VGP dispersion shows that our data set likely averages the geomagnetic secular variations during both reverse and normal-polarity intervals. Furthermore the secondary NRM (VRM) has been totally removed in all cases. The failure of the test is likely to be due to the difference of the data set size between the normal ($N=12$) and reverse ($N=39$) polarity, even if the κ values are similar (same order of magnitude). We note that if we accept all results of normal polarity sites with shallow inclination, the test passed at 95%.

4.2.5 Overall Rotation or Tilting

The grand mean is discordant to the expected direction (**table 2**) with an inferred rotation of $26.2^\circ \pm 7.3^\circ$ and a flattening of $-3.5^\circ \pm 5.6^\circ$. The offset can be adequately explained by either a 26° clockwise vertical axis rotation or a 28° WNW-side-down tilt about an average N009 trending tilt axis (**Fig.13C**). The rotation or tilt could, however, affect (a) only the sampled part of the rift zone or (b) the entire island. Several possibilities can explain the discordant data. Geophysical studies have revealed a lithosphere flexure under Tenerife Island due to the edifice load (*Watts, 1994; Canales et al. 2000*). Sagging of the island can induce an inward tilt of the units as observed in various volcanoes and confirmed by analogue models (*Byrne et al., 2009*). Such flexure would explain the deduced tilt, but would also affect the entire island (*Collier and Watts, 2001*). The rotation could also be related to the plate motion associated with the Atlantic rift transform faults or the hypothetical transform fault running from the High Atlas mountains to the Canaries (*Beck and Lehner, 1974; Anguita and Hernán, 1975; Robertson and Stillman, 1979*). In this case, the transform faults (whatever the origin) associated with strike-slip movements would be responsible for the island's rotation. In order to test these hypotheses, paleomagnetic data collected away from the rift would be required. We sampled several lava flows and dykes in the Anaga massif (NE extremity of the island), in a supposedly stable area. Unfortunately, the sampled sites did not adequately average

secular variation (**Table 3**). However, *Leonhardt and Soffel (2006)* recently studied late Miocene lava flows (Old Basaltic Series, *Fúster et al., 1968*) in the Teno massif for paleomagnetic work (**Table 4**). The mean of the reverse polarity of their data set is $D=184.9^\circ$, $I=-47.3^\circ$, $\alpha_{95}=22.9^\circ$ while the mean of their normal polarity data set yields $D=7.8^\circ$, $I=38.1^\circ$, $\alpha_{95}=10.2^\circ$ (excluding sites BC2 and P1 as they likely record transitional fields). These data passed the reversal at 95% confidence (cf. *McFadden and McElhinny, 1990*). After inverting the reversed polarity sites through the origin, we calculated a grand mean of $D=3.2^\circ$, $I=38^\circ$, $\alpha_{95}=5.6^\circ$, $K=29.4$, $N=24$. The dispersion of the data is 14.9° which fall within the confidence limit of the predicted VGP dispersion for the latitude of Canaries (*Merrill and McElhinny, 1983*). Their results are statistically indistinguishable to the average expected direction determined by *Watkins (1973)*, *Storevedt et al. (1979)*, *Najjid et al. (1981)*, *Carracedo and Soler (1995)*, *Quidelleur and Valet (1996)*, (**Tables 2, 3** and **Fig. 15**). We conclude therefore that the tilt/rotation affects only the sampled part of the rift and not the island as a whole.

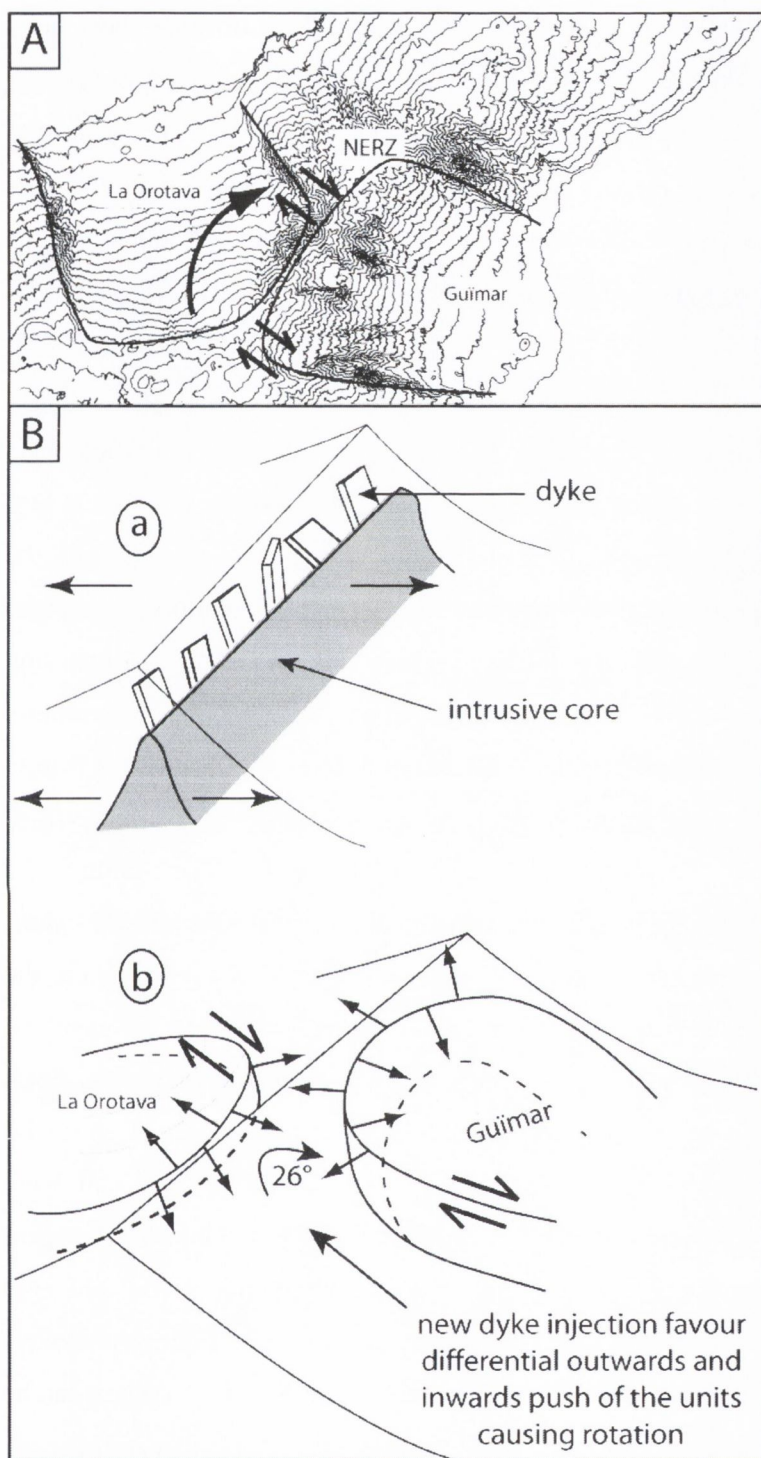


Figure 15: Interpretative sketch of the NERZ upper unit rotation inferred with paleomagnetic data. **A:** map view; **B:** 3D-sketch. **A:** The arrows show the deduced strike-slip movements related to the clockwise rotation. As the sampled section is in-between the two valleys of La Orotava and Guimar, we interpreted the dextral strike-slip regime as the results of flank collapse accommodation. **B:** Model to explain the rotation of the upper unit of the NERZ; **(a)** intrusive core emplacement and its associated dykes push the rift flanks outwards that will trigger rotation of the units and flank collapses; **(b)** the rotation and collapses are accommodated by strike-slip movements. Continued dyke emplacement will push the surroundings and favour further destabilisation.

The lava flows observed along the sampled area are mostly subhorizontal. Down the rift exposure, along the La Orotava scar (see Fig.6), the lava flows are also mainly

subhorizontal. If a WNW-side-down tilt of 28° had occurred, the different lava units would have been tilted towards the west. It is thus more likely that the paleomagnetic data reflect a vertical axis rotation of the units rather than a tilt (**Fig.15**). It is not unusual to identify vertical axis rotation using paleomagnetic data. Indeed, such phenomenon has been documented in paleomagnetic studies in various tectonic contexts (mountain chains or extensional basins) and is commonly associated with strike-slip movements and extension (Gillett and Van Alstine, 1982; Ron et al., 1984; Hudson and Geissman, 1987; Janecke et al., 1991; Hudson, 1992; Hudson et al., 1994; Wawrzyniec et al., 2002; Petronis et al., 2002; Petronis et al., 2007). In a volcanic context, a vertical axis clockwise rotation has been recognized in the rift and El Golfo sequences of El Hierro island (Szérméta et al., 1999). Strike-slip movements have been recognised along the Güímar and La Orotava collapse scars through dyke and fracture orientation pattern (see **chapter one**). As the dykes record the prevailing magnetic field at the time of intrusion (likely Pleistocene regarding the Ar-Ar ages; see **Fig.12**), it is evident that the rotation occurred after dyke emplacement and cooling, and thus it is most likely a recent feature of the rift. Consequently, we interpret the vertical axis clockwise rotation as a result of strike-slip movements that accommodate flank destabilization (**Fig.15**). The sliding of the flanks creates strike-slip faults along the scars that will rotate the host rock and the dykes. La Orotava and Güímar valleys are on each side of the sampled section. Hence, the dextral strike-slip faults that accommodated sliding along the NE scar of La Orotava and along the SW scar of Güímar are the most likely candidates for this rotation. Such mechanisms have been observed in many volcanoes cut by landslides, such as on La Réunion Island where tension gashes are described along the crater rim and were interpreted as the results of strike-slip movements that accommodate flank displacement (Carter et al., 2006). Furthermore, strike-slip movements and rotation associated with flank sectorial spreading have been previously observed using analogue models (Delcamp et al., 2008b; Delcamp et al., in review). In such scenarios, flank sliding can be due to the emplacement of a deep intrusive core and its associated dykes (Swanson et al., 1976; Dieterich, 1988; Cayol et al., 2000; Brooks et al., 2008; Delcamp et al., in review). Then, intense intrusive activity can generate flank destabilisation that may lead to sliding and this displacement is accommodated by block rotation in coherent portions of the

landslide. Note that the Ar-Ar ages are in agreement as the dated dykes predate or are contemporaneous to the NERZ collapse events.

5. Conclusion

The range of the site mean directions from individual dykes of the rift provides evidence that the NERZ is a long-lived and dynamic system. This is also confirmed with recent field data. The sampled dykes cover two polarity intervals, the Matuyama (reverse) and Bruhnes (normal) periods. The intrusive activity is characterised by several magmatic events which are relatively closely spaced in time and seem to be linked to flank deformation and collapses, where the destabilisation of the ridge flanks creates fractures and space for magma to intrude. The discordant paleomagnetic data are interpreted as a vertical-axis rotation of the sampled part of the rift. We interpret the 26° clockwise vertical-axis rotation as the result of strike-slip movements related to flank sliding that occurred on the edge of the collapse scar to accommodate the intrusive core at depth and dyke emplacement in the near surface environment. The volcanic rift-zones are therefore not a “permanent” structural feature, and the activity is characterized by concentrated phases of activity separated by longer periods of quiescence. They thus represent a dynamic and changing geological and geo-morphological environment.

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TABLES

Table 1: Paleomagnetic data from the sampled area. The sites are labelled NER1, NER2 (sometimes also with a capital letter as in NER77B) etc. and each specimen from a site is labelled by a lowercase letter (a, b etc). For each site, No is the number of samples analyzed per site, N is the number of data retained to calculate the site mean. R corresponds to resultant vector length, α_{95} to the 95% confidence interval about the estimated mean direction (assuming a circular distribution) and κ is the best estimate of the precision parameter (Fisher). D and I correspond respectively to the in situ declination and inclination. Lat/Long are the latitude and longitude of the VGP's and the site. *: site rejected due to high dispersion between the samples. In bold: rejected site for the grand mean calculation (high dispersion or records a transitional field).

	site							VGP		Site location		
SITE	No	N	R	α95	κ	D	I	LAT	LONG	LAT	LONG	
NER1	9	7	6.961	4.54	152.4	21.5	51.7	71.06	56.03	28.346	343.5185	
NER3	8	5	4.945	8.08	72.6	187	-10.8	66.16	146.05	28.346	343.519	
NER4	7	7	6.817	9.89	32.8	188.8	-45.4	82.07	82.06	28.346	343.5188	
NER5	9	5	4.888	11.56	35.8	224.3	-86.7	32.95	348.90	28.347	343.5192	
NER6	9	8	7.878	6.88	57.5	209.1	-53.2	67.78	52.58	28.347	343.5193	
NER7	8	6	5.757	13.76	20.6	195.5	-35			28.348	343.519	
NER8	7	6	5.906	8.44	53.3	206.5	-22.8	60.28	101.79	28.348	343.5192	
NER9	10	10	9.922	4.28	115.4	203.3	-27.9	64.54	100.70	28.348	343.5192	
NER10	6	6	5.991	2.55	575.4	206.3	-44.7	66.60	72.25	28.348	343.5193	
NER11	8	8	7.934	5.07	105.4	209.1	-36.6	62.40	83.78	28.348	343.5193	
NER12	10	*									28.348	343.5194
NER13	9	*								28.348	343.5202	
NER14	9	8	7.882	6.78	59.3	197.5	-38.7	72.89	91.97			
NER15	8	8	7.967	3.55	213.6	224.5	-48.4	51.26	60.90	28.349	343.5204	
NER16	9	8	7.890	6.54	63.6	192.2	-55.2	77.31	34.87	28.349	343.5204	
NER17	8	7	6.908	6.97	65	193.8	-45.6	77.72	76.35	28.349	343.5205	
NER18	9	8	7.981	2.69	373	199	-30.7	68.88	103.48	28.349	343.5204	
NER18A	8	8	7.770	9.52	30.5	197.7	-30.4	69.75	106.09	28.349	343.5204	
NER19	8	8	7.905	6.08	73.6	219	-41.9	54.92	71.17	28.346	343.5212	
NER20	7	7	6.985	2.77	408.7	199.2	-38.6	71.43	89.97	28.346	343.5212	
NER21	9	9	8.979	2.49	379.5	198.8	-49.6	73.51	61.68	28.346	343.5212	
NER22	11	10	9.965	2.86	258.16	229	-49.2	47.53	58.80			
NER23	8	7	6.969	4.05	191.4	196.3	-16.5	64.85	122.74			
NER24	9	9	8.939	4.26	130.8	188.4	-46.4	82.56	76.50			
NER25	9	9	8.982	2.31	443.9	213.2	-37	58.93	80.36			
NER26	9	7	6.970	3.98	197.7	245.8	-29.4	28.43	70.18			
NER27	9	9	8.899	5.47	79.5	199.5	-30.1	60.38	22.99			
NER28	8	8	7.945	4.62	126.4	216.9	-44.2	57.20	68.92			
NER29	8	7	6.976	3.53	251	185.5	-55.1	81.35	14.73			

NER31A	9	8	7.983	2.58	403.8	209.4	-45.7	64.01	69.19	28.352	343.5244
NER31	9	7	6.969	4.02	193.6	197.5	-53	66.38	337.18	28.352	343.5244
NER32	6	5	4.785	16.19	18.6	187.5	-47.9	83.39	66.42	28.352	343.5249
NER33	11	7	6.810	10.09	31.5	202	-39.3	69.24	85.58	28.352	343.5247
NER35	9	6	5.974	4.39	194.8	196.6	-6.5	60.35	128.31	28.353	343.525
NER36AN	9	6	5.856	10.49	34.8	126.4	29.1				
NER36	8	8	7.915	5.73	82.8	186.5	-31.3	77.09	134.52		
NER36A	8	8	7.967	3.57	211.3	208.3	-44.8	64.85	71.24		
NER37	6	*								28.353	343.5258
NER38	7	7	6.887	7.74	53	49.3	33.1	43.85	75.24		
NER39	9	8	7.907	6	75.4	50.7	38.7	43.95	69.69		
NER40	9	9	8.877	6.05	65.2	193.1	-32.3	73.83	112.64	28.353	343.5264
NER41	6	*								28.353	343.5264
NER42	8	*								28.353	343.5264
NER43	9	7	6.959	4.61	147.5	204.9	-33.3	65.04	92.11		
NER44	9	5	4.856	13.18	27.7	188.3	-56.7	78.68	19.33	28.353	343.5264
NER44B	9	*									
NER45	9	9	8.969	3.01	261.5	200.8	-41.7	70.86	81.82	28.353	343.5268
NER47	9	7	6.807	10.15	31.1	197.1	-23.5	67.35	115.27		
NER48B	9	7	6.945	5.35	110	27.6	53	65.84	54.12	28.353	343.528
NER50	9	9	8.952	3.78	165.6	198	-33.6	70.74	100.78	28.353	343.5284
NER51	11	10	9.903	4.78	92.8	195.1	-32.5	72.50	107.87		
NER52	11	11	10.963	2.64	273.6	26.3	14.3	57.43	108.80		
NER53	6	6	5.953	5.97	105.6	30	56.6	63.41	46.39	28.353	343.5289
NER54	5	4	3.900	14.68	30.1	206.9	-23.3	60.14	100.86	28.353	343.5284
NER55	9	6	5.986	3.28	347.8	23.5	55.3	68.80	46.87	28.353	343.529
NER56	7	7	6.930	6.08	85.4	205.8	-33.6	64.37	90.77	28.354	343.5292
NER57	9	9	8.957	3.55	187.5	28.5	45.8	64.82	69.32	28.354	343.5294
NER58	7	6	5.908	8.37	54.2	198.2	-19.5	64.96	116.91	28.354	343.5298
NER59	9	9	8.820	7.35	44.5	201.6	-27.2	65.60	103.84	28.354	343.53
NER60	7	8	7.893	6.45	65.4	206.1	-10.1	56.03	111.87	28.355	343.53
NER61	7	7	6.912	6.81	68.1	204	-20.1	61.20	107.36		
NER65	9	9	8.821	7.32	44.8	204.6	-57.1	67.44	42.69	28.362	343.53
NER70	7	6	5.950	6.13	100.5	224.4	-32.2	47.93	78.59		
NER70A	5	4	3.945	10.87	54.3	219.6	-32.5	52.21	81.16		
NER71	8	8	7.919	5.61	86	210.8	-37.3	61.10	81.59		
NER72	9	*									
NER73	7	5	4.872	12.4	31.2	31.3	49.7	62.74	61.23		
NER74	9	8	7.947	4.51	133.1	20.6	48.8	71.96	64.13	28.341	343.52
NER75	11	10	9.882	5.28	76.1	20.5	51	71.97	57.78		
NER76B	7	6	5.954	5.86	109.7	29.1	11.5	54.40	107.30		
NER77b	10	8	7.815	8.51	37.9	19.3	52.4	72.79	53.06	28.316	343.50
NER79	8	7	6.937	5.76	94.8	188.3	-51	82.08	46.51		
NER81	8	8	7.877	6.93	56.7	197.3	-57.5	72.58	34.92	28.309	343.50
NER82	8	7	6.859	8.64	42.7	4.8	48.4	85.67	58.48	28.309	343.4957
NER86	8	8	7.956	4.12	158.7	197.2	-43.3	74.35	81.02	28.373	343.5387
NER87	9	8	7.911	5.89	78.3	25.2	48.4	67.95	64.71	28.377	343.5404

Table 2: Expected declinations and inclinations for the Canary Islands compiled from several literature sources; mean of the data set is in bold.

Reference	latitude	longitude	$\Lambda 95$	age range (Ma)	expected declination	expected inclination	error exp. declination	error exp. inclination
Watkin, 1973	81.9	114.4	4.1	5 - 21	353.4	40.2	4.45	5.64
Storevedt et al., 1974	77.8	146.2	5.2	2 - 14	356.2	30.9	5.43	8.34
Najid et al., 1981	85.8	161.5	5.9	5 - 13	359.8	41.9	6.47	7.85
Quidelleur and Valet, 1996	84.9	166.9	5.8	0 - 1	359.7	40.7	6.32	7.90
Carracedo and Soler 1995	84.9	173.9	7.0	0 - 2	359.0	40.1	7.60	9.64
		Mean	4.7		357.6	38.8		

Table 3: Paleomagnetic data from the area sampled in Anaga massif. See table 1 for explanations.

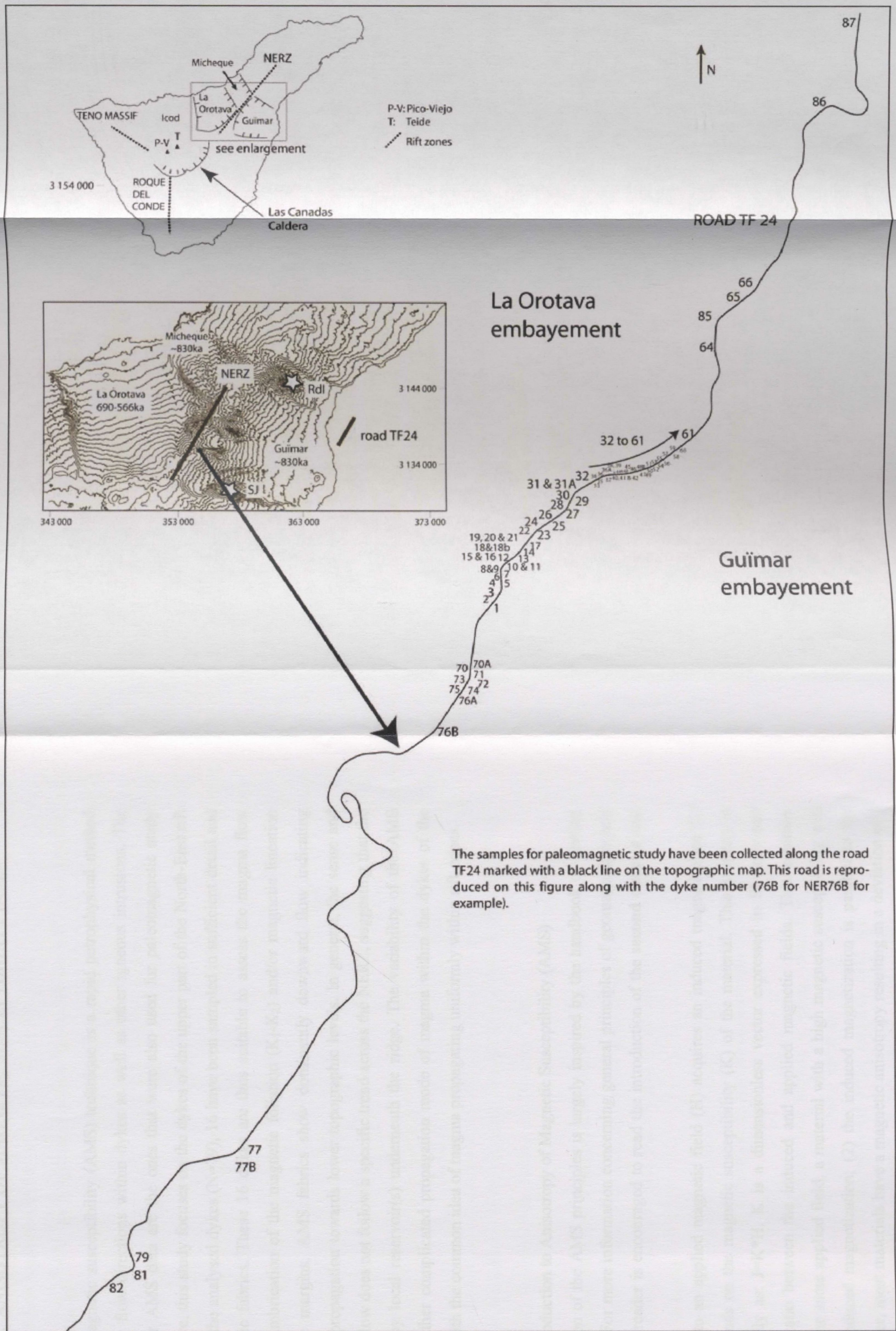
	site							VGP		Site location	
	No	N	R	$\alpha 95$	κ	D	I	LAT	LONG	LAT	LONG
LF2	7	7	6.9	6.23	81	13.3	-43.8	28.511	343.808	34.38	149.29
LF3	6	6	6	5.55	122	1.7	-29.3	28.512	343.808	45.780	161.460
LF4	6	5	5	4.65	217	15.5	-54.8	28.507	343.677	24.51	149.81
LF5	6	5	5	5.73	143	198	5.5	28.508	343.677	54.41	132.22
LF6	6	6	6	5.17	141	191	42.4	28.509	343.677	35.96	151.63
LF7	altered samples										
LF8	7	7	6.9	9.06	39	14.2	41.9	28.511	343.677	76.56	89.28
LF10	6	5	5	8.56	65	8.9	-46.3	28.508	343.684	39.880	233.650

Table 4: Paleomagnetic data from Leonhardt and Soffel (2006).

	site						VGP		Site location	
	No	N	$\alpha 95$	κ	D	I	LAT	LONG	LAT	LONG
BC1	6	5	8.4	97	340.4	25.7	28 319	343 130	66.5	218.1
BC2	6	5	9.2	71	97.8	9.8	28 321	343 131	-4.5	65.1
C1	5	4	8.7	113	21	30.7	28 320	343 143	67.3	100
C2	4	4	5.1	322	24.5	56	28 323	343 142	67.8	45.1
C3	5	5	6.6	135	11.4	58.5	28 328	343 144	75.6	21.1
LM1	4	4	4.6	398	161.7	-51	28 293	343 140	-73.8	89.8
LM2	5	5	8.1	104	184.7	-38.3	28 293	343 146	-82	309.8
LM3	4	4	3.8	575	175.5	-32.2	28 296	343 148	-78.4	5
LM4	6	4	4.8	370	212.7	-38.1	28 298	343 152	-59.6	258.8
LM5	4	4	4.1	513	169.4	-31.1	28 299	343 155	-74.9	25.7
UM1	6	5	6.8	129	11.5	39.6	28 308	343 160	78.1	99.9
UM2	6	6	5.8	134	335.7	50.9	28 305	343 170	68.7	268.1
UM3	5	5	3.6	441	19.3	52.6	28 300	343 176	72.7	52

P1	5	5	4.7	270	322.9	-67.5	28 344	343 145	4.7	185.9
P2	4	4	4.1	512	1.8	23.3	28 344	343 145	73.7	156.9
P3	5	5	7.1	118	357.9	24.2	28 344	343 145	74.2	170.7
P4	5	5	7.7	100	10.4	26.7	28 344	343 145	72.8	126.8
P5	6	5	6.1	167	354.5	24	28 342	343 143	73.4	182.3
P6	6	4	6.6	194	6.9	32.2	28 339	343 141	77.4	131.4
P7	4	4	3	962	355	38.6	28 339	343 141	82	198.8
P8	4	4	8.7	113	358.2	35.4	28 342	343 143	81.1	174.1
P9	4	4	7.9	136	5.6	22.5	28 339	343 141	72.6	144.6
P10	4	4	8.3	122	11.3	41.7	28 339	343 141	79	93.7
P11	5	4	4.9	350	9.9	43.2	28 347	343 145	80.6	90.7
TA1	4	4	2.2	1703	0.6	40.9	28 339	343 130	85.1	156.7
TA2	5	5	6.2	161	184.7	-36.8	28 348	343 118	-81.1	313.5
Mean					3.2	38				

APPENDIX 1: MAP OF PALEOMAGNETIC SITES.



CHAPTER 3: ANISOTROPY OF MAGNETIC SUSCEPTIBILITY (AMS)

Abstract

The anisotropy of magnetic susceptibility (AMS) technique is a rapid petrophysical method used to infer magma flow directions within dykes as well as other igneous intrusions. The samples collected for AMS data are the ones that were also used for paleomagnetic study (**chapter 2**). Therefore, this study focuses on the dykes of the upper part of the North-East rift zone (NERZ). From the analysed dykes (N=79), 16 have been sampled in sufficient detail and have normal magnetic fabrics. These 16 dykes are thus suitable to assess the magma flow direction, using the imbrication of the magnetic foliation (K_1 - K_2) and/or magnetic lineation (K_1) along the dyke margins. AMS fabrics show dominantly downward flow indicating deflation or magma propagation towards lower topographic levels. In general, the sense and direction of magma flow does not follow a specific trend across the NERZ, suggesting that the dykes are supplied by local reservoir(s) underneath the ridge. The variability of the AMS fabrics suggests a rather complicated propagation mode of magma within the dykes of the NERZ, contrasting with the common idea of magma propagating uniformly within rift-zones.

1. Introduction:

1.1. Brief introduction to Anisotropy of Magnetic Susceptibility (AMS)

This short introduction of the AMS principles is largely inspired by the handbook of *Tarling and Hrouda (1993)*. For more information concerning general principles of geomagnetism and rock magnetism, the reader is encouraged to read the introduction of the **second chapter** and reference therein.

A material exposed to an applied magnetic field (H) acquires an induced magnetization (J) whose strength depends on the magnetic susceptibility (K) of the material. This property is written mathematically as: $J=K \cdot H$. K is a dimensionless vector expressed in SI units and corresponds to the ratio between the induced and applied magnetic fields. This equation implies that (1) for the same applied field, a material with a high magnetic susceptibility will acquire a stronger induced magnetization; (2) the induced magnetization is parallel to the applied field. However, most materials have a magnetic anisotropy resulting in a deviation of J from H , thus creating an induced magnetization that varies in three dimensions. In that case,

the magnetic susceptibility is represented by a 3*3 matrix with three principal susceptibility axes K_1 , K_2 , K_3 , where K_1 is the maximum, K_2 is the intermediate and K_3 the minimum susceptibility axes. The anisotropy of magnetic susceptibility (AMS) is a measure of the orientation of these principal susceptibility axes that can be visually represented by a susceptibility ellipsoid (Hrouda, 1982). The shape of the susceptibility ellipsoid is defined by the “shape factor” (T) where $-1 < T < 1$. A prolate shape is defined by $K_1 > K_2 = K_3$ ($T < 0$), an oblate shape by $K_1 = K_2 > K_3$ ($T > 0$) and a triaxial shape corresponds to the intermediate configuration, i.e. $T \approx 0$ (**Fig.1**). The degree of anisotropy is given by $P = K_1/K_3$. The corrected degree of anisotropy P' is, however, preferentially used to measure the strength of the magnetic fabric (see Jelínek, 1981). A material with a value of $P' = 1$ will have an isotropic fabric while a value of $P' = 1.1$ corresponds to a material with 10% anisotropy etc. A degree of anisotropy P' of 0-5% is referred as a weak anisotropy, while a moderate anisotropy represents values between 5-10%, and a strong anisotropy between 10-20%. Any value of P' above 20% indicates a very strong anisotropy (e.g. Tarling and Hrouda, 1993; Rochette et al., 1999; Aubourg, 2002; Petronis and Geissman, 2009).

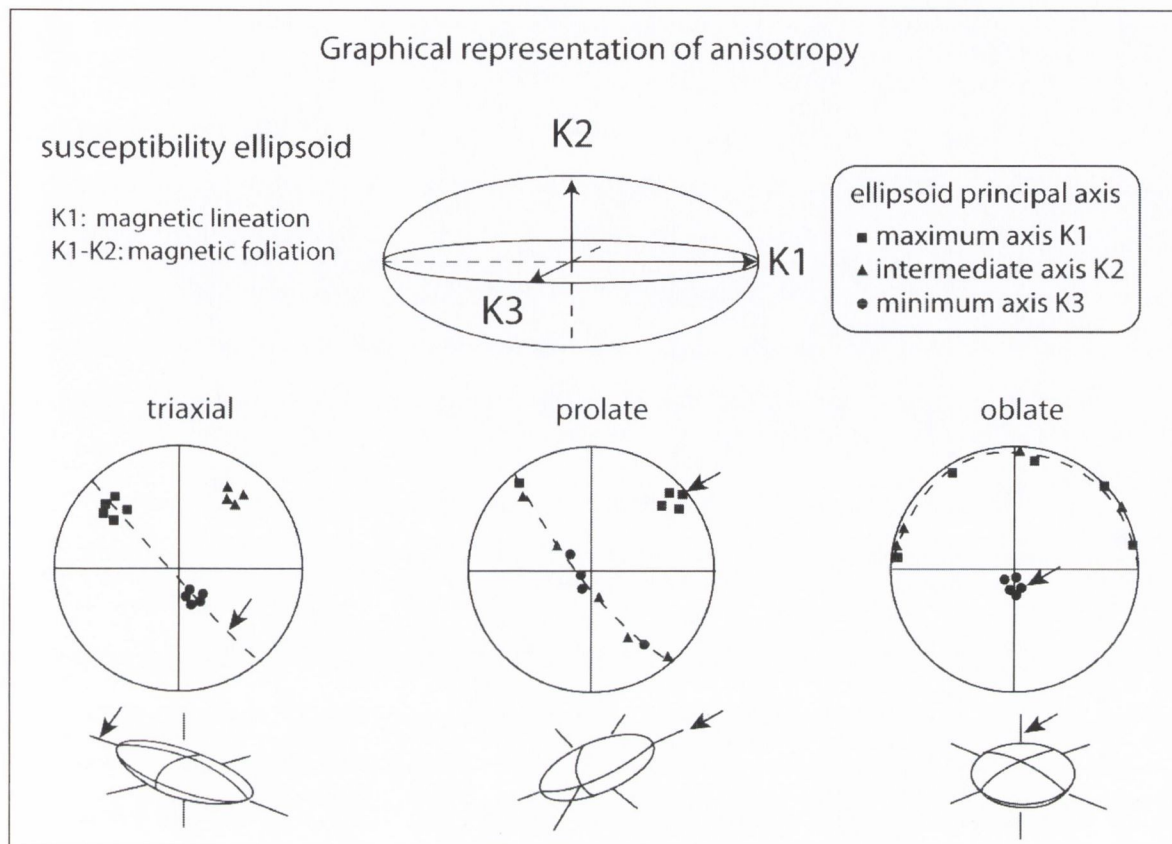


Figure 1: Graphical representation of the magnetic susceptibility ellipsoid, defined by three principal axes, K_1 , K_2 , K_3 (modified from Tarling and Hrouda, 1993).

Materials can have different behaviour when they are exposed to a magnetic field:

- A diamagnetic material will acquire a small induced magnetization J_i opposite to the applied magnetic field that reduces to zero once the field is removed. Example: quartz.
- A paramagnetic material will acquire a small induced magnetization parallel to the magnetic field which decreases to zero when the magnetic field is removed. Examples: olivine, biotite.
- A ferromagnetic (s.l.) material will acquire a strong induced magnetization parallel to the magnetic field. These materials are of major interest for paleomagnetic study as after removal of the magnetic field, they do not reduce to zero but preserve a record of the applied field. Examples: hematite, magnetite.

AMS is often used as a proxy for magma flow direction (e.g., *Ellwood, 1978; Hrouda, 1985; Rochette et al., 1991; Stevenson et al., 2007*). As early ascent structures are likely obliterated during magmatic flow, only the final stages of magma flow can be inferred from petrographic and AMS studies (*Hutton, 1992; Vigneresse and Clemens, 2000*). The K_1 axis and the plane (K_1 - K_2) define, respectively, the magnetic lineation and the magnetic foliation plane. The foliation and the lineation describing the magnetic fabric can be defined as normal or inverse (*Hrouda, 1982; Rochette et al., 1991; Rochette et al., 1999*). A normal fabric implies that the magnetic foliation is parallel to the dyke plane, while an inverse fabric implies that the magnetic foliation is perpendicular to the dyke plane. Any configuration in between these two end members is referred to as an intermediate fabric. A normal fabric is related to magmatic flow with the lineation K_1 defining the flow direction, while the origin of an inverse fabric is still debated (*Park et al., 1988; Potter and Stephenson, 1988; Hargraves et al., 1991; Rochette et al., 1991; Rochette et al., 1992; Dragoni et al., 1997; Rochette et al., 1999; Aubourg et al., 2002; Ferré, 2002; Hrouda et al., 2002; Cañón-Tapia and Chávez-Álvarez, 2004; DeFrates et al., 2006; Chadima et al., 2009*). Recent works highlight the role of magnetic fabric imbrication along dyke margins that help determine the direction and sense of magma flow (*Moreira et al., 1999; Aubourg, 2002; Geoffroy et al., 2002; Aubourg et al., 2008; Fig.2*). In this study, we use the orientations of the magnetic lineation K_1 and the magnetic foliation plane (K_1 - K_2), relative to the dyke orientation to determine flow direction, following the method of *Aubourg et al. (2002)*.

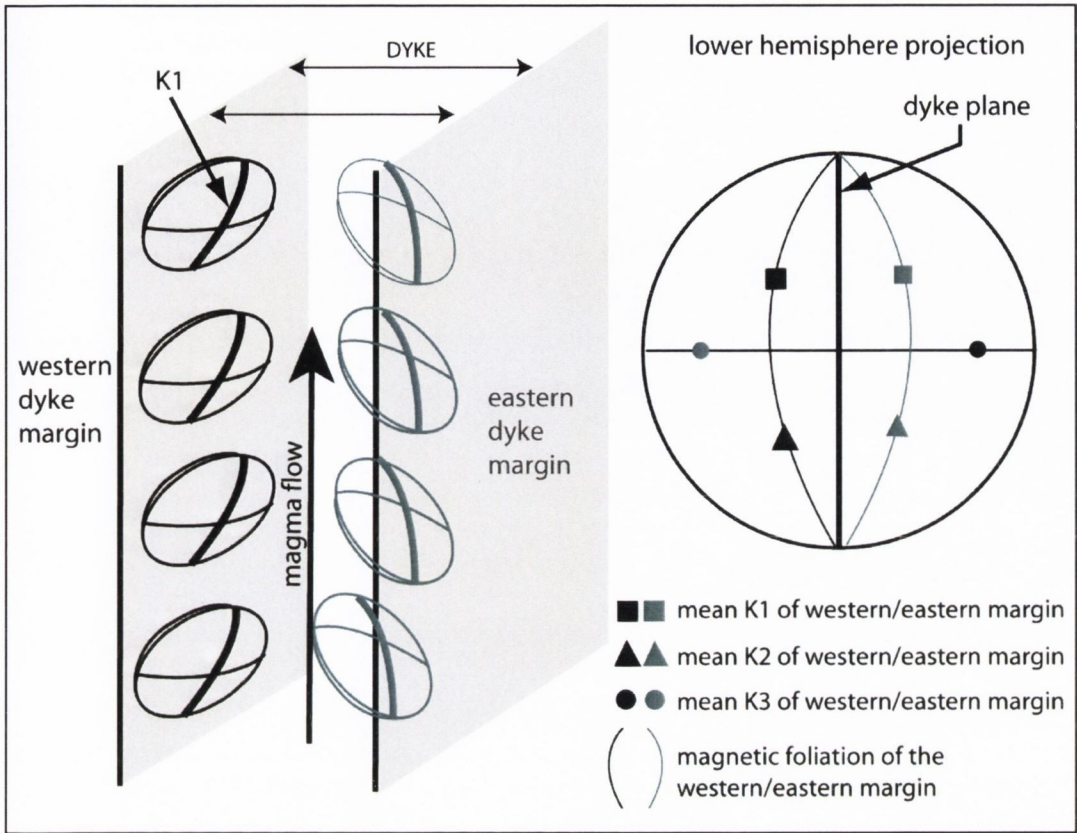


Figure 2: Representation of the magnetic fabric imbrications along dyke margins. Modified after Tauxe (1998).

1.2. Sampling area

The samples used for AMS analyses are those collected along the road TF24 for paleomagnetic study (**chapter 2; Fig.3A**). As previously explained (**chapter 2**), it is not possible to use the samples collected in the galerias due to orientation uncertainties. We are thus focusing on dyke emplacement in the upper part of the North-East rift-zone (NERZ). Three major lateral collapses occurred along the rift (**Fig.3A**). The Micheque collapse is thought to be the oldest and is estimated at about 830ka (Carracedo *et al.*, 2009). The Güímar collapse seems to be approximately contemporaneous with the Micheque event. The La Orotava collapse is the last one to have occurred on the NERZ and is bracketed between 690ka and 566ka (Carracedo *et al.*, 2009).

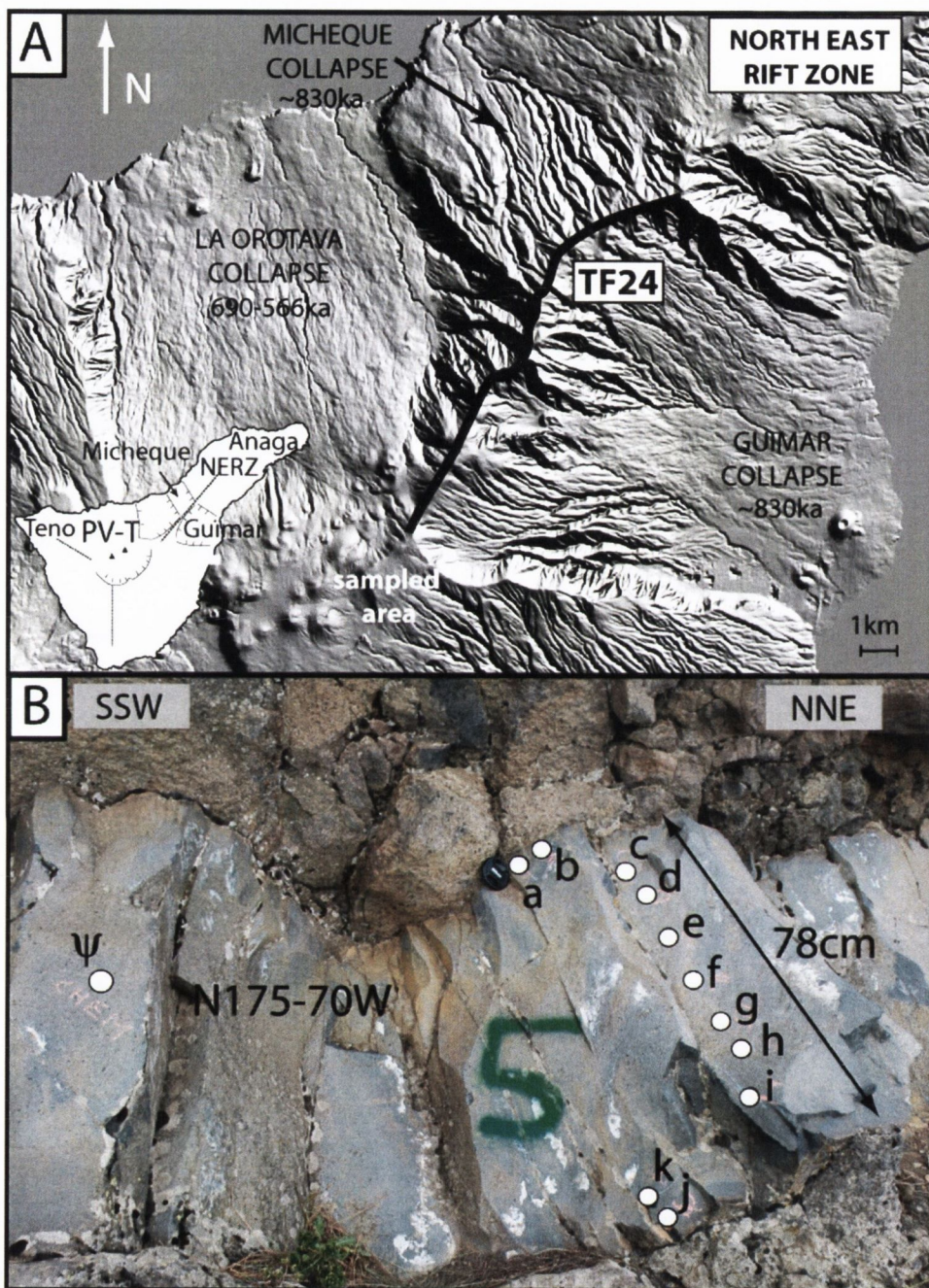


Figure 3: A: DEM of the study area; the black line corresponds to the sampled section for the AMS study, PV-T: Pico Viejo-Teide complex (DEM: courtesy of R.Paris, source: Grafcan©). B: Example of a representative dyke drilled for our AMS study (dyke NER5). The a-j letters correspond to drilled holes for AMS and paleomagnetism studies, ψ corresponds to a specimen drilled for the geochemistry study.

1.3. Aims

This AMS study was carried out to assess the direction and sense of magma flow within the dykes of the NERZ to understand better how magma emplacement occurred within the rift. Two main hypothesis are to be tested: (1) the rift is fed by a central reservoir below the Pico Viejo-Teide complex, or (2) the rift is supplied by shallow reservoir(s) within the ridge. In the first case, we expect magma to propagate laterally from the SW to the NE as the Pico Viejo-Teide complex is to the SW of the NERZ (**Fig.3A**). Alternatively, if the rift is fed by shallow reservoir(s) within the ridge, we should expect to see various flow patterns i.e. lateral, inclined and/or vertical, without following any uniform or specific direction.

2. Field analyses

2.1. Brief dyke description

A full description of dyke morphologies and features can be found in **chapter 1**, and the following review merely summarises key information.

The majority of the dykes are between 30 and 90cm-thick and are basaltic. The occurrence of inverse and intermediate fabrics is common in basaltic dykes (*Knight and Walker, 1988; Tauxe et al., 1998; Rochette et al., 1999; Geoffroy et al., 2002; DeFrates et al., 2006*).

2.2. Thin sections

Six petrographic groups have been determined using hand samples and thin sections (see **chapter 1**). The main minerals found are feldspar, olivine and pyroxene with rare brown hornblende. Since these minerals are paramagnetic, they contribute to the total AMS, although their magnetic susceptibilities are several orders less than those of the ferromagnetic minerals such as magnetite (*Richter and van der Pluijm, 1994*).

The main six groups are: (1) aphyric/subaphyric dykes (aph)

(2) feldspar-rich dykes (Fsp rich)

(3) clinopyroxene and feldspar-rich dykes (Cpx = Fsp)

(4) ankaramitic dykes (anka)

(5) olivine and clinopyroxene-rich dykes (Ol = Px)

(6) pyroxene-rich dykes (Px-rich)

Oxide phases are relatively abundant (up to 40% of the total phenocryst content in some samples) and magnetite has been identified as the main phase using reflective light microscopy (**Fig.4**). No exsolution lamellae are observed on the magnetite grains, indicating that they were stable after growth and cooling. The thin-section textures indicate that the magnetite grains likely crystallized during or after dyke emplacement and cooling. This ferromagnetic phase, based on petrographic observations as well as rock magnetic experiments, controls the AMS fabric.

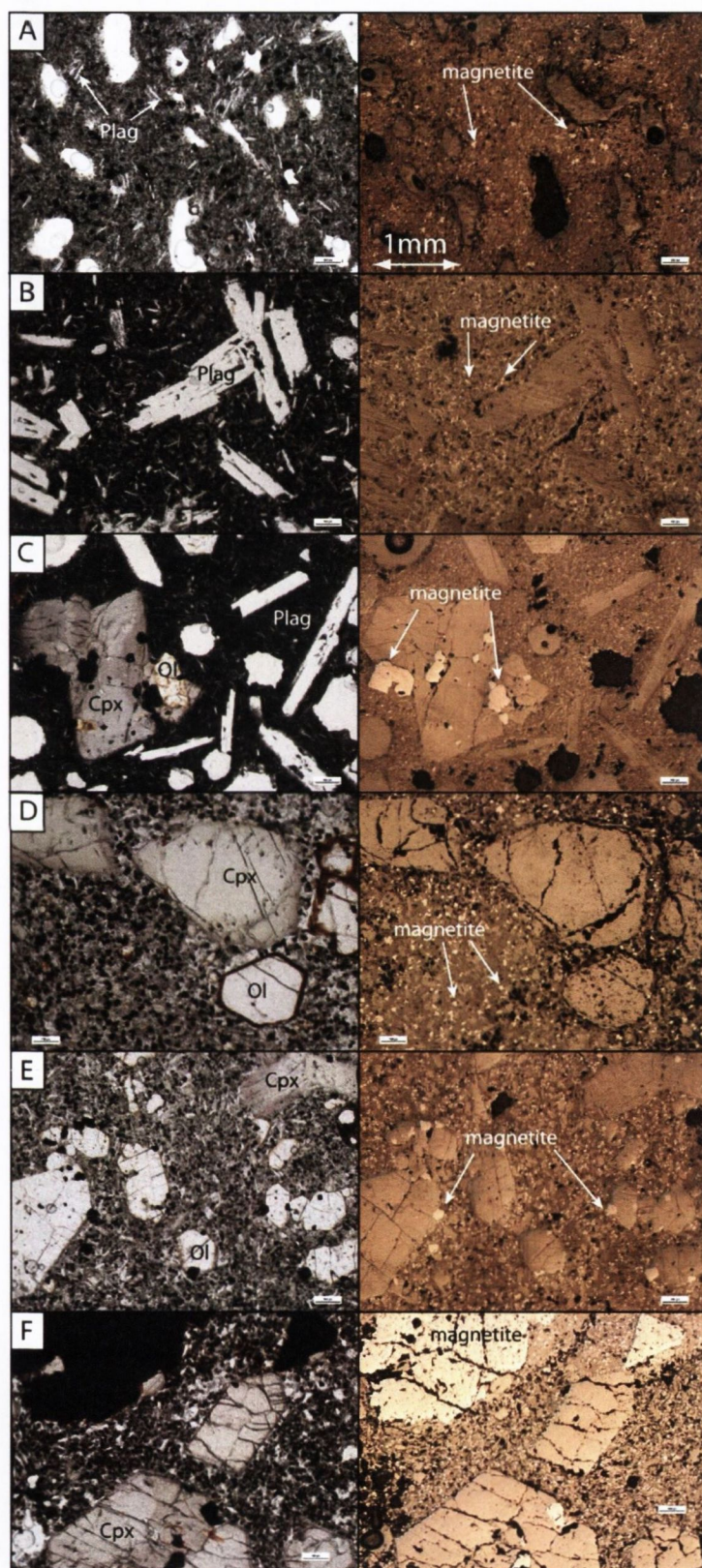


Figure 4: Pictures of representative thin sections of the different petrographic groups: Plane polarized light on the left and reflective light on the right. Cpx: clinopyroxene, Ol: olivine, Plag: plagioclase. **A:** aphyric/subaphyric group **B:** Feldspar-rich group **C:** Feldspar + pyroxene group **D:** Ankaramite **E:** olivine + pyroxene group **F:** Pyroxene-rich group. Plag = plagioclase; Cpx = clinopyroxene; Ol = olivine.

3. Magnetic analysis

3.1. Sampling

The sampling method followed the procedure outlined in the paleomagnetic chapter (**chapter 2**) as the same cores were used. Typically for paleomagnetic study, 9 to 11 samples are sufficient to provide robust statistics. However, for AMS, more samples are necessary to infer the direction and sense of the magma flow. The samples were usually collected along section perpendicular to the dyke walls and a few samples were drilled parallel to the margins (**Fig.3B**). AMS measurements were carried out at the University of New Mexico (USA) using an AGICO Kly4-S Kappa-Bridge, operating at 875Hz with an applied field of 300 A/m.

3.2. Rock magnetism and magnetic fabric: results

3.2.1. Rock magnetism

Rock magnetism experiments were conducted to determinate the minerals that carry the remanent magnetization and contribute to the AMS fabric (**chapter 2**). Standard experiments include alternating field (AF) demagnetization, IRM acquisition (Isothermal Remanent Magnetization), thermal demagnetization (Thermal Remanent Magnetization) and the investigation of the variation of the susceptibility vs temperature. These analyses showed that single domain (SD) and pseudo-single domain (PSD) grains of cubic titanomagnetite constitute the main magnetic phase of the dykes. We also note that the relatively high magnetic susceptibility magnitude (K_m ; **table 1** and **Fig.5C**) is consistent with the presence of a titanomagnetite phase (*Knight and Walker, 1988; Hargaves et al., 1991; Rochette et al., 1991*). Titanomagnetite phases grow late in the crystallization process (*Hargaves et al, 1991*), which is consistent with the thin section observations (**section 2.2**). The contribution of paramagnetic grains (i.e. feldspar, olivine, pyroxene and hornblende) to the total anisotropy is thus negligible.

The SD and PSD domains are more suitable for paleomagnetic data than for AMS study (*Butler, 1992*). Indeed, contrary to multi-domains grains (MD), the short axes of SD grains are parallel to the maximum susceptibility (K_1), while the long axes are parallel to the minimum susceptibility (K_3) (*Hrouda, 1982; Potter and Stephenson, 1988*). This configuration can be responsible for the occurrence of inverse or intermediate fabrics.

3.2.2. AMS fabric (Fig.5)

It was possible to determinate the AMS fabric of 72 dykes of the 79 analysed as 7 dykes yielded scattered results. The corrected degree of anisotropy (P') of these 72 dykes is relatively low and varies from 0.4% ($P'=1.004$) to 10.8% ($P'=1.108$), with an average of 3.3% ($P'=1.033$). Of these dykes, 46% show a prolate shape, 44% an oblate shape and the remaining 10% yield roughly triaxial shapes. 56% of the dykes show a normal fabric, 18% an inverse fabric, and 26% an intermediate fabric. Of the normal fabrics, 57% are of oblate shape, 35% of prolate shape and 8% of roughly triaxial shape. Of the inverse fabrics, 38% are of an oblate shape, 54% of a prolate shape, and 8% of a roughly triaxial shape. From the intermediate fabric, 21% are of an oblate shape, 63% of a prolate shape, and 16% of a triaxial shape. No clear relation between anisotropic parameters is observed on **figure 5**, except a slight positive correlation between the corrected degree of anisotropy and the magnetic susceptibility (**Fig. 5C**).

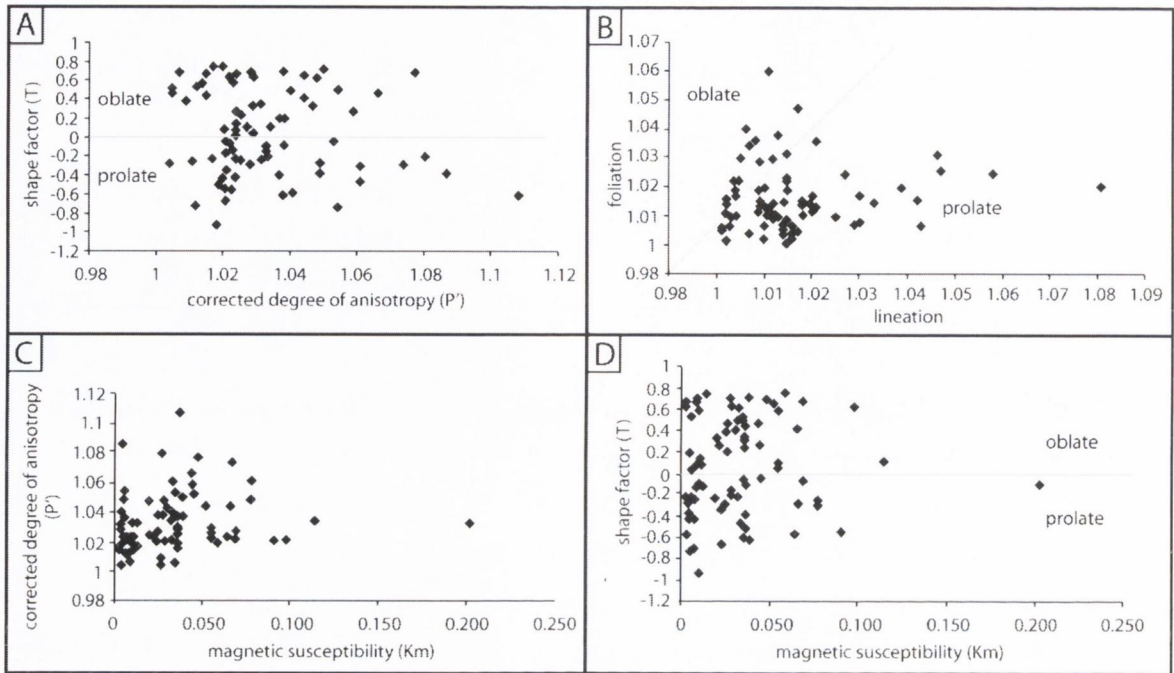


Figure 5: Plots showing the relationships between the selected anisotropy parameters. **A:** Shape factor (T) vs corrected degree of anisotropy (P') **B:** Magnetic foliation vs magnetic lineation (Flinn diagram) **C:** Corrected degree of anisotropy (P') vs magnetic susceptibility (Km) **D:** Shape factor (T) vs magnetic susceptibility (Km), for all the AMS data. A modest correlation exists between the corrected degree of anisotropy and the magnetic susceptibility (plot C), while no clear relation is observed between the other parameters.

The petrographic information was added to plots 5A and 5C to evaluate if the degree of anisotropy maybe linked with dyke petrography (Fig.6). The feldspar-clinopyroxene group is dominantly prolate shape, while the aphyric, ankaramite and olivine-cpx groups are dominantly oblate shape. The ankaramite and px-rich groups have higher magnetic susceptibilities. A plot similar to figure 6 has been drawn with the chemical composition (not presented here), but little systematic relationships can be gleamed given the low range of compositional diversity-dykes fall mainly into the basanite field of the TAS diagram; chapter 1.

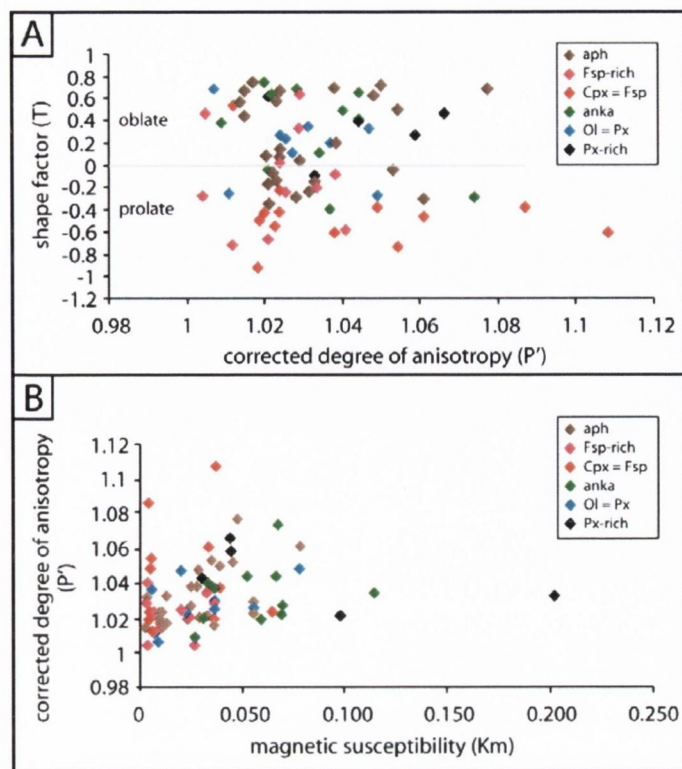


Figure 6: Selected anisotropy diagrams. **A:** Shape factor (T) vs corrected degree of anisotropy (P') **B:** Corrected degree of anisotropy (P') vs magnetic susceptibility (K_m), for each petrographic group. Aph: aphyric dykes; Fsp-rich: feldspar-rich dykes; Cpx = Fsp: clinopyroxene and feldspar-rich dykes; anka: ankaramitic dykes; Ol = Px: olivine and clinopyroxene-rich dykes; Px-rich: pyroxene-rich dykes. The shape factor (6A) and the magnetic susceptibility (6B) seem to be slightly linked to the petrography.

3.3. AMS fabrics and flow direction.

3.3.1. Method.

To infer magma flow direction and sense, we selected only dykes with a normal magnetic fabric and with a sufficient number of samples along the margins. Consequently, the number of dykes studied in detail is 16, from a total set of 79 dykes analysed. In addition, sections of five dykes (with normal fabrics) were considered as insufficient samples were collected along the margins. All AMS data and magmatic flows determined from the imbrications of the magnetic foliations and/or lineations along the dyke margins and from the dyke-sections are outlined in **table 2**.

The magnetic lineation (K_1) has been widely used to infer magma flow directions (e.g., Rochette *et al.*, 1991, Tauxe *et al.*, 1998). More recently, the imbrication of the magnetic foliation planes along the margin has been preferred, as dykes displaying an oblate shape can have their K_1 and K_2 axes reversed (e.g., Moreira *et al.*, 1999; Aubourg, 2002). Using this principle, Geoffroy *et al.* (2002) developed a method to determine the flow vector that, however, is sometimes inappropriate to apply if the imbrications of the magnetic foliations are

not symmetric (e.g., *Pueyo Anchuela et al., 2006*). This is a common case with the dykes of the NERZ. Consequently, we only use geometric methods based on the orientation of the magnetic foliation and lineation along the margins relative to the dyke orientation following *Aubourg et al. (2002)*.

We divided our data into two classes (**Fig.7**) and the results for all the dykes studied are shown in **figure 8**. Class 1 comprises dykes with an overall oblate shape and class 2 includes dykes with an overall prolate shape. While a dyke presents an overall oblate/prolate shape, it can happen that one of the margins (or sometimes even both) shows a prolate/oblate shape. Within “class 1” (oblate shape), three sub-groups are found:

Class 1-A: one margin displays a magnetic foliation, or lineation, parallel to the dyke margin and the other margin displays a magnetic foliation, or lineation, oblique to the dyke margin.

Class 1-B: the magnetic foliations, or lineation, along each margin are oblique (slightly or strongly) to the margin.

Class 1-C: the magnetic foliations, or lineation, along each margin are parallel to the margins.

Within the “class 2” (overall prolate shape), two sub-groups are found:

Class 2-A: one margin has an oblate fabric and the other has a prolate fabric, strongly oblique to the dyke margins.

Class 2-B: both margins present a prolate or an oblate fabric.

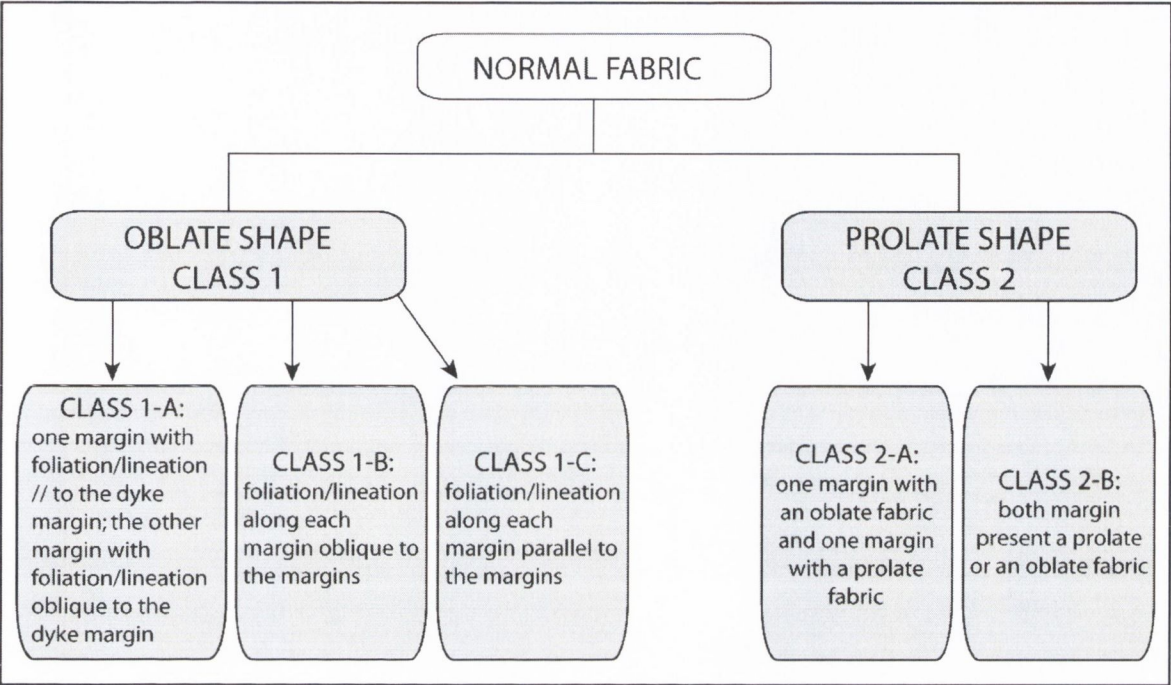
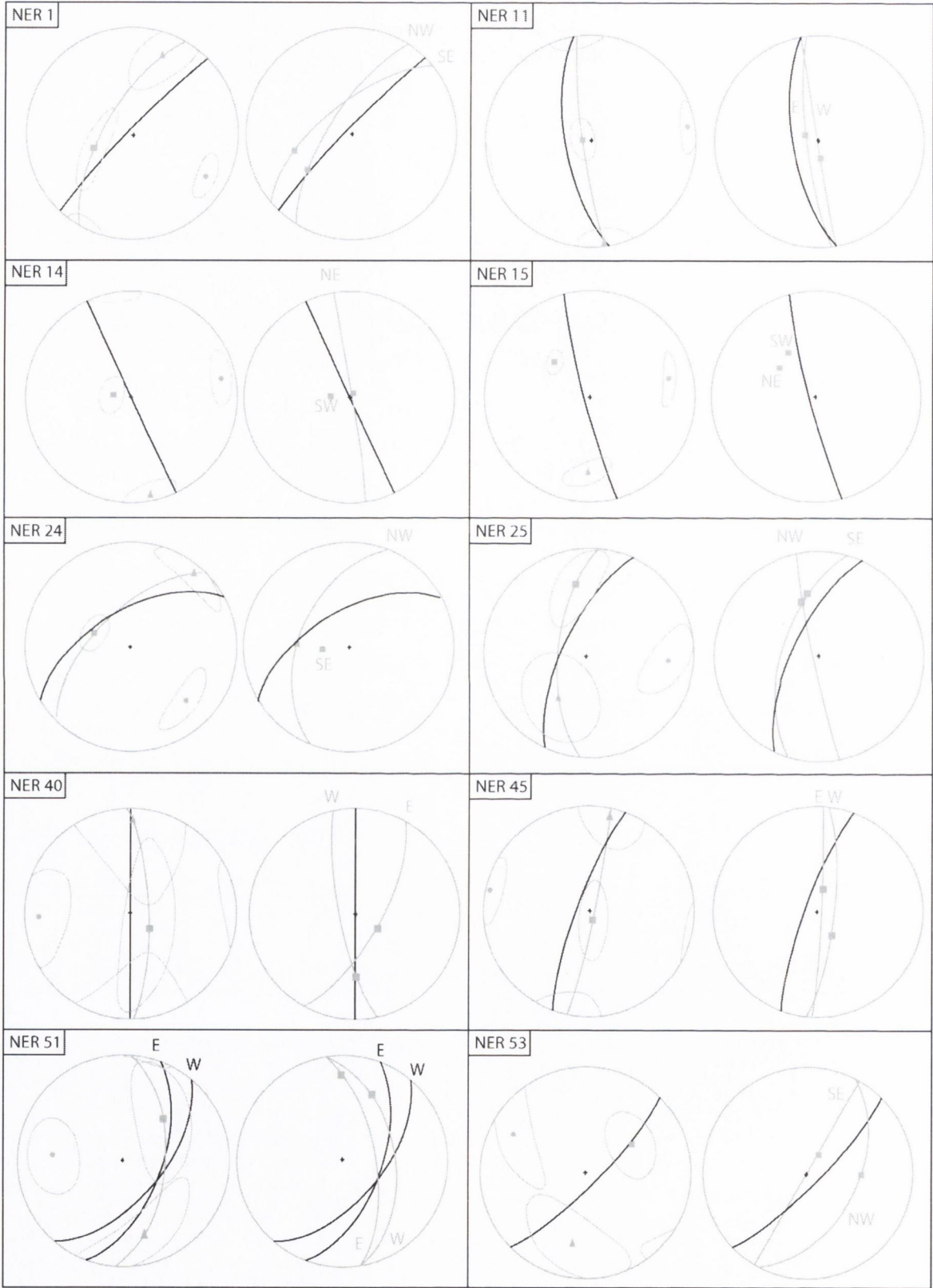


Figure 7: AMS fabric classification. The fabrics are divided in two main classes: dykes with an overall oblate shape (class 1) and dykes with an overall prolate shape (class 2). The two classes are sub-divided according to the orientation of the magnetic fabric related to the orientation of the dyke margins (Class 1-A, 1-B, 1-C), or according to the shape of the magnetic fabric along each margin (Class 2-A and 2-B).



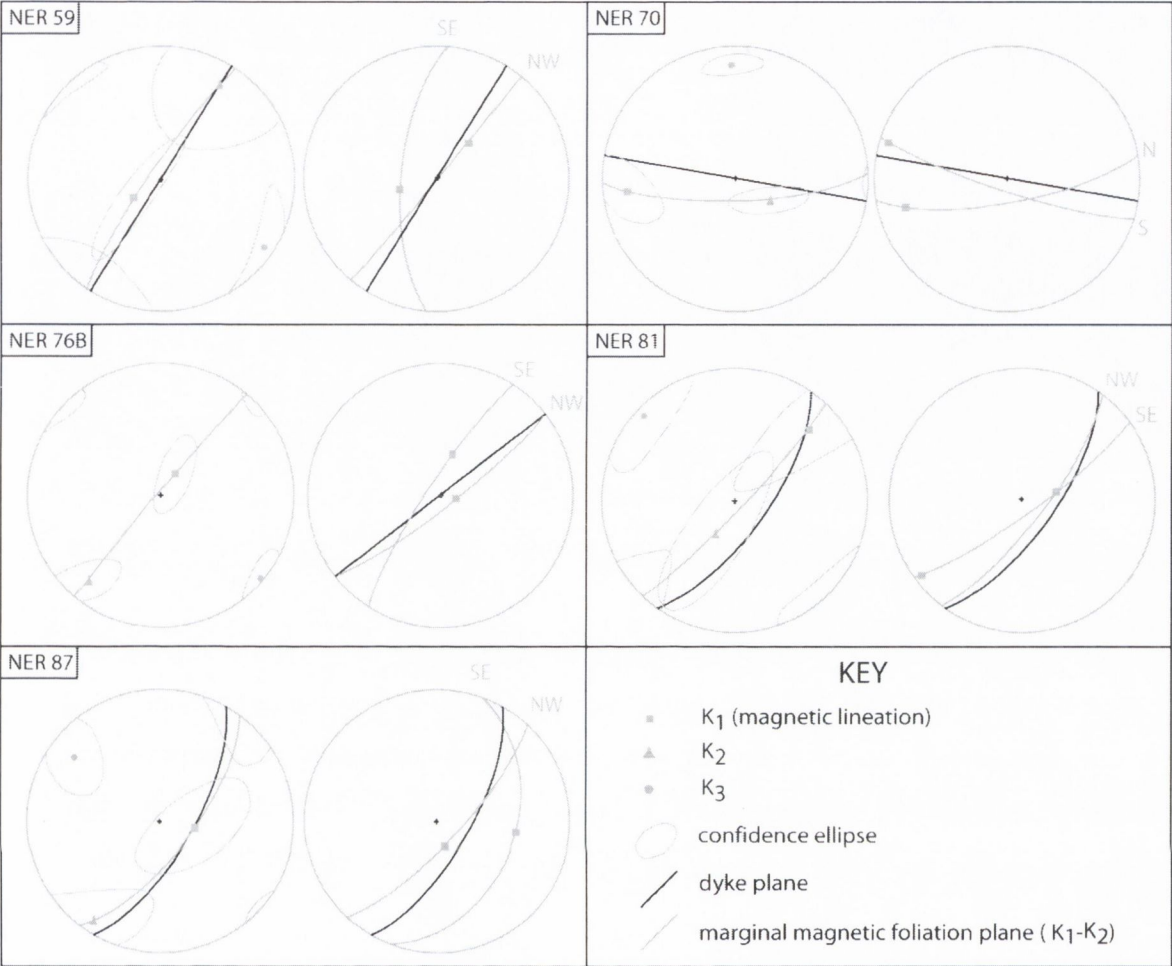


Figure 8: Equal area stereoplots (lower hemisphere projection) of the dykes studied that display a normal AMS fabric. The plots on the left are the AMS fabrics for the whole dyke (i.e. margin and centre) and the plots on the right correspond to the AMS fabric for the margins (e.g., the right plot for the NER1 shows the AMS fabric for the NW and SE margin).

3.3.2. Inferred flow direction

3.3.2.1. Class 1

We used the geometry and the imbrications of the magnetic foliation and lineation trend along the dyke margins to infer the direction and the sense of the flow following *Aubourg et al. (2002)* (section 3.3.1). We note that, sometimes, the plunge of the magnetic lineation K_1 is not consistent with the flow direction deduced from the imbrication of the magnetic foliation plane. One explanation is that the degree of anisotropy is low and that the K_1 and K_2 axes yield

similar orientations between specimens from the same dyke, as oblate fabric are characterized by $K_1 \approx K_2$ and the confidence ellipses overlapped (e.g. NER25 and NER59; **table 2**).

3.3.2.1.1. Class 1-A (**Fig.9**)

The occurrence of one magnetic plane parallel to a margin, and the other oblique to the margin, can be due to lateral shearing along the margins (e.g. *Clemente et al., 2007*), or due to the displacement of the country rock along one margin during dyke emplacement.

The arrangement of the dips of the magnetic foliation planes of three dykes (NER 25, 59 and 76B) shows a downward-directed oblique flow. The directions and senses are specified in **figure 9** and **table 2**.

The dip of the magnetic foliation plane along each margin, the intersection of K_1 - K_2 , and the lineations of the dyke NER 70 indicate a sub-horizontal flow towards the west.

Even though the dyke NER81 has an overall oblate magnetic fabric, the south-east margin is prolate with a shallow K_1 . The K_1 - K_2 of the NW margin is parallel to the dyke plane. This arrangement suggests sub-horizontal flow along the dyke strike. Note that if we determined the magnetic foliation (K_1 - K_2) of the south-east margin that is prolate, the resulting plane dips towards the south. The two magnetic foliation planes of each margin would then indicate a downward flow.

To summarise, for the dykes of “class 1-A”, the magnetic foliation planes along each margin indicate downward flow. One dykes show downward flow towards the SW and one shows downward flow towards the north. NER 70 and 81 show sub-horizontal flows.

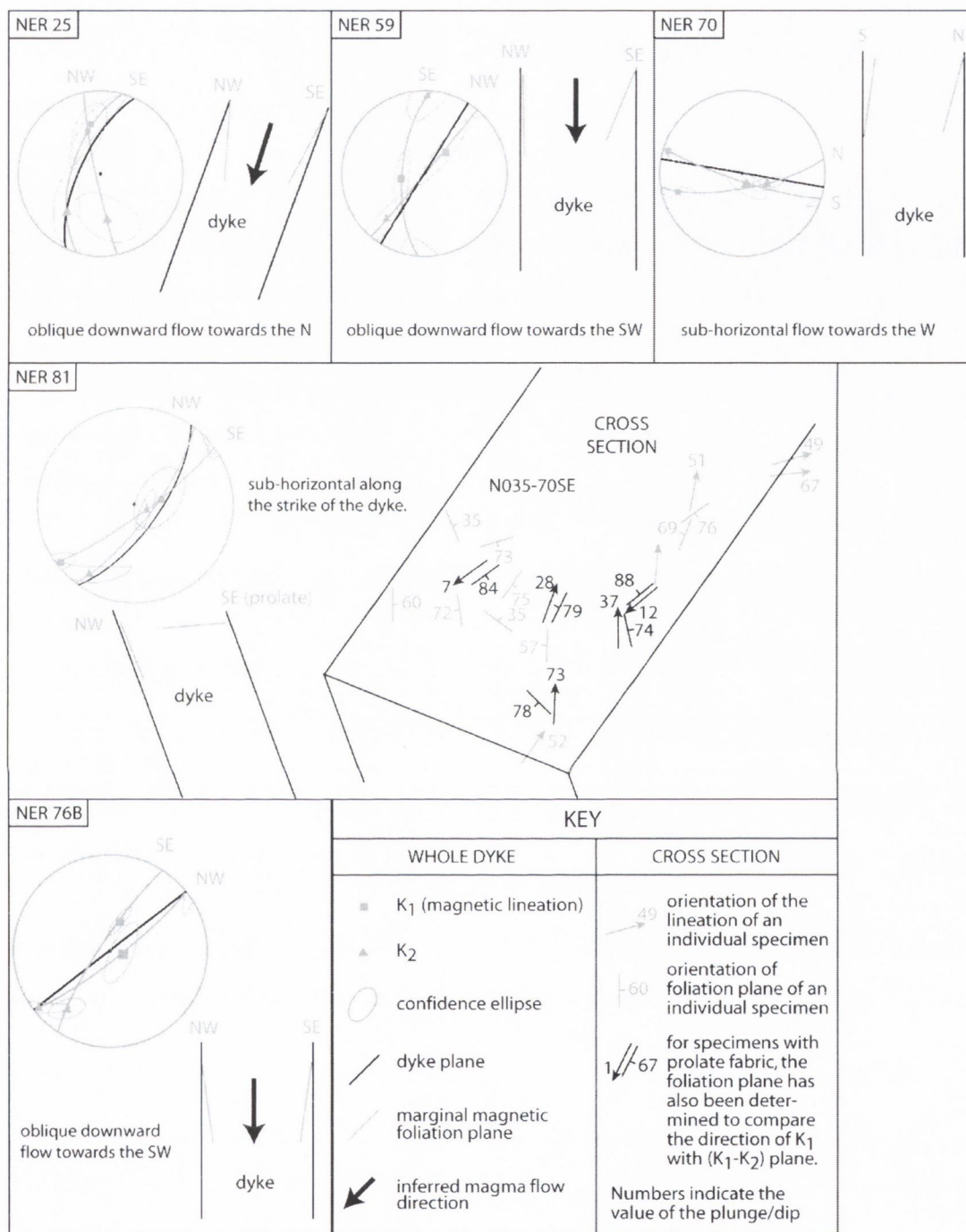


Figure 9: Representation of AMS fabric for each dyke that belongs to the class 1-A (one margin displays a magnetic foliation/lineation, parallel to the dyke margin; the other margin displays a magnetic foliation/lineation oblique to the dyke margin). The magnetic fabric patterns of the dykes “class 1-A” seem to indicate downward oblique flow, or sub-horizontal flow.

3.3.2.1.2. Class 1-B (Fig.10)

The dips of the K_1 - K_2 planes along each margin of the dyke NER1 are almost parallel. The lineations along each margin are consistent and steep. Thus, the arrangement of the fabric suggests a steep flow, either upward or downward.

NER5 is similar to NER81 (class 1-A) in the sense that one margin has an oblate fabric almost parallel to the dyke margin while the other margin has prolate fabric with a shallow inclination. The arrangement of the marginal fabrics can be thus interpreted as sub-horizontal flow along the dyke strike. Note that if we determine the magnetic foliation plane of the eastern margin that is prolate, this plane dips towards the west and is almost parallel to the (K_1 - K_2) of the western margin, indicating an oblique flow. The individual magnetic fabrics across the dyke seems also to show an oblique flow.

The eastern margin of NER40 displays a prolate shape with a steep lineation (K_1) while the western margin has an oblate shape with foliation plane dipping steeply. Their arrangement indicates an oblique to sub-vertical upward flow. However, as one of the margin is prolate, it is not possible to infer the direction of the magma flow.

Similar to NER40, NER45 has one oblate margin and one prolate margin. Their arrangement indicates an oblique to vertical flow. However, it is difficult to deduce the direction of the flow. The arrangement of the magnetic fabric along each margin (both oblate) of NER51 likely indicates a downward vertical flow.

The fabrics along each margins of NER87 are both oblate and show an oblique upward flow. The lineations along each margin are not consistent. It is thus difficult to infer any flow direction.

In summary, dykes that belong to the class 1-B indicate either steep or oblique flow. The flows are either downward or upward. However, as observed for some dykes of class 1-A, it is sometimes not possible to establish the flow sense and/or direction.

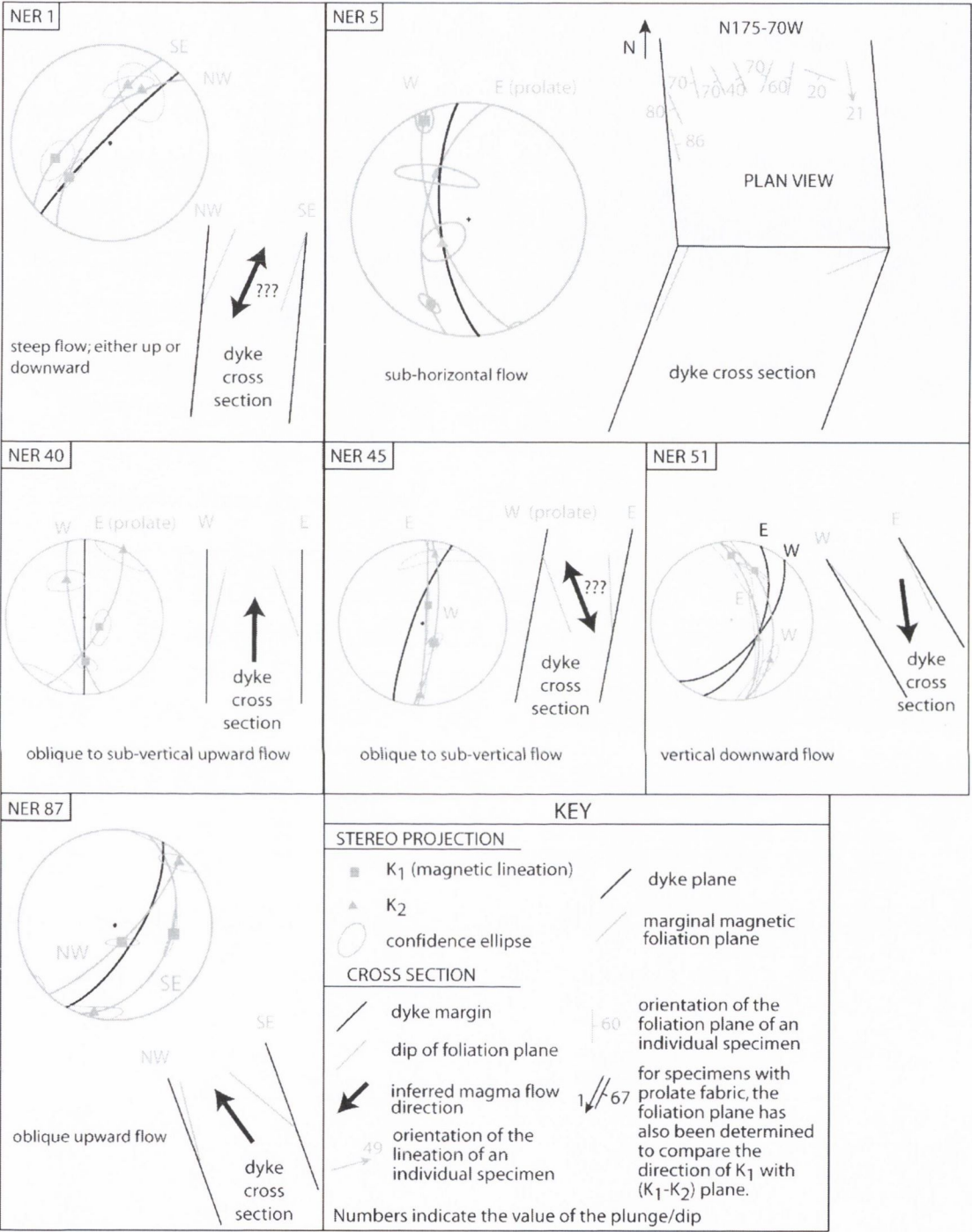


Figure 10: Representation of the AMS fabric for each dyke that belongs to the class 1-B (magnetic foliations/lineations along each margin are oblique to the margin, either slightly or strongly) to the margin). The dykes of this class indicate mainly oblique to sub-vertical flow, either downwards or upwards.

3.3.2.1.3. Class 1-C (Fig.11)

NER11 is the only case in this class. The imbrication of K_1 - K_2 planes and the vertical lineations suggest a vertical flow. The directions of the two K_1 - K_2 planes are approximately parallel to the dyke direction, which is consistent with a sub-vertical flow.

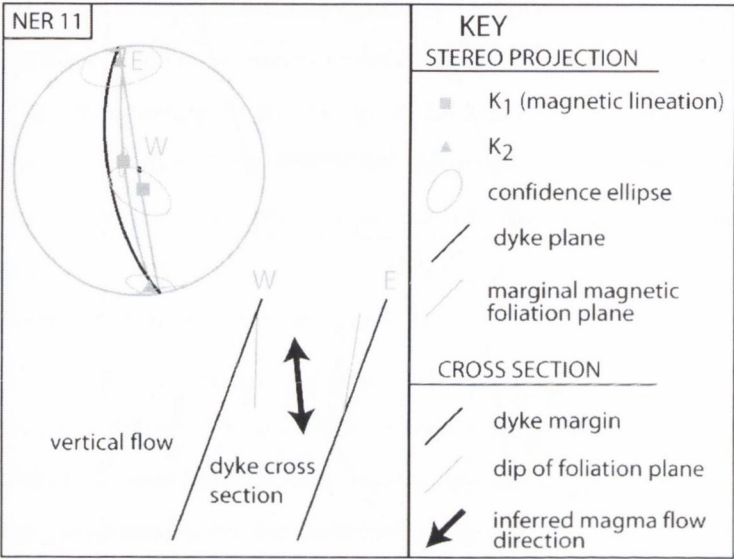


Figure 11: Representation of the AMS fabric for NER11 (magnetic foliation / lineation along each margin are parallel to the margin). The marginal magnetic foliation planes and the lineations indicate a vertical flow.

3.3.2.2. Class 2 (Fig.12)

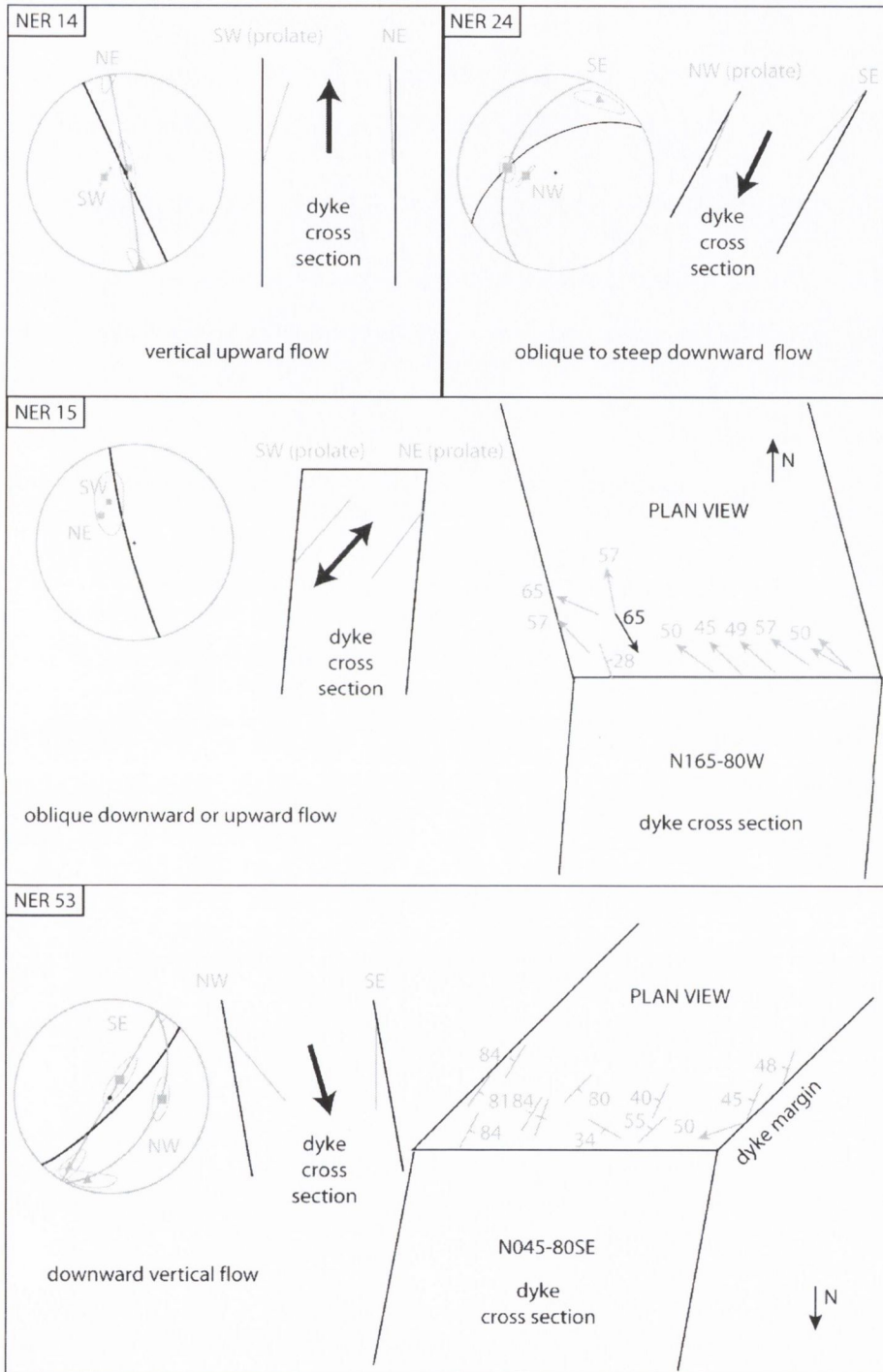


Figure 12: Representation of the AMS fabric for each dyke that belongs to class 2. See **figures 9 or 10** for key. Class 2-A: One margin has an oblate fabric and the other has a prolate one, strongly oblique to the dyke (NER14, 24). The AMS fabrics along the margins of NER14

suggest a vertical upward flow while they suggest an oblique to steep downward flow for NER24 (flow direction inconclusive). Class 2-B: Both margins present a prolate or an oblate fabric (NER15 & 53). The AMS fabrics along the margins of NER15 seem to indicate an oblique flow (flow direction inconclusive) and they suggest a downward vertical flow for NER53.

3.3.2.2.1. Class 2-A (Fig.12; dyke NER 14, 24)

The south-west margin of NER14 has a prolate fabric with reasonably well-defined lineation of high confidence. The magnetic foliation plane along the north-east margin and the magnetic lineation along the south-west margin suggest a vertical upward flow.

The magnetic lineation and the magnetic foliation of NER24 are both inclined westward. This pattern corresponds likely to an oblique to steep downward flow. However, as the NW margin has a prolate fabric, it is not possible to deduce the direction of the magma flow.

3.3.2.2.2. Class 2-B (Fig.12; dyke NER 15 and 53)

The fabric of NER15 is strongly prolate, attested to by the small confidence ellipse, and reflected in the margin fabrics that are also both prolate (Fig.8). The trends and the plunges of K_1 along each margin suggest oblique flow, consistent with the orientations of each individual sample lineation across the dyke. However, it is not possible to infer the direction of the flow as both margins are prolate.

The fabric of NER53 is overall prolate but almost triaxial, with large confidence ellipses (Fig.8). The margin are fortunately better defined and are both oblate. The dips and the imbrications of the magnetic foliation planes, as well as the lineations, suggest a downward vertical flow. The magnetic fabrics of each sample across the dyke are in general scattered but they still suggest downward flow.

3.3.2.3. Cross-section (Fig 13).

For five dykes, we attempted to infer the magma flow direction by drawing cross-sections showing the fabrics for individual specimens across the dyke-width. The results are at times difficult to interpret as the fabrics seem randomly distributed across the dykes, especially for the dykes NER29 and NER35. However, the orientations of the individual magnetic fabrics

across NER8 indicate a sub-vertical flow. The lineations at the centre of NER26 are all tilted towards the west or south-west, suggesting an oblique flow. The magnetic lineations at the centre of NER28 indicate an oblique flow. Nevertheless, all the flow directions and senses determine using the cross-sections have to be handled with care as the patterns are not obvious.

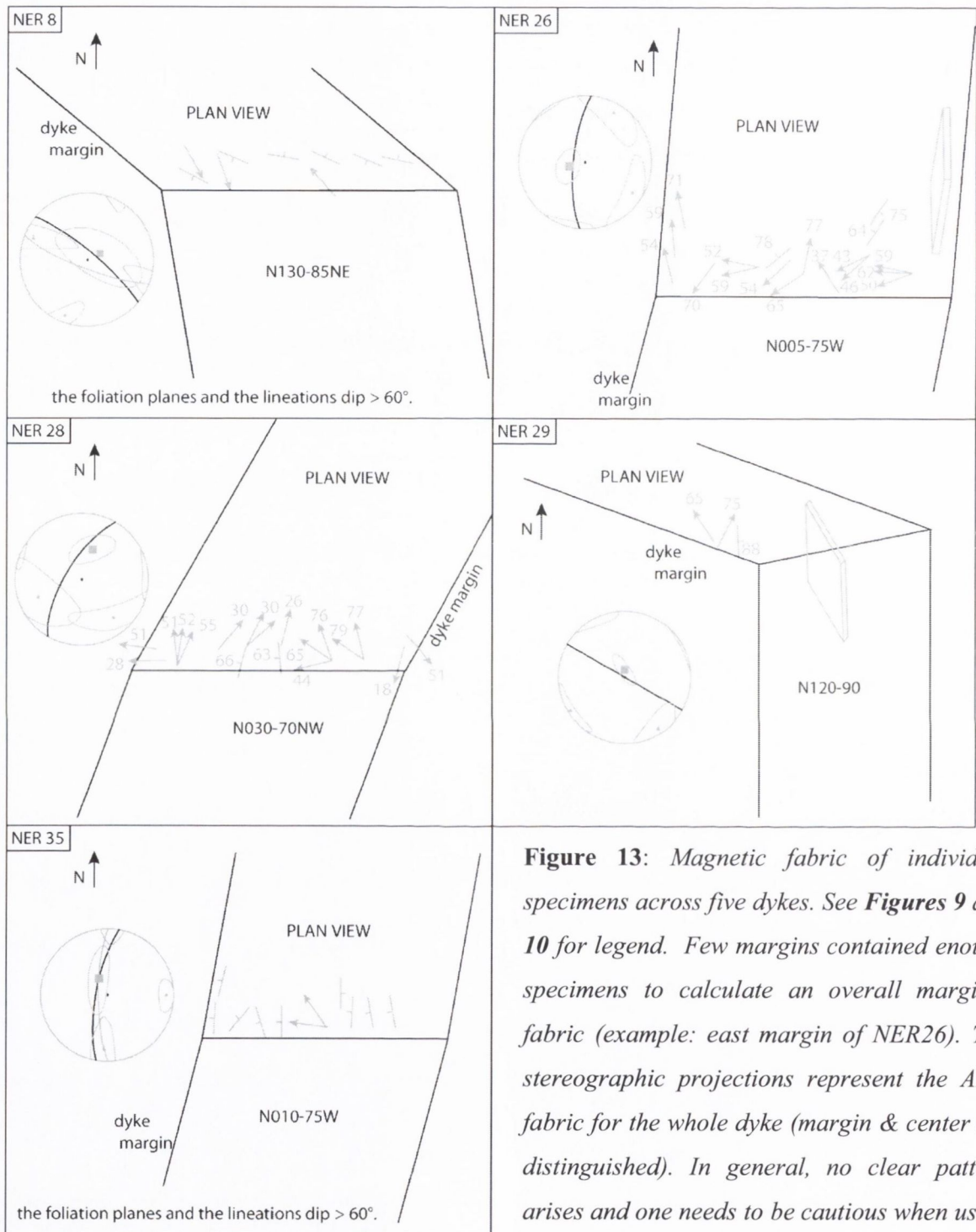


Figure 13: Magnetic fabric of individual specimens across five dykes. See **Figures 9 and 10** for legend. Few margins contained enough specimens to calculate an overall marginal fabric (example: east margin of NER26). The stereographic projections represent the AMS fabric for the whole dyke (margin & center not distinguished). In general, no clear pattern arises and one needs to be cautious when using the directions and senses of the flow inferred from these cross-sections.

4. Magnetic analysis: interpretation and discussion

4.1. AMS fabric

Single-domain or pseudo-single-domain titanomagnetite grains are characterized by a low degree of anisotropy compared with the larger multi-domain grains. Thus, the weak to moderate degree of anisotropy (**Fig.6** and **table 1**) is likely due to the presence of fine-grained titanomagnetite (*Hargraves et al., 1991*). The shape factor (**Fig.6A**) and the magnetic susceptibility (**Fig.6B**) seem to be only slightly related to the petrography. We especially note that the ankaramite and px-rich groups have higher magnetic susceptibilities, which is coherent with the high-magnetite content observed in the thin-sections of the dykes that belong to these petrographic groups.

In dykes of basaltic composition, the occurrence of inverse and intermediate fabrics is common (*Knight and Walker, 1988; Tauxe et al., 1998; Rochette et al., 1999; Geoffroy et al., 2002; DeFrates et al., 2006*). Several studies address the origin of such fabrics. This behaviour can be explained by the presence of single-domain titanomagnetite grains where K_1 and K_3 correspond respectively to the short and long axis of the grain (*Potter and Stephenson, 1988*), the anisotropic distribution of ferromagnetic particles (*Hargraves et al., 1991*), the anomalous orientation of magnetic grains with respect to dyke margins due to strain variations in the magma (*Dragoni et al., 1997*) and the dyke composition (*Aubourg et al., 2002; DeFrates et al., 2006; Chadima et al., 2009*). They can also occur when a primary fabric is overprinted by growth of a secondary fabric due to flow and post-flow processes or tectonic stresses (*Park et al., 1988; Rochette et al., 1991; Hrouda et al., 2002*). It has also been proposed that intermediate fabrics can result from a combination between normal and inverse fabrics (*Rochette et al., 1992; Rochette et al., 1999; Ferré, 2002*). Recently, *Chadima et al. (2009)* showed that such configuration can also depend on the individual grain orientation and the domain state. The dykes analyzed for this study are mostly composed by SD and PSD domains and 18% exhibit an inverse fabric, and 26% an intermediate fabric. The occurrence of such fabrics with SD domain grains can probably be explained by the combination of the possibilities discussed above. The origin of the reverse and intermediate fabrics is, however, not the focus of the present study.

4.2. AMS: flow direction

The upper part of the rift has been clockwise rotated by 26° along a vertical axis (**chapter 2**). The inferred directions of flows are thus similarly rotated. However, only three “geographic flow directions” have been conclusively deduced from the selected dykes (NER25, 59, 70). So even if this rotation is taken into account, the global magma flow pattern within the rift-zone will not change as no preferential trend of magma flow has been deduced.

The flow directions and senses are, admittedly, difficult to assess from the magnetic fabrics in the present study. We find some recurrent patterns (**Fig.7**), but for several dykes, the fabric arrangements do not allow the directions and/or the senses of magma flow to be inferred. Some dykes were sampled just before a change of dip (NER8 and 15), before going en-échelon (NER14, 24, 25), or at their tips (NER29, 59, 81). Those specific locations can be a reason why it is difficult to determine the flow direction in these places, as it might be highly variable on a very small scale. However, some of the dykes were drilled in more “regular” locations, and it is interesting to see that the AMS fabrics are ambiguous and hard to resolve in those cases too. This can be due to complicated internal flow, involving, for example, differential shearing rates and multiple flow generations as well as non-ideal rock magnetic properties.

Six dykes show a downward flow, steep or inclined (NER24, 25, 51, 53, 59 and 76B). Such flow can be simply due to cessation of magma supply or cooling that can cause the magma to shrink and the dyke contents to sink (magma deflation), resulting in an apparent downward flow (e.g., *Aubourg et al., 2002*). Downward flows have been also previously observed in dykes and have been interpreted as primary flow (*Knight and Walker, 1988; Varga et al., 1998; Poland et al., 2004; Soriano et al., 2008*). As downward flows have been observed more than once and across the NERZ, it may reflect a general process. Furthermore, we would expect steep downward flows in case of magma deflation and yet most of the downward flows are oblique. We can thus interpret these downward flows as primary. Such flows imply that magma propagates toward lower topographic levels and constitutes a potential hazard with eruption along the rift flanks (**Fig.14**). It is interesting to note that such flows have also been recently observed in the rift-zones of Teno and Anaga massifs (**Fig.3A**; *Soriano et al., 2008*). It thus seems a general process in the shallow levels of the Tenerife rift-zones. *Soriano et al. (2008)* interpreted downward flow due to magma emplacement above neutral buoyancy levels

and conclude that buoyancy is the magma driving force in the shallow level of the rift. This conclusion may well hold true in the NERZ too.

The main volcanic center, the Teide-Pico Viejo complex, is located to the SW of the NE rift-zone (**Fig.3A**). One hypothesis would be that the main complex acts as a focal point, and is supplied by a major magmatic reservoir below. Once a critical point is reached, the complex induces compressive stress that blocks vertical magma ascent and favours lateral magma propagation (e.g., *Pinel and Jaupart, 2004; Pueyo Anchuela et al., 2006; cf. chapter 4*). If the magmatic reservoir below the Teide Pico-Viejo complex would feed the NERZ exclusively, we should expect to see a consistent north-eastward direction of sub-horizontal magma flow with the AMS fabrics. However, we frequently observed vertical or oblique flow without a clear directional pattern. It can be thus inferred that the dykes that fed the NERZ are supplied by reservoir(s) that store magma just below the rift zone crest (**Fig.14**). Furthermore three dykes record a vertical upward flow (NER14, 40 and 87), suggesting a direct supply from a reservoir beneath. This suggestion is consistent with the idea developed in the first and second chapters that implies an intrusive complex below the rift. Note that it is, however, also possible to imagine a deeper feeding system coming from Teide-Pico Viejo complex with magma flowing laterally in the deeper part of the NERZ to then rise up to the ridge. It could be thus interesting to sample the lower part of the ridge to check this possibility.

The feeding of the upper part of the NE rift-zone is thus considerably more complicated in term of magma supply sources and also in terms of dyke propagation and magma flow within the dykes, thus negating simple, non-dynamic rift models as recently proposed by *Acocella and Neri (2009)*. Furthermore, as highlighted before, we only have access to the final stages of magma flow with AMS studies (*Hutton, 1992; Vigneresse and Clemens, 2000*). Consequently, the flows recorded in this work reflect the final stages of magma emplacement, which can be very different from the early phase, as it has been observed in real-time on Stromboli in 2002 (*Acocella et al., 2006*). During this eruption, magma emplacement was complex with dyke propagating in several directions. As suggested earlier, the role of deeper intrusive complex(es) and deep magma propagation needs to be considered to unravel the evolution of the NERZ.

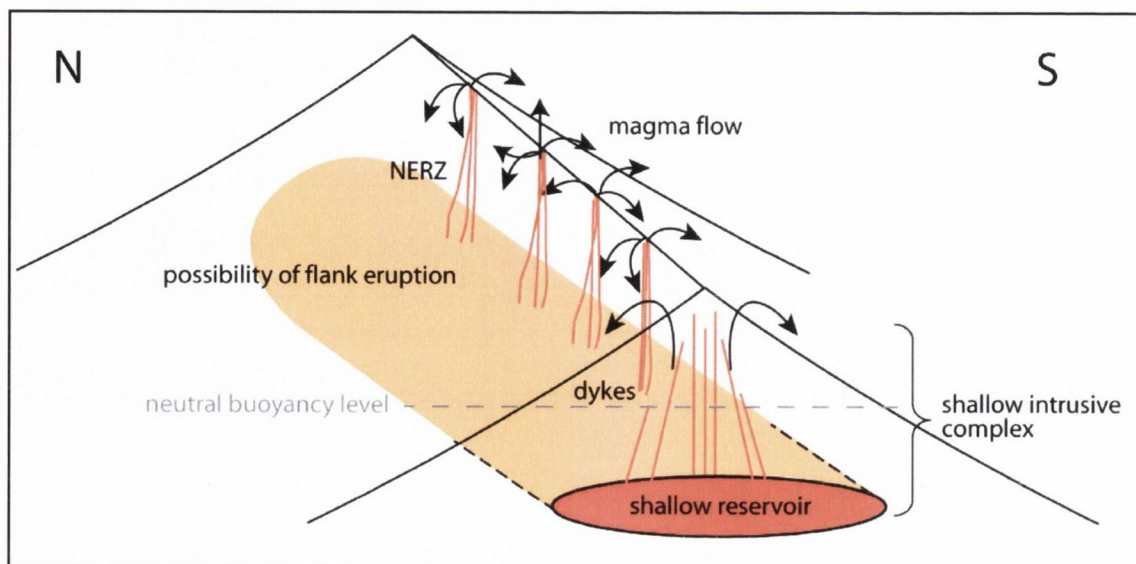


Figure 14: Model of magma propagation across the North-East Rift Zone. The dykes are fed by local shallow reservoir(s) within the rift and the magma propagates multidirectionally, usually from above the neutral buoyancy level. Magma in these dykes will thus flow towards lower topographic levels. This implies a potential hazard for future eruption that may likely occur along the rift flank.

5. Conclusion

From among the 79 dykes drilled along the NERZ, few are suitable to infer reliable magma flow information. This is either because of limited samples or because of the high dispersion of the data and/or the occurrence of inverse and intermediate fabrics. The arrangements of the magnetic fabrics of the dykes that are suitable are, in turn, not always easy to resolve, and ambiguities remain concerning the direction and or the sense of the magma flow, reflecting a complicated mode of dyke emplacement. A lot of dykes show oblique or vertical flow, mostly downward. Such flows imply magma flowing towards lower topographic levels. Future dyke emplacement can thus possibly lead to flank eruption rather than summital eruption. The direction and sense of magma flow within dykes do not follow a consistent pattern across the NERZ. It is thus more likely that the dykes are fed by a reservoir (or several reservoirs) located below the rift-zone instead of a uniform lateral supply from the island's centre. This reservoir(s) would form, together with the dykes, the intrusive rift-complex.

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TABLES:

Table 1: AMS data from the sampled area. The sites are labelled NER1, NER2 (sometimes also with a capital letter as in NER18A) etc. For each site, No is the number of samples analyzed per site, N is the number of data retained to calculate the site mean. Km is the magnetic susceptibility (SI units), K₁ and K₃ (in situ) are the azimuth and plunge respectively of the magnetic lineation and the normal to the magnetic foliation plane; K₁ and K₃ (corrected) are the azimuth and plunge of respectively the magnetic lineation and the normal to the magnetic foliation plane corrected according to the 26° vertical axis clockwise rotation (cf **chapter 2**). L: lineation; F: foliation; P: degree of anisotropy ; P': corrected degree of anisotropy (*Jelinek, 1981*); T: shape parameter; S: shape (O: oblate; P: prolate; T: triaxial). Fabric: N: normal, INTER: intermediate, I: inverse.

Site	Rock type	No	N	Km	K1 (in situ)	K3 (in situ)	K1 (corrected)	K3 (corrected)	L	F	P	P'	T	Sp	GPS coordinates		dyke orientation	fabric
															Latitude	Longitude		
NER1	aph	25	26	0.048	249\58	120\21	223\58	94\21	1.011	1.06	1.071	1.077	0.691	O	28.346	343.5185	044-85NW	N
NER3	ol+px	14	14	0.022	348\64	210\20	322\64	184\20	1.009	1.015	1.023	1.024	0.267	O	28.346	343.519	110-90	N
NER4	cpx=fd	7	12	0.010	258\82	36\6	232\82	10\6	1.015	1.001	1.016	1.018	-0.925	P	28.346	343.5188	100-90	scatter
NER5	fd>cpx	21	21	0.036	177\9	81\32	151\9	55\32	1.009	1.019	1.029	1.029	0.340	O	28.347	343.5192	175-70W	N
NER6	fd>cpx	11	13	0.004	301\42	192\20	275\42	166\20	1.030	1.008	1.039	1.041	-0.575	P	28.347	343.5193	010-70W	INTER
NER7	anka	9	18	0.037	262\60	104\28	236\60	78\28	1.025	1.010	1.036	1.037	-0.384	P	28.348	343.519	130-80NE	INTER
NER8	aph	12	13	0.013	74\72	201\11	48\72	175\11	1.002	1.014	1.016	1.017	0.747	O	28.348	343.5192	130-85NE	N
NER9	anka	25	26	0.033	158\31	60\13	132\31	34\13	1.009	1.029	1.038	1.040	0.506	O	28.348	343.5192	155-90	N
NER10	fd=cpx	13	13	0.006	277\67	92\23	251\67	66\23	1.043	1.007	1.050	1.054	-0.727	P	28.348	343.5193	170-50W	N
NER11	px-rich	12	12	0.045	283\83	82\7	257\83	56\7	1.021	1.036	1.058	1.059	0.258	O	28.348	343.5193	170-70W	N
NER12	fd>cpx	23	23	0.026	62\28	313\31	36\28	287\31	1.001	1.004	1.005	1.005	0.476	O	28.348	343.5194	080-70N	INTER
NER13	fd>cpx	27	28	0.008	261\12	157\50	235\12	131\50	1.010	1.002	1.011	1.012	-0.711	P	28.348	343.5202	060-65NW	N/INTER
NER14	anka	20	21	0.067	276\75	78\14	250\75	52\14	1.047	1.026	1.073	1.074	-0.283	P			155-90	N
NER15	fd=cpx	15	15	0.037	314\51	77\24	288\51	51\24	1.081	1.020	1.102	1.108	-0.597	P	28.349	343.5204	165-85W	N
NER16	aph	14	14	0.010	128\27	28\19	102\27	2\19	1.003	1.010	1.013	1.014	0.580	O	28.349	343.5204	130-90	N
NER17	fd>cpx	21	21	0.020	339\50	233\13	313\50	207\13	1.015	1.009	1.025	1.025	-0.244	P	28.349	343.5205	135-70NE	N
NER18	anka	33	35	0.069	232\50	43\40	206\50	17\40	1.012	1.010	1.022	1.022	-0.062	P	28.349	343.5204	000-90	INTER
NER18A	fd=cpx	17	19	0.037	280\77	21\3	254\77	355\3	1.014	1.004	1.018	1.019	-0.512	P	28.349	343.5204	060-70NW	INTER

NER19	fd>cpx	18	18	0.007	101\50	355\13	75\50	329\13	1.011	1.012	1.024	1.024	0.029	S/O	28.346	343.5212	030-75NW	INTER
NER20	aph	18	18	0.023	257\61	351\2	231\61	325\2	1.014	1.007	1.020	1.021	-0.347	P	28.346	343.5212	160-90	I
NER21	aph	21	22	0.024	284\64	78\23	258\64	52\23	1.018	1.010	1.027	1.028	-0.290	P	28.346	343.5212	140-90	N/INTER
NER22	fd=cpx	17	17	0.005974	81\59	172\1	55\59	146\1	1.003	1.009	1.012	1.012	0.536	O			175-80W	I
NER23	fd=cpx	13	14	0.0037	271\65	48\19	245\65	22\19	1.002	1.001	1.004	1.004	-0.282	P			090-50N	I/INTER
NER24	fd>cpx;ol+px	20	21	0.003	291\59	135\28	265\59	109\28	1.010	1.007	1.017	1.017	-0.221	P			060-60N	N
NER25	ol+px	25	26	0.055	349\33	95\24	323\33	69\24	1.012	1.015	1.027	1.027	0.102	O			025-70NW	N
NER26	fd=cpx	26	29	0.007	259\68	123\16	233\68	97\16	1.016	1.007	1.023	1.024	-0.429	P			005-75W	N
NER27	aph	17	17	0.010	308\74	218\0	282\74	192\0	1.013	1.010	1.023	1.023	-0.143	P			175-85	INTER
NER28	fd=cpx	20	21	0.039	18\49	141\26	352\49	115\26	1.029	1.007	1.036	1.038	-0.617	P			030-70N	N
NER29	aph	19	20	0.014	352\80	238\4	326\80	212\4	1.019	1.014	1.033	1.033	-0.130	P			120-90	N
NER33	anka	35	35	0.059	284\45	15\1	258\45	349\1	1.002	1.016	1.018	1.020	0.756	O	28.352	343.5247	025-90	scatter
NER35	anka	18	19	0.053	329\67	90\12	303\67	64\12	1.007	1.034	1.041	1.044	0.658	O	28.353	343.525	010-75W	N
NER36NAN	fd=cpx	11	11	0.035	24\3	284\74	358\3	258\74	1.001	1.004	1.005	1.005	0.518	O			030-90	scatter
NER36ANK	anka	15	15	0.026	327\72	87\9	301\72	61\9	1.003	1.006	1.009	1.009	0.389	O			015-90	N
NER36A	fd=cpx	15	15	0.005	29\74	123\1	3\74	97\1	1.014	1.006	1.020	1.020	-0.418	P			010-90	INTER
NER37	anka	25	26	0.114	295\0	25\12	269\0	359\12	1.015	1.019	1.034	1.034	0.114	S/O	28.353	343.5258	005-90	I
NER37A	anka	6	6	0.033	323\6	53\0	297\6	27\0	1.009	1.013	1.022	1.022	0.615	O	28.353	343.5258	005-90	INTER
NER38	ol+px	16	16	0.007	346\39	81\7	320\39	55\7	1.007	1.004	1.011	1.011	-0.245	P			080-70N	I
NER39	ol+px	22	24	0.009	280\60	18\5	254\60	352\5	1.001	1.005	1.006	1.007	0.695	O			070-80N	I
NER40	px-rich	19	20	0.098	130\72	268\14	104\72	242\14	1.004	1.017	1.021	1.022	0.622	O	28.353	343.5264	000-90	N
NER41	anka	15	16	0.069	191\34	282\1	165\34	256\1	1.004	1.022	1.026	1.028	0.679	O	28.353	343.5264	025-80W	N
NER42	aph	12	13	0.003	354\16	98\39	328\16	72\39	1.002	1.011	1.014	1.015	0.669	O	28.353	343.5264	030-80W	N
NER43	fd=cpx	20	20	0.033	19\51	151\28	353\51	125\28	1.042	1.016	1.059	1.061	-0.459	P			120-50NE	I
NER44	fd=cpx	16	16	0.005	23\28	283\18	357\28	257\18	1.015	1.009	1.024	1.024	-0.222	S/P	28.353	343.5264	155-75NE	INTER
NER44B	fd=cpx	17	18	0.005	24\48	253\31	358\48	227\31	1.033	1.015	1.048	1.049	-0.379	P			170-70E	N
NER45	anka	24	24	0.067	160\82	282\4	134\82	256\4	1.012	1.03	1.043	1.044	0.424	O	28.353	343.5268	020-80W	N
NER47	anka pxrich	35	35	0.027	332\85	112\4	306\85	86\4	1.046	1.031	1.079	1.08	-0.194	S/P			020-90	N
NER48B	aph	20	20	0.009	191\69	101\0	165\69	75\0	1.009	1.011	1.02	1.02	0.091	S	28.353	343.528	020-90	N
NER50	aph	27	29	0.029	58\74	241\16	32\74	215\16	1.012	1.009	1.021	1.021	-0.157	P			030-85SE	I
NER51	px-rich	21	24	0.030	43\45	274\33	17\45	248\33	1.012	1.03	1.043	1.044	0.418	O			030-60SE	N
NER52	aph	30	30	0.056	47\63	310\4	21\63	284\4	1.004	1.017	1.021	1.023	0.596	O			030-90	N
NER53	aph	19	19	0.035	58\48	298\24	32\48	272\24	1.011	1.01	1.021	1.021	-0.048	P	28.353	343.5289	045-80SE	N
NER54	ol+px	8	8	0.078	54\39	271\44	28\39	245\44	1.03	1.017	1.048	1.049	-0.270	P	28.353	343.5284	170-50E	INTER

NER55	fd=cpx	16	17	0.065	22\52	259\23	356\52	233\23	1.017	1.005	1.022	1.023	-0.564	P	28.353	343.529	005-80E	INTER
NER56	fd>cpx	14	18	0.024	127\62	346\23	101\62	320\23	1.016	1.003	1.02	1.021	-0.661	P	28.354	343.5292	010-70E	INTER
NER57	aph	32	35	0.055	56\65	313\6	30\65	287\6	1.014	1.015	1.029	1.029	0.046	S/O	28.354	343.5294	022-70SE	N
NER59	aph	17	19	0.010	233\70	124\7	207\70	98\7	1.004	1.019	1.022	1.024	0.675	O	28.354	343.53	032-90	N
NER60	aph	18	18	0.011	59\28	153\7	33\28	127\7	1.011	1.013	1.024	1.024	0.088	S	28.355	343.53	034-90	INTER
NER61	aph	9	9	0.003	10\26	103\7	344\26	77\7	1.02	1.012	1.032	1.032	-0.231	P			010-90	N
NER65	fd>cpx	34	34	0.032	133\8	5\76	107\8	339\76	1.021	1.013	1.034	1.034	-0.212	P	28.362	343.53	140-wiggly	I
NER66	fd=cpx	23	23	0.004	2\24	115\41	336\24	89\41	1.058	1.025	1.085	1.087	-0.391	P			140-90	INTER
NER70	px-rich	15	15	0.044	263\19	358\14	237\19	332\14	1.017	1.047	1.064	1.066	0.47	O			100-90	N
NER70A	aph.px	7	7	0.011	91\69	217\13	65\69	191\13	1.018	1.015	1.033	1.033	-0.098	P			114-90	N
NER71	aph	15	16	0.037	259\62	139\15	233\62	113\15	1.004	1.01	1.014	1.015	0.441	O			150-80NW	I
NER72	px	16	17	0.202	127\61	316\29	101\61	290\29	1.018	1.014	1.033	1.033	-0.102	P			120-90	I
NER73	aph	10	14	0.028	226\68	12\18	200\68	346\18	1.005	1.03	1.035	1.038	0.711	O			075-80SE	INTER
NER76B	ol+px	12	12	0.021	33\74	130\2	7\74	104\2	1.015	1.031	1.046	1.047	0.326	O			052-90	N/INTER
NER77B	ol+px	24	27	0.036	199\77	40\12	173\77	14\12	1.009	1.015	1.025	1.025	0.239	O	28.316	343.50	135-80NE	N/INTER
NER79	aph	11	11	0.011	88\72	206\8	62\72	180\8	1.01	1.013	1.024	1.024	0.145	O			145-80NE	N
NER81	ol+px	21	21	0.006	46\24	313\6	20\24	287\6	1.015	1.022	1.037	1.037	0.2	O	28.309	343.50	035-70SE	N
NER82	aph	26	27	0.029	183\43	39\40	157\43	13\40	1.008	1.036	1.045	1.048	0.621	O	28.309	343.4957	120-65S	N
NER86	fd>cpx	15	15	0.004	83\46	301\37	57\46	275\37	1.005	1.022	1.027	1.029	0.637	O	28.373	343.5387	090-90	I
NER87	aph	24	24	0.035	101\68	307\20	75\68	281\20	1.013	1.038	1.052	1.054	0.504	O	28.377	343.5404	030-70SE	N
G1D3	fd>cpx	14	14	0.037	163\68	282\11	137\68	256\11	1.02	1.017	1.038	1.038	-0.079	P			190-60E	N
G1D4	aph	18	18	0.078	95\75	186\0	69\75	160\0	1.039	1.02	1.06	1.061	-0.315	P			190-90	I
G1D5	ol+px	12	12	0.036	13\4	207\85	347\4	181\85	1.01	1.02	1.03	1.031	0.354	O			035-90	I
G1D6	aph	14	14	0.025	184\51	88\5	158\51	62\5	1.015	1.023	1.038	1.038	0.202	O			000-80E	N
G2D1	aph?alt II	13	13	0.091	271\82	129\6	245\82	103\6	1.016	1.004	1.02	1.021	-0.557	P			055-70SE	N
G2D4	aph	15	15	0.045	79\50	343\5	53\50	317\5	1.027	1.025	1.053	1.053	-0.041	P			050-65SE	I
G2D7	aph	11	11	0.038	62\71	163\4	36\71	137\4	1.006	1.04	1.046	1.05	0.720	O			050-90	scatter

Table 2: AMS fabrics for the dykes studied in detail, with the inferred magma flow. K_1 and K_3 (in situ) are the azimuth and plunge respectively of the magnetic lineation and the normal to the magnetic foliation plane K_1 - K_2 .

CASE 1		Shape factor (T)	K1	K3	(K1-K2)	Flow direction
NER 25	dyke	oblate	349\33	95\24	N005-66W	Oblique downward (northward)
	NW margin	oblate	343\46	78\4	N168-86W	
	SE margin	oblate	350\40	107\28	N017-62W	
NER 59	dyke	oblate	233\70	124\7	N034-83W	Oblique downward (southwestward)
	SE margin	oblate	254\66	95\22	N005-68W	
	NW margin	oblate	43\60	310\1	N040-89E	
NER 70	dyke	oblate	263\19	358\14	N088-76S	Sub-horizontal (westward)
	N margin	oblate	254\21	350\15	N080-75S	
	S margin	oblate	287\7	18\8	N108-82S	
NER 76B	dyke	oblate	33\74	130\2	N040-88W	Oblique downward
	SE margin	oblate	17\63	123\8	N033-82W	
	NW margin	oblate	100\80	322\8	N052-82S	
NER 81	dyke	oblate	46\24	313\6	N043-84S	Sub-horizontal
	SE margin	prolate	233\5	324\9	N054-81S	
	NW margin	oblate	74\68	307\14	N037-76E	
NER 1	dyke	oblate	249\58	120\21	N030-69W	steep
	NW margin	oblate	253\43	140\23	N050-67W	
	SE margin	oblate	231\45	122\18	N032-72W	
NER 5	dyke	oblate	177\9	81\32	N171-58W	Sub-horizontal
	E margin	prolate	205\22	101\31	N011-59W	
	W margin	oblate	336\7	69\23	N159-67W	
NER 40	dyke	oblate	130\72	268\14	N178-76E	Oblique to sub-vertical upward
	W margin	oblate	179\40	78\12	N168-78W	
	E margin	prolate	123\70	299\20	N029-70E	
NER 45	dyke	oblate	160\82	282\4	N012-86E	Oblique to sub-vertical
	W margin	prolate	150\68	276\13	N006-77E	
	E margin	oblate	17\77	273\3	N003-87E	
NER 51	dyke	oblate	43\45	274\33	N004-57E	Vertical downward
	E margin	oblate	0\20	260\25	N170-65E	
	W margin	oblate	23\32	260\41	N170-49E	
NER 87	dyke	oblate	101\68	307\20	N037-70S	Oblique upward flow
	SE margin	oblate	98\40	290\50	N020-40E	
	NW margin	oblate	162\74	313\14	N043-76E	
NER 11	dyke	oblate	283\83	82\7	N172-83W	Vertical
	W margin	oblate	176\77	80\1	N170-89W	
	E margin	oblate	305\78	83\9	N173-81W	
CASE 2		Shape factor (T)	K1	K3	(K1-K2)	Flow direction
NER 14	dyke	prolate	276\75	78\14	N168-76W	Vertical upward
	SW margin	prolate	257\72	61\17	N151-73W	
	NE margin	oblate	26\86	261\2	N171-88E	

NER 24	dyke	prolate	291\59	135\28	N045-62W	Oblique to steep downward
	NW margin	prolate	266\67	127\18	N037-72W	
	SE margin	oblate	274\49	112\40	N022-50W	
NER 15	dyke	prolate	314\51	77\24	N167-66W	Oblique
	SW margin	prolate	328\49	80\18	N170-72W	
	NE margin	prolate	308\53	86\29	N176-61W	
NER 53	dyke	prolate	58\48	298\24	N028-66E	Vertical downward
	SE margin	oblate	30\74	120\0	N030-90	
	NW margin	oblate	92\47	297\40	N027-50E	
CROSS SECTION		Shape factor (T)	K1	K3	(K1-K2)	Flow direction
NER 8	dyke	oblate	74\72	201\11	N111-79S	Sub-vertical
	margins					
NER 26	dyke	prolate	259\68	123\16	N033-74NW	Oblique
	W margin					
	E margin	oblate	253\81	89\9	N179-81W	
NER 28	dyke	prolate	18\49	141\26	N051-64NW	Oblique
	margins					
NER 29	dyke	prolate	352\80	238\4	N148-86NE	?
	centre	oblate	125\85	239\2	N149-88NE	
NER 35	dyke	oblate	329\67	90\12	N000-78W	?
	margins					

CHAPTER 4: ANALOGUE MODELLING

Abstract

Three different set ups have been tested to constrain the relationship between intrusive complex emplacement and growth, and the structural evolution of the ridge. The first one consists of a sand-plaster ridge lying on a basal ductile layer. This set up produces interesting results, but they do not correspond to the NERZ geometries. The second and the third set up consist of sand-plaster ridges intruded by golden syrup, which was used as magma analogue. Intrusive complex emplacement within a ridge seems to be mainly accommodated by flank creeping. This process allows the rift to stabilize until a critical point is reached where further intrusion will generate collapse. The analogue model results are in good agreement with the field observation and paleomagnetic results.

1. Introduction

1.1. Principle

The purpose of analogue modelling is to reproduce phenomena observed in nature at a laboratory scale. In volcanology, these phenomena include dynamic processes such as pyroclastic and lava flows, debris avalanches, ash plumes, vesiculation and fragmentation mechanisms, as well as volcano-tectonic processes such as fracture formation and propagation, landslides, spreading and intrusion emplacement. All these processes are governed by physical and mechanical laws that are not easily comprehended in nature due to scale, time and difficulties in making suitable observation. Reproducing these phenomena at a laboratory scale allows the parameters within a simplified system to be controlled. The key point of analogue modelling is to keep a coherent balance of geometry, forces and time between the natural case and the model reproduced in the laboratory, which is achieved by using scaling rules (*Hubbert, 1937; Ramberg, 1981; Middleton and Wilcock, 1994*).

1.2. State of the Art

1.2.1. Scaling and dimensionless analysis

Dimensionless analysis is an analytical method applied to solve physical problems using scaling laws that ensures the necessary conditions to maintain the same geometric, time

and force ratios between models and natural cases (*Hubbert, 1937; Ramberg, 1981; Middleton and Wilcock, 1994*).

The first step in scaling consists in listing the different parameters (**N**) describing the system with their corresponding dimensions (**n**) (*Middleton and Wilcock, 1994*).

The second step in scaling consists in defining the parameters that have independent dimensions, called repetitive variables (*Middleton and Wilcock, 1994*). These parameters correspond to variables that are of secondary importance for the analysis.

1.2.2. Dimensionless numbers

Once scaling is completed, the parameters are cast into dimensionless numbers to quantitatively describe the model (*Hubbert, 1937; Ramberg, 1981; Middleton and Wilcock, 1994*). There are usually a large number of parameters, but if they are held constant, they can be removed from the dimensionless analysis.

The Buckingham Π -theorem (e.g. *Middleton and Wilcock, 1994*) gives the quantity of dimensionless numbers that can be established using this simple equation: **N-n**. As these numbers are without dimension, they can be combined and compared in graphs to ascertain physical or mechanical processes that govern the phenomena in the laboratory and thus, by analogy, in nature.

1.2.3. Material

The scaling, previously completed, is used to find the materials that preserve the same geometric, time and force ratios between models and natural cases. Usually pure sand or a mix of plaster and sand is used to model brittle rock (*Merle and Borgia, 1996; Donnadieu, 2000*). Silicon gel is the preferred analogue material for ductile rocks (*Merle and Borgia, 1996; van Wyk de Vries and Francis, 1997; Oehler et al., 2005*). Golden Syrup or honey is frequently used as an analogue for magma (e.g. *Mathieu et al., 2008; Holohan et al., 2008*). The scaling and the properties of materials used will be developed in section 2.4.

2. Ridge models

2.1. Introduction

Volcanic rift-zones are vent alignments fed by dense swarms of dykes and sills that are associated with shallow intrusive complexes. The intrusive complexes are usually composed by a partially solid basal cumulate body resulting from crystal fractionation of the magma. Because of its density, the crystal-poor magma is concentrated at the top of the complex. The intrusive complex is a motor for the deformation as it induces long-term strain, associated with ductile and gravitational deformation during its growth and development, and is also the feeder source of dykes and sills, which are superficial intrusions associated with short-term deformation. Surface deformation due to intrusion emplacement and growth has been commonly observed. For example, a bulge due to cryptodome development was observed at the surface of Mount St Helens prior the 1980 eruption. Furthermore, inflation and displacements of volcano flanks related to dyke emplacement has been observed in several edifices such as Hawaii, Piton de la Fournaise and Etna, (Swanson *et al.*, 1976; Dieterich, 1988; Cayol *et al.*, 2000; Brooks *et al.*, 2008). Fiske and Jackson (1972) studied the mechanisms of dyke emplacement within a volcanic rift-zone using gelatine models applied to the Hawaiian case (**Fig.1A**). They concluded that the gravitational field is the main factor controlling rift geometry. Using gelatine models, Walter and Troll (2003) demonstrated that dyke orientation within the rift is influenced by collapse structure and applied this model to the Canary Islands (**Fig.1B**). Both studies, however, did not investigate the temporal evolution of extended magma emplacement and the associated surface deformation. Building on Fiske and Jackson (1972) and Walter and Troll (2003), who focused on dyke emplacement, this chapter studies the structures and the deformation field related to emplacement and growth of an intrusive complex within sand-plaster ridge models.

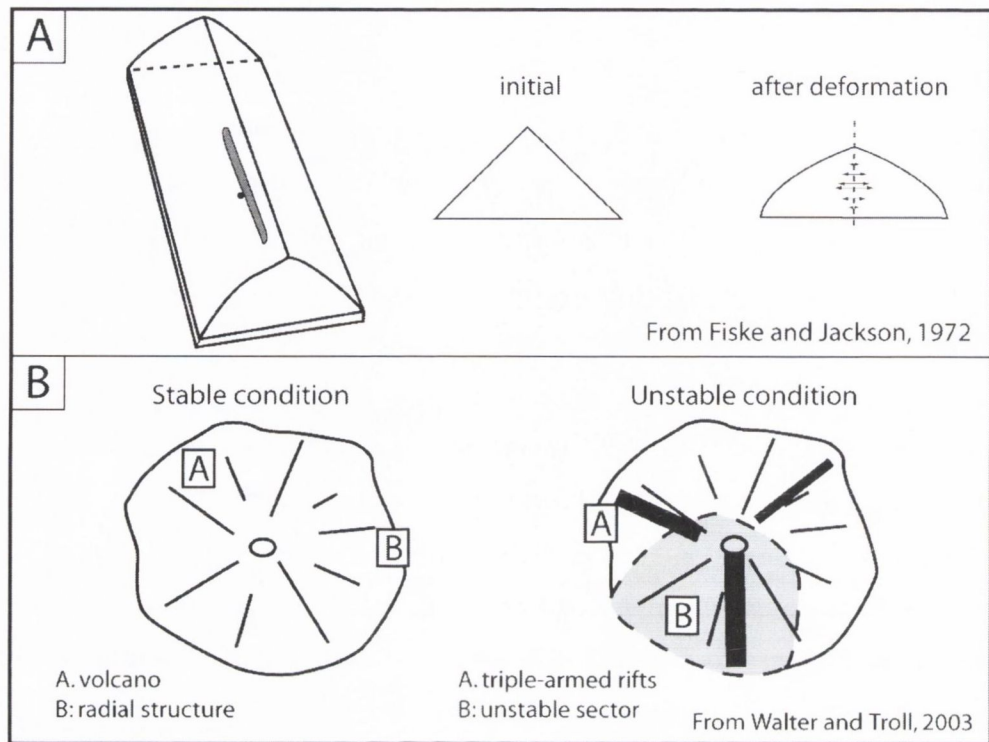


Figure 1: **A-** Fiske and Jackson gelatine rift-zone model (1972), where the gravitational field controls the rift geometry. **B-** Walter and Troll (2003) used gelatine conical models to demonstrate that rift zone orientation is modified by collapse structures.

Flank sector creeping, spreading and sagging were the main phenomena observed during the experiments. Flank sector creeping implies a progressive displacement of one sector of the ridge. Spreading and sagging are caused by the flow of a basal ductile substratum (for example low strength sediments or an intrusive complex) due to gravitational loading. Gräben and en-échelon fractures are the characteristic features of spreading and correspond to an extensional regime (**Fig.2A**; van Bemmelen, 1949; Merle and Borgia, 1996; Borgia et al., 2000; Borgia and van Wyk de Vries, 2003). Sagging is characterized by terrace formation, corresponding to a compressional regime (**Fig.2B**; Byrne et al., 2009).

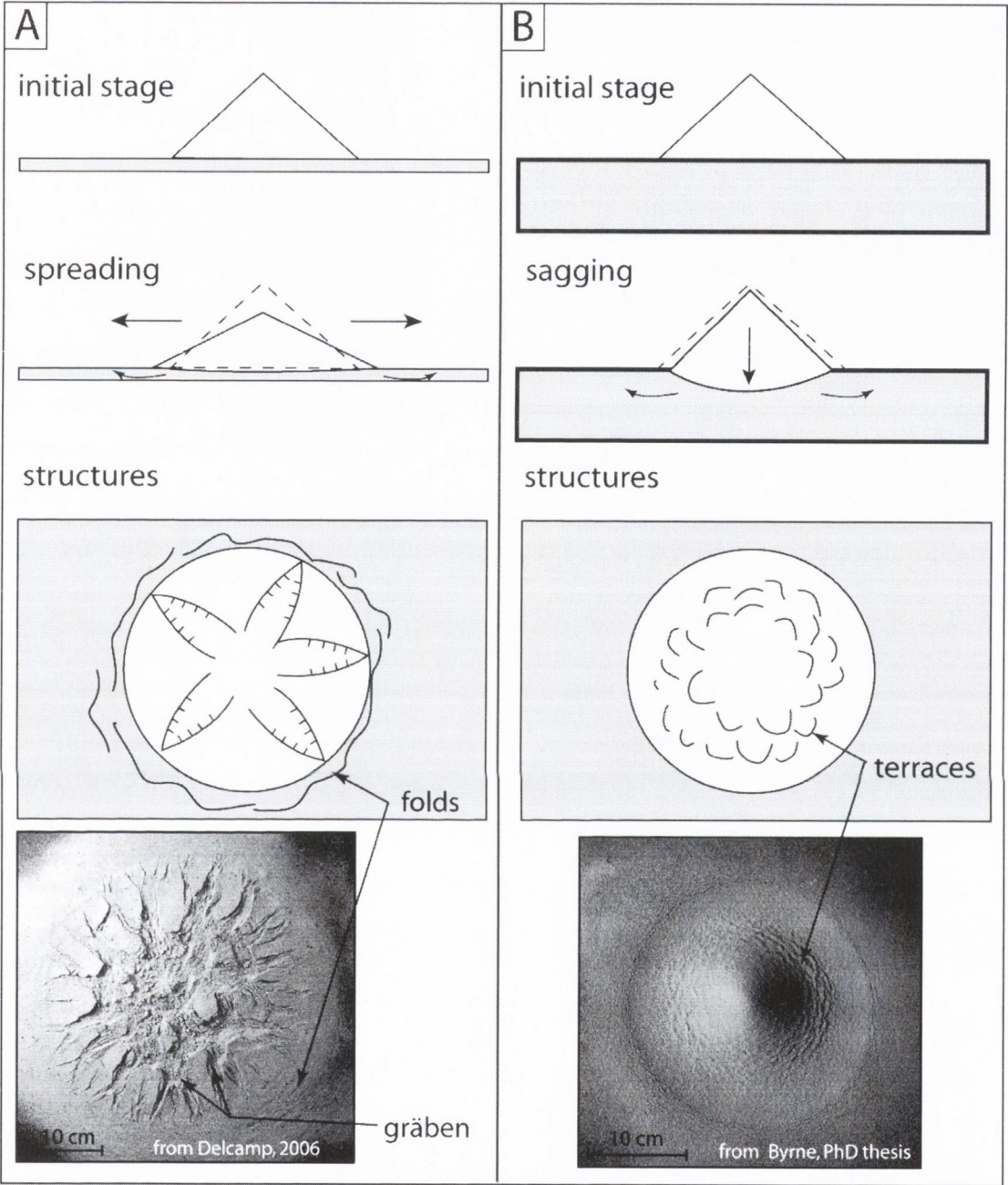


Figure 2: A- Illustration of spreading and associated structures. B- Illustration of sagging and associated structures.

2.2. Experimental set up

Different set-ups have been designed to simulate the morphology of a typical rift-zone, i.e. a ridge, and the nature of the constituent rocks (brittle and ductile). An ODP leg (Oceanic Drilling Program) has been carried out around Gran Canaria through the island's volcanic apron (*Weaver et al., 1998*). Two drill sites, close to Tenerife's east coast (sites 954 and 956) provide a rough picture of the sediment types and properties that probably constitute the submarine layers in the vicinity of Tenerife. Another site (397), approximately 100 km to the south of Gran Canaria, sampled 1.3 km of the 2-3 km of sediment cover (Mesozoic to Neogene). They recognized fine-grained compacted Cretaceous silt and claystones, radiolarian nannofossil mudstones, pebbly mudstones, nanofossil chalks and marls. The sites 954 and 956 are mainly composed of debris avalanche deposits. *Schmincke and Segschneider (1998)* noted minor thin layers of nanofossil ooze and clay in between the volcanoclastic rocks. From these observations it can be deduced that the surroundings and also the basement of Tenerife are partly composed up such deposits that may act as ductile layers, referred to herein as Low Strength Layers (LSLs) after *Oehler et al. (2005)*, that favour edifice destabilization. These sediment layers are partly replaced by intrusions that also help the edifice to spread (*Borgia, 1994*). The basal hyaloclastites formed during the early submarine stage of volcanic growth, and the highly crushed and deformed lava flows and the associated scoria observed in the field (**chapter 1**) may equally act as LSLs. Periods of instability are attested by debris avalanche deposits found in the volcano aprons around Gran Canaria. There, landslides occurred even during the early stages of the volcano growth (ODP, *Weaver et al., 1998*; *Schmincke and Segschneider, 1998*).

Three different set-ups were designed to assess in detail the surface deformation associated with the emplacement and growth of an intrusive complex within a rift-zone (**Fig.3**).

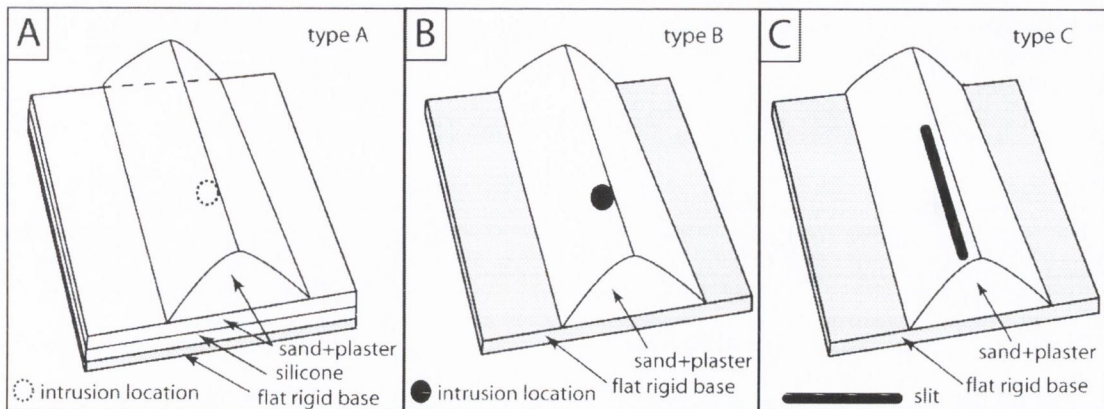


Figure 3: The different set ups designed for this study. **A-** Type A: sand and plaster ridges lying on a basal ductile layer, sometimes topped by a brittle layer. Golden Syrup was intruded in some experiments through a hole at the base of the model. **B-** Type B: sand and plaster ridges intruded by Golden Syrup from a circular hole. **C-** Type C: sand and plaster ridges intruded by Golden syrup from a slit.

The first models (type A) consist of sand (90%) and plaster (10%) ridges, lying on a basal ductile layer (silicone). A brittle layer (sand/plaster) covered the silicone in a few experiments (**Fig.3A**), but the brittle layer was not always added. Several experiments were performed with Golden Syrup intrusion. In total 25 experiments were performed.

The second set-up consists of ridges of plaster and sand mix (50/50) intruded below the centre or under the flank by Golden Syrup rising up from a circular hole (**Fig.3B**). In some experiments, the pipe was slightly above the base, allowing Golden Syrup to propagate directly into the ridge without spreading along the rigid base. 17 experiments were performed.

The last set up consists of ridges of plaster (50% or 40%) and sand (50% or 60%) intruded at the centre or along the flank by Golden syrup coming up through a slit (**Fig.3C**). 14 experiments were performed.

The basal silicone layer in the type A models represents the underlying ductile substratum. Although the presence of such layers is conceivable, it is not fully confirmed and other models without a basal ductile layer were also performed (types B and C). A Golden Syrup intrusion within a 90% sand, 10% plaster ridge only infiltrates and wets the model, the porosity of this mixed material being too high. This effect is overcome by

using a more cohesive and less porous material and thus by changing the proportion of plaster and sand to 50/50 and 60/40 (types B and C).

All the models were built on a flat and rigid base. The aperture from where the magma rises up within the models (either a circular hole or slit) is connected by a pipe to a reservoir containing Golden Syrup. The analogue magma rises through the model by simple overpressure. The experiments with the “intrusion” are then frozen to preserve the shape of the intrusive complex. Pictures were taken at regular steps during the experiments (every 1 or 5 minutes), enabling 2D-horizontal deformation to be tracked. Tracking was possible using a patch correlation technique written in MatlabTM, from which the x-y coordinates of the image are obtained from individual black markers (usually magnetite grains or pepper) tracked in successive images. Two velocities have been calculated from these data: “instantaneous” velocity (i.e. velocity between two consecutive pictures) and “total” velocity (average velocity between the first and the last image). The variation of horizontal displacement velocities through time can thus be followed and compared between the models. Deformation fields were then computed using SurferTM. This method, developed by *James* (in *Delcamp et al., 2008*), gives satisfying results. However, most of the models performed here are characterized by a non-uniform and localized deformation. In other words, within the same time interval, the values and direction of the horizontal displacement vectors are different within the model. The calculated velocities (total and instantaneous) correspond to an average of individual velocities determined from individual particle horizontal displacements. They provide thus an average of the horizontal displacement rate. It is more instructive therefore to look at the deformation fields as they show the variation of the displacements along the ridge within the same time interval, reflecting the asymmetrical distribution of the deformation.

2.3. Scaling and dimensionless numbers.

The necessary conditions to maintain the same geometric, time and force ratios between models and natural cases are guaranteed by scaling (*Hubbert 1937; Ramberg 1981; Middleton and Wilcock 1994*). The presented scaling is similar to a previous study where

a comparable set up was used (thus involving similar parameters) but with a different geometry (i.e. volcano models instead of ridge models; *Delcamp et al., in review*).

Length scale (l^*) is 4×10^{-5} , density scale (ρ^*) is 0.6 and gravity scale (g^*) is 1 (**Table 1**). The product of these provides the stress and cohesion scale (σ^*), that is thus about 10^{-6} :

$$l^* \cdot \rho^* \cdot g^* = \sigma^* \quad (\text{eq. 1})$$

Using this, the sand-plaster mix with a cohesion of about 10 Pa (varies between 7 and 11Pa) corresponds to a rock with a cohesion of about 10^5 Pa.

Time scale (t^*) is obtained from velocity data. The horizontal displacement velocity is preferentially utilised rather than the ‘time of deformation’, as this latter parameter is often hard to define in nature and in experiments (*Merle and Borgia, 1996*). Characteristic flank displacement rates in the models are about 10^{-7} ms^{-1} . Generally, the flank displacement rates on volcanoes are about $3 \times 10^{-10} \text{ m.s}^{-1}$ (equivalent to 1 cm per year). Consequently the velocity scale (u^*) is 3×10^3 . The time scale can be obtained also using the length and velocity scale:

$$l^* / u^* = t^* \quad (\text{eq. 2})$$

Thus, the time scale (t^*) is about 10^{-9} . The viscosity scale (μ^*) is extracted by using stress and time scales:

$$\mu^* = \sigma^* \cdot t^* \quad (\text{eq. 3})$$

The viscosity scale is thus determined at 10^{-14} , corresponding to a sediment layer viscosity of 10^{16} Pa.s and to an intrusive complex viscosity of 10^{13} Pa.s . The substrata values are coherent with those obtained in other studies (*Merle and Borgia, 1996; Cecchi, 2003*) and the intrusive values are consistent with the expected viscosity for a semi-solidified magma body (e.g. *Holohan et al., 2008*). Consequently, the deformation of the whole intrusive complex is considered as an entity. Indeed, additional low viscosity

intrusions that may create dykes from the intrusive body are not taken into account. The unknown natural values in the scaling are not expected to influence the style of deformation observed (**table 1**).

The 17 parameters that may play a role in the deformation are listed in the **Table 1**. Previous studies show that the viscosities of the substratum and the intrusion, and the intrusive flux play an important role in the displacement velocity but do not interfere in the style of deformation (*Merle and Borgia, 1996; Delcamp et al., in review*). Other parameters such as the cohesion and the densities do not influence the character of the deformation (*Merle and Borgia, 1996*). These parameters can thus be removed from the list, reducing the number of variables to 5 expressed with one dimension: the length (**Table 2**).

Table 1: List of the parameters describing the models; *: for the slit, two sizes have been used (see *appendix*).

Parameters	Definition	Values		Units	Ratio M/N
		Model	Nature		
h	ridge height	0.02	500	m	4×10^{-5}
l	ridge length	0.2	5000	m	4×10^{-5}
w_r	Ridge $\frac{1}{2}$ width	0.2	5000	m	4×10^{-5}
w_s	substratum $\frac{1}{2}$ width (brittle + ductile)	0.2	5000	m	4×10^{-5}
t_b	brittle substratum thickness	0.02	500	m	4×10^{-5}
t_d	ductile substratum thickness	0.02	500	m	4×10^{-5}
μ_s	substratum viscosity	10000	10^{16}	Pa.s	10^{-14}
μ_i	intrusion viscosity	10	10^{13}	Pa.s	10^{-14}
ρ_r	ridge density	1500	2500	Kg.m^{-3}	0.6
ρ_s	ductile substratum density	935	1558	Kg.m^{-3}	0.6
ρ_i	intrusion density	1625	2708	Kg.m^{-3}	0.6
u	displacement velocity	10^{-7}	3.18×10^{-10}	m.s^{-1}	3.15×10^3
τ_0	cohesion	10	10^5	Pa	10^{-6}
Φ	angle of internal friction	30	30		1
g	gravity acceleration	9.81	9.81	m.s^{-2}	1
f	intrusive flux	10^{-8}	10^{-2}	m.s^{-3}	10^{-7}
$\varnothing_i$	circular hole diameter*	0.02	500	m	4×10^{-5}

Table 2: Range of parameters in model and nature equivalents that play a major role in the style of deformation

Parameters	Definition	Values		Units	Ratio M/N
		Model	Nature		
h	ridge height	0.02	500	m	4×10^{-5}
w _r	Ridge ½ width	0.2	5000	m	4×10^{-5}
w _s	substratum ½ width (brittle + ductile)	0.2	5000	m	4×10^{-5}
t _b	brittle substratum thickness	0.02	500	m	4×10^{-5}
t _d	ductile substratum thickness	0.02	500	m	4×10^{-5}

Following the Π -Buckingham theorem (e.g. *Middleton and Wilcock, 1994*) 4 dimensionless numbers describing the system geometry can be established. Note that if the initial 17 parameters were all taken into account, 14 Π -numbers could have been established in total, i.e. 10 supplementary numbers could have been determined apart from the previous 4 dimensionless numbers. These 10 other Π -numbers ($\Pi 5$ to $\Pi 14$) are displayed in **table 3** for reference, but as previously mentioned they do not influence the deformation style and are thus of secondary relevance (*Merle and Borgia, 1996; Delcamp et al., 2008*).

$\Pi 1$ corresponds to the ratio between the ridge height and its ½ width.

$$\Pi 1 = \frac{\text{ridge height}}{\text{ridge } 1/2 \text{ width}} = \frac{h}{w_r} \qquad (eq. 4)$$

$\Pi 2$ is the ratio of the ridge height on the substrata thickness (brittle and ductile). This number equals zero for the set ups B and C as there is no basal substratum.

$$\Pi 2 = \frac{\text{ridge height} + \text{thickness of brittle substrata}}{\text{thickness of ductile substrata}} = \frac{h + t_b}{t_d} \qquad (eq. 5)$$

$\Pi 3$ corresponds to the ratio of the ridge width over the substratum width. This number equals zero for the set ups B and C.

$$\Pi 3 = \frac{2 \times \text{ridge } 1/2 \text{ width}}{\text{substratum width}} = \frac{2 \times w_r}{w_s} \quad (eq. 6)$$

$\Pi 4$ is the ratio between the thickness of the brittle substrata and the thickness of the ductile layer. For the set ups B and C, $\Pi 4$ is always equal to zero.

$$\Pi 4 = \frac{\text{thickness of brittle substrata}}{\text{thickness of ductile substrata}} = \frac{t_b}{t_d} \quad (eq. 7)$$

Obviously, the values may vary during the experiments according to the evolution of the deformation.

Table 3: Description of the 10 supplementary *II*-numbers that do not play a significant role in the style of deformation. They correspond to the numbers established previously by Merle and Borgia (1996).

	Formules	Comments
II5	$\frac{\text{ridge height}}{\text{ridge length}} = \frac{h}{l}$	This number describes the model geometry.
II6	$\frac{\text{ridge height}}{\text{circular hole diameter}} = \frac{h}{\varnothing_i}$	This two numbers describe the system geometry. The critical factors in the deformation are however the shape rather than the diameter of the pipe and its position related to the ridge crest. These two factors can not be expressed by dimensionless numbers.
II7	$\frac{\text{ridge } 1/2 \text{ width}}{\text{circular hole diameter}} = \frac{w_r}{\varnothing_i}$	
II8	$\tan\Phi$ = coefficient of internal friction	Internal friction angle (Φ) corresponds to the volcano repose angle. The tangent of this angle gives the internal friction coefficient.
II9	$\frac{\text{potential energy}}{\text{viscous force}} = \frac{g[(\rho \times h) + (\rho \times tb) + (\rho \times td) + (\rho \times \varnothing_i)]}{(\mu_s + \mu_i) \times u}$	This number describes the balance between the potential energy accumulated by the volcano (van Bemmelen, 1949) and the viscous forces related to the basal ductile layer and the intrusion.
II10	$\frac{\text{resistance force to the failure}}{\text{viscous force}} = \frac{[\tau_0(1 + 2 \tan \Phi \sqrt{b}) + g(\rho h + \rho tb) \tan \Phi (1 + b)] \times h}{(\mu_s + \mu_i) \times u}$	This number represents the balance between the resistance force to the failure of the brittle material (i.e. ridge and brittle layers) and the viscous forces, responsible of the flow of the ductile layers (basal substratum and intrusion).
II11	$\frac{\text{inertial force}}{\text{viscous force}} = \frac{(t_d \times u \times \rho_s) + (f \times \left(\frac{1}{\varnothing_i}\right) \times \rho_i)}{(\mu_s + \mu_i)}$	This number is the Reynolds number and corresponds to the balance between inertial and viscous forces occurring within the ductile material (i.e. silicone and golden syrup).
II12	$\frac{\text{spreading force}}{\text{intrusive flux}} = \frac{u \times h \times (tb + td)}{f}$	This number describes the influence of the intrusive flux and the horizontal displacement velocity on the deformation.

Π13	$\frac{\text{substratum viscosity}}{\text{intrusion viscosity}} = \frac{\mu_s}{\mu_i}$	This number corresponds to the balance between the viscosity of the substratum and the viscosity of the intrusion.
Π14	$\frac{\text{ridge density}}{\text{intrusion \& ductile layer densities}} = \frac{\rho_r}{\rho_s + \rho_i}$	This number describes the physical properties of the materials.

2.4. Material

Typically, pure sand is the analogue material of choice for brittle volcanic rocks (e.g. *Merle and Borgia, 1996*). Recently, sand has been mixed with plaster to obtain a slightly more cohesive material which scales better to natural rocks (*Donnadieu, 2000*). Furthermore, using this mix helps to observe much more structural detail at the surface due to the increased cohesiveness.

Silicone is the analogue material for the ductile rocks that can constitute part of volcanoes and/or the volcanic basement such as hyaloclastites, weathered horizons, and colluvium produced by erosion of the volcano, intrusive complexes, hydrothermally altered layers, and also basal sedimentary layers (*Merle and Borgia, 1996; van Wyk de Vries and Francis, 1997; Oehler et al., 2005*). The viscosities of these LSLs are poorly constrained, but are estimated at 10^{18} Pa.s by *Carena et al. (2000)* and at 10^{14-15} Pa.s by *Arnaud (2005)*. From these values, *Delcamp et al. (2008 and in review)* chose an intermediate value of 10^{17} Pa.s. In this study, as the same materials are used and the scaling is similar, the value of 10^{17} Pa.s for the viscosity of the natural ductile layers is maintained.

Golden Syrup is a good analogue for magma and has been recently used in several studies (*Holohan et al., 2008; Mathieu et al., 2008*). The viscosity of this material is temperature dependant. The experiments done in the present study were conducted at room temperature varying from 10 to 20°C, corresponding respectively to a variation of viscosity from 11.5Pa.s to 6.5Pa.s. Fortunately, intrusion viscosity is an important parameter for the horizontal displacement velocity, but has no influence on the style of deformation (*Delcamp et al., 2008*).

Magma viscosity varies from 10 to 10^{18} Pa.s depending on crystal content and melt composition (e.g. *Dingwell, 1999*). Following the scaling, Golden Syrup represents magma of intermediate viscosity that is sufficiently crystal-rich.

Eruption volumed data estimated ages and edifice volumes, deformation and degassing data allow us to approximate magma fluxes in nature (*Wadge, 1982; McNight and Williams, 1997; Hasenaka and Carmichael, 1985; Beauducel et al., 2000; Wicks et al., 2002*). The magma flux values estimated vary from 10^{-3} to 10^{-2} $\text{m}^3.\text{s}^{-1}$. Fluxes in models (without eruption) vary from 10^{-10} to 10^{-7} $\text{m}^3.\text{s}^{-1}$, corresponding respectively through scaling to fluxes of 10^{-5} - 10^{-2} $\text{m}^3.\text{s}^{-1}$ in nature. Consequently, models scale here from 1/10 to 10 times the estimated magma fluxes. Golden syrup fluxes thus cover the range of variation of natural fluxes.

2.5. Type A-models: results

2.5.1. Deformation

2.5.1.1. Surface structures

All the experiments were characterized by the early formation of two central faults on each side of the ridge summit and parallel to the ridge axis. For a similar ridge geometry and side-constrained experiments (using a frame), a Π_3 ratio (ridge width : silicone width) of around 0.8 allows the formation of only two parallel fractures (**Fig.4A**-model ridge30_1; $h= 4.3\text{cm}$; $w_r= 8.5\text{cm}$; $w_s= 9.5\text{cm}$). A $\Pi_3 < 0.8$ favoured more faults to form, still parallel to the ridge axis. The number of formed faults thus increases with diminishing Π_3 values (**Fig.4B**).

All the models were side-constrained within a wooden frame except the models ridge30_5 and ridge30_11 where multiple faults formed (**Fig.4C**).

In the case of multiple fault formation and constrained experiments, two central parallel faults formed first, and secondary faults formed afterwards, along one preferential flank (**Fig.4B**) or within the early-formed two faults. The secondary faults did not usually form sequentially. All the fractures remain active and widened throughout time.

Variation of ridge length did not influence the deformation.

Folds have been observed in the silicone at the foot of or at the surface of the sand-plaster ridge on several models (for example, ridge30_2, ridge30_13). One fold was observed along the flank of the model ridge30_7 ($h = 5\text{cm}$; $t_d = 1.9\text{cm}$).

Two models, where one side of the ridge had a slope at 20° and the other at 30° , tested the effect of a slight asymmetry (model ridge30_18 and 19). No difference was noticed compared to symmetric ridges.

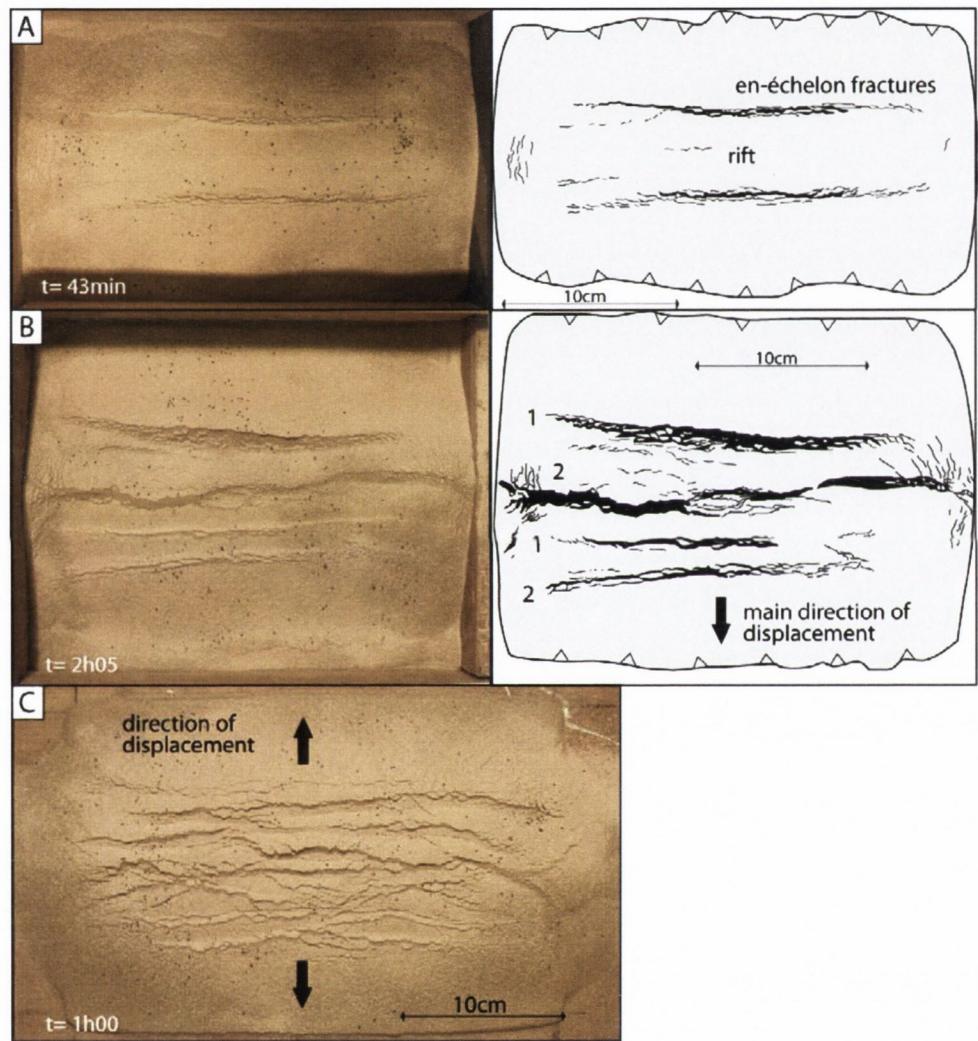


Figure 4: Type A-models are sand-plaster ridges lying on a basal ductile layer **A-** Only two parallel fractures formed during the experiment ridge30_1 (ridge width = 17cm; silicone width = 19cm). **B-** A decrease of $\Pi 3$ number allows more fractures to be formed (model ridge30_2; ridge width = 18cm, silicone width = 25cm). **C-** Model ridge30_5,

where multiple parallel fractures formed as the ridge was not constrained (ridge width = 16cm, silicone width = 19cm).

2.5.1.2. Deformation related to intrusive bodies and intrusion morphologies

Experiments with a ridge intruded by golden syrup have been performed. The deformation was characterized by the usual formation of faults parallel to the ridge axis. The deformation due to the intrusion emplacement was either concentrated perpendicular to the ridge axis (for example ridgeint30_1: $h=5\text{cm}$; $w_r=9.5\text{cm}$; **Fig.5A**) or above the pipe parallel to the ridge axis (for example ridgeint30_6: $h=4.5\text{cm}$; $w_r=7.2\text{cm}$; **Fig.5B**). The golden syrup propagated within the silicone layer, except when it erupted.

The models where the deformation due to the intrusion was mostly perpendicular to the ridge axis occurred for $\Pi_1 \leq 0.5$ (Π_1 = ridge height: ridge $\frac{1}{2}$ width; **Fig.5A**). At the end of the experiment, the golden syrup sometimes erupted at the foot of the ridge. The resulting intrusive complex exhibited a component of extension parallel to the ridge axis and another component that extended perpendicular to the ridge axis. The extremity of the perpendicular extension could present morphological variations. For example, the extremity was thicker at the foot of the model ridgeint30_1 (**Fig.5A**: $h=5\text{cm}$, $w_r=9.5\text{cm}$), but was flat for the model ridgeint30_8 ($h=6\text{cm}$, $w_r=12\text{cm}$).

For a smaller model, i.e. $\Pi_1 > 0.5$ (ridge height: ridge $\frac{1}{2}$ width), the intrusive complex propagated parallel to the ridge axis (**Fig.5B**), but we still observed some deformation perpendicular to the ridge axis. The golden syrup in that case erupted at the ridge axis (ridgeint30_6) or slightly offset from the axis, but no longer at the ridge foot. The resulting intrusive bodies are either slightly elongated and parallel to the ridge axis or have a pancake shape (more or less circular).

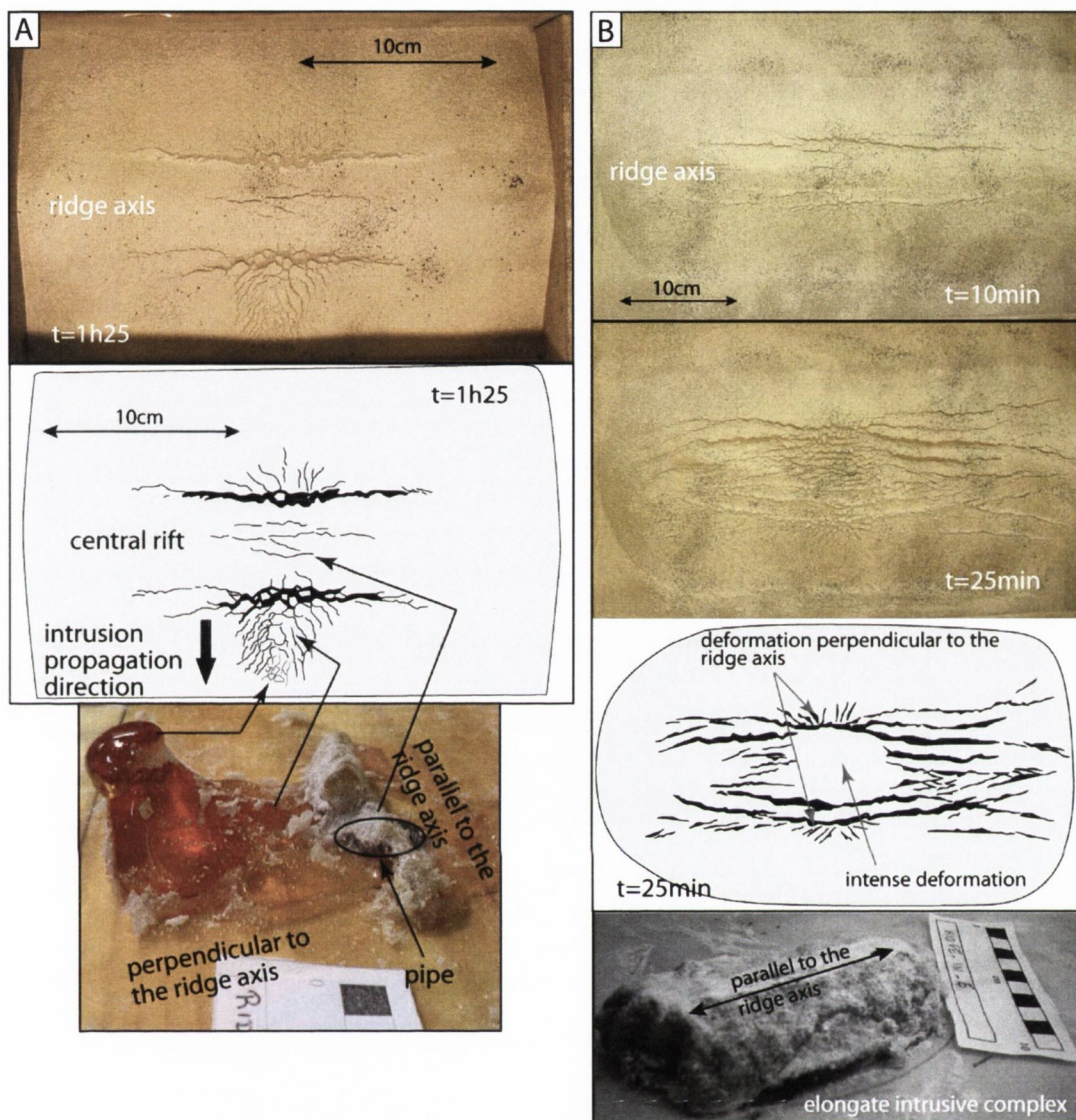


Figure 5: A- The intrusion (now frozen) propagated perpendicular to the ridge axis along one flank during the experiment ridgeint_30_1 (ridge height= 5cm; ridge width= 19cm). B- The intrusion mostly propagated parallel to the ridge axis, resulting in symmetric deformation (model ridgeint_30_6; ridge height= 4.5cm; ridge width= 14.4cm).

2.5.2. Horizontal displacement

The maximum horizontal displacement occurred at the first stage of the experiments, usually the first 10 minutes (**Fig.6**). Then, the velocity decreased gradually until 30 to 40

minutes where it became slow and almost constant. For the first 10 to 20 minutes, the horizontal displacement velocity of the model ridge30_7 ($h= 5\text{cm}$; $t_d= 1.9\text{cm}$) is higher than the other experiments.

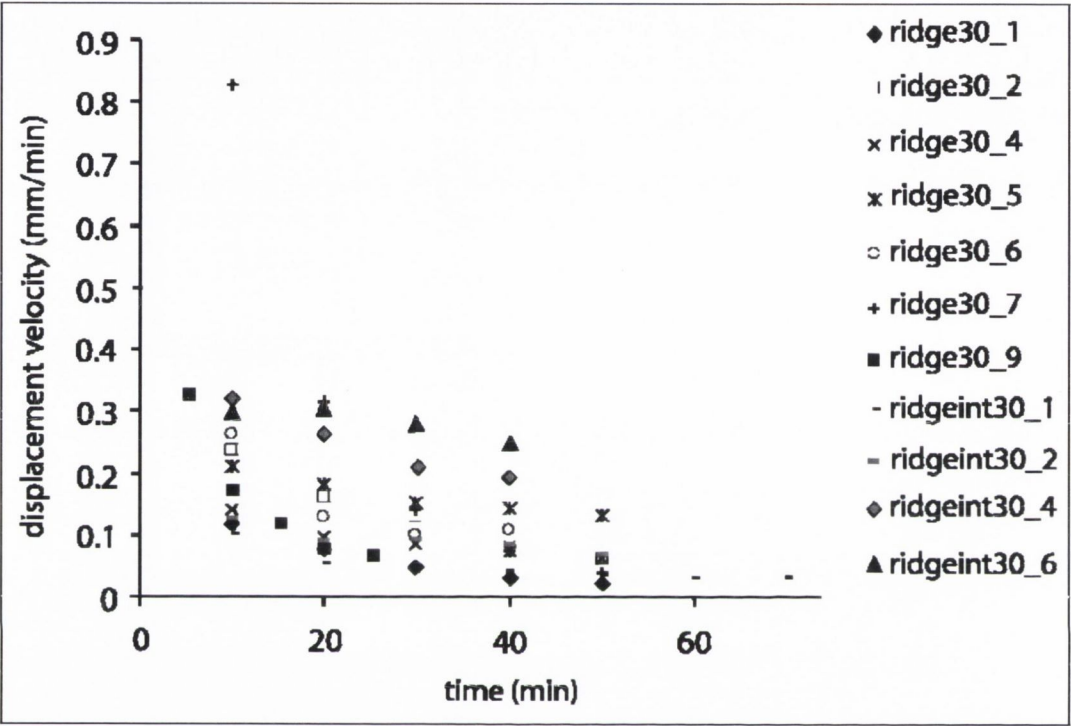
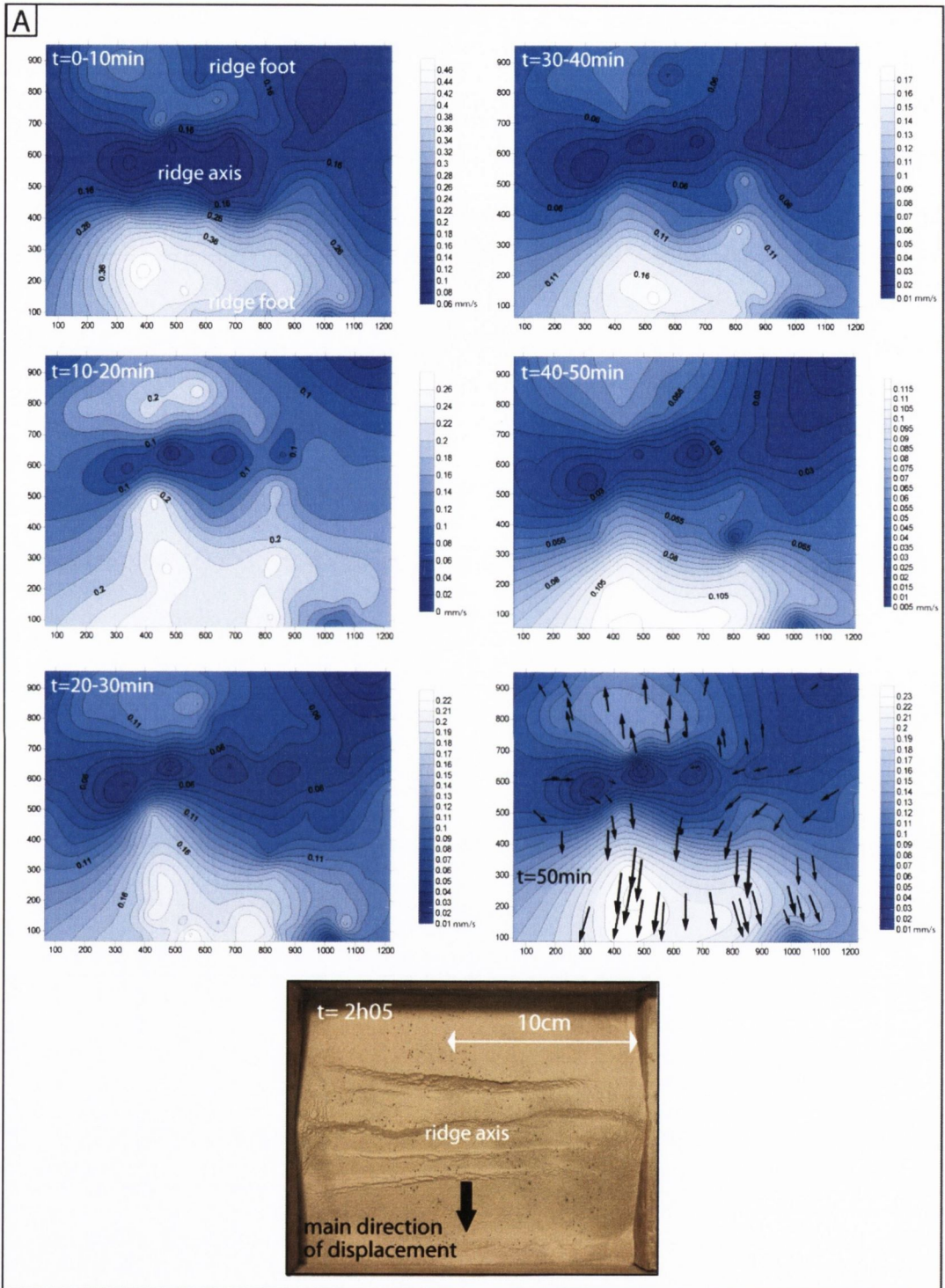


Figure 6: Horizontal displacement vs. time for the models of set up A. Both experiments with (ridgeint30_x) and without intrusion (ridge30_x) are displayed for comparison.

The majority of the models followed a similar deformation pattern where the ridge spread outward perpendicular to its crest (**Fig.7A**). The horizontal displacements were either symmetric or asymmetric to the ridge axis. This latter case occurred when the secondary rifts formed preferentially along one flank (e.g., model ridge30_2, $h= 4.8\text{cm}$; $w_r= 9\text{cm}$; **Fig.4B** and **7A**). The resulting displacement rates were greater along this flank (south side on **Fig.7A**). At the two extremities of the ridge, the particles were moving towards the centre of the ridge before moving outward along a curved trajectory (see arrows on **Fig.7A** on the last sketch showing the total displacement field over 50min). An example of symmetric horizontal displacement is shown in **Fig.7B**. We note that at the extremities of the ridge axis, the black markers are still moving towards the centre. In all the

experiments, the maximum horizontal displacement did not occur along the ridge axis but rather along the ridge flanks (**Fig.7**).

In contrast, for the experiments with intrusion, the maximum of horizontal displacement occurred around the location of the intrusive complex (**Fig.7C**).



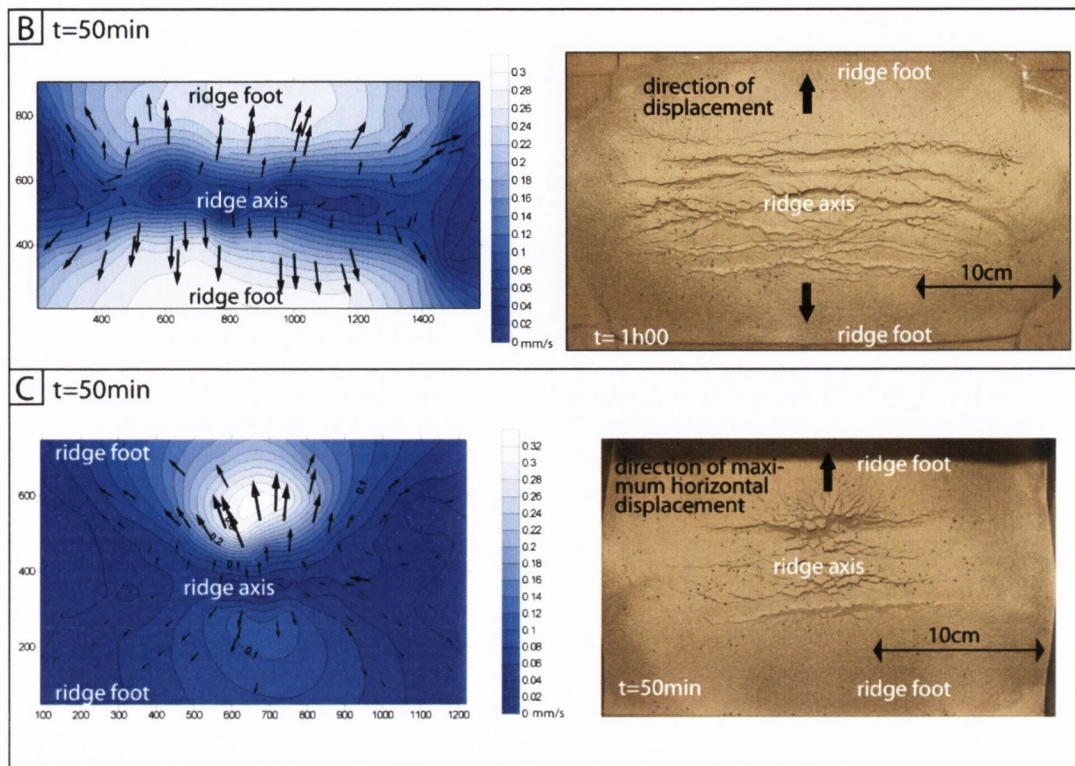


Figure 7: A- Horizontal displacement fields for the experiment ridge30_2 during regular time intervals ($h= 4.8\text{cm}$; $w_r= 9\text{cm}$). The last diagram corresponds to the average displacement field for the first 50 minutes and the vectors show the total displacement of the black markers during this time. B- Average of the horizontal displacement field for the experiment ridge30_5 ($h= 4.7\text{cm}$; $w_r= 8\text{cm}$) for the first 50 minutes with associated displacement vectors. C- Average of the horizontal displacement field for the experiment ridgeint_30_2 ($h= 4.8\text{cm}$; $w_r= 9.5\text{cm}$) for the first 50 minutes with associated displacement vectors. The intrusion emplacement is clearly apparent with the maximum displacement located above the intrusive complex. The scale bars displayed on the right are in mm/s.

2.5.3. Interpretation

For comparable ridge geometry (width, length, slope angle), the number of faults formed increases with the width of the basal ductile substratum. The models were all constrained by rigid boundaries except the models ridge30_5 and ridge30_11, where multiple faults were formed (**Fig.4C**). A wider basal ductile substratum involves less influence from the

model boundaries. Consequently, the ridges will be able to spread more easily, allowing more faults to form. The number of faults thus depends on the degree of constraint from the model boundaries.

In the case of multiple fault formation, the later faults preferentially appeared most of the time along one flank, offset from the ridge axis (**Fig.4B**). This shift reflects an asymmetry of the deformation that might be linked to a difference of morphology between the flanks, even if the models were built with care to avoid any asymmetry. I thus performed two models with different slope values: one side of the ridge was 30° and the other side 20° (models ridge30_18&19) to check if slight flank slope asymmetry has an influence on the deformation field. However, no systematic difference has been noticed. The reason why we sometimes observe symmetric or asymmetric deformation relative to the ridge axis is therefore likely due to slight defects in the basal layers. Indeed, a slight variation of basal layer thickness or dip in a spreading volcano can induce asymmetric deformation, i.e. the volcano will spread preferentially towards one direction (*Wooler et al., 2004; Delcamp et al., submitted*).

The folds sometimes observed at the foot of the ridges (**Fig.4B**) are likely to be due to spreading (*van Bemmelen, 1949; Borgia, 1994; Merle and Borgia, 1996; Borgia and van Wyk de Vries, 2003*). The ridge load pushes and forces the silicone to flow outwards resulting in fold formation at the ridge foot.

The experiment ridge30_6 ($h = 4.5\text{cm}$; $t_d = 3.8\text{cm}$) has a thick basal ductile layer compared to the other models, corresponding to a low Π_2 (ridge height + brittle substratum thickness : ductile substratum thickness). A high Π_2 allowed the model to spread while a low Π_2 results in sagging of the model (*van Wyk de Vries and Matela, 1998*). The model ridge30_6 is thus expected to sag rather than to spread. The south flank shows an inward displacement that may reflect the progressive sagging of the ridge. However, as the north flank was not constrained (this side of the frame broke during the experiment), this side was able to spread outwards. Consequently, there is only one direction of displacement. The displacement vectors of the model ridge30_7 ($h = 4.5\text{cm}$; $t_d = 1.9\text{cm}$) reflect a spreading of the ridge (i.e. outwards displacement) but a fold within the ridge was observed, reflecting compression. This fold can be regarded as an analogue to the terraces observed in sagging giant volcanic edifices (*Byrne et al., 2009*). This experiment displays

both spreading and sagging characteristics with an intermediate Π_2 (slightly higher than the ridge30_6) and thus stands at the limit between spreading and sagging. On the two experiments ridge30_15 ($h=4\text{cm}$; $t_d=3\text{cm}$) and ridge30_16 ($h=3\text{cm}$; $t_d=3\text{cm}$) only fractures have been observed and no terraces formed. However, these ridges did spread and sag. Similarly to the experiment ridge30_7, these two experiments covered a spreading and sagging regime.

The displacement velocity was high during the first minutes of the experiments and stabilized afterwards (**Fig.6**). This has been already noticed for spreading volcano analogue models by *Delcamp et al. (2008)*, where they deduced that the velocity decrease corresponded to a progressive decrease of volcanic load while the edifice spread.

The intrusive bodies developed only a slight along-axis propagation. For relatively high models ($h \geq 5\text{cm}$), the intrusive complex instead propagates perpendicular to the ridge axis in a lateral direction. The perpendicular extension developed different morphologies depending on whether basal eruption occurred. If a golden syrup eruption occurred, the intrusion developed a protusion that fed the eruption (see **Fig.5A**) while it remained “flat” when no eruption occurred (ridgeint30_8; $h=6\text{cm}$; $w_r=12\text{cm}$). For less high models, the intrusive complex still propagates perpendicularly, but also slightly parallel to the ridge axis and erupts at the centre or slightly offset from the ridge centre. The propagation perpendicular to the ridge axis is confirmed by the irregular pancake shape displayed by most of the intrusive complexes.

On the horizontal displacement vs. time graphs, the experiments with intrusion show the same patterns to the models without intrusion except the model ridgeint_30_6 (**Fig.6**). The slight increase of velocity at 20 minutes followed by a slow decrease might correspond to the golden syrup rising within the model. It is, however, not reflected in the other experiments, probably because the local effects due to intrusion are hidden by the other individual velocity values (see **section 2.2**). Also, the black markers placed around the intrusive complex were usually lost during the experiments due to large horizontal displacements. Plaster was sieved onto the models allowing the structures formed to be easily observed. Around the intrusive complex, the surface of plaster was intensely cracked forming individual plates that progressively separated. The increase in horizontal

displacement around the intrusive complex can be observed with the naked-eye, but can not be followed with the markers on time-lapsed images.

2.6. Type B-models: results

2.6.1. Deformation

2.6.1.1. Surface structures

Two main types of structures have been observed: a progressive sector sliding along one flank or a summital bulge. Usually, a bulge was preferentially formed when the intrusion was just under the ridge crest, while sector creep was favoured when the intrusion was offset from the summit, i.e. underneath the flank. Note that the position of the pipe (at the base or slightly above the base) did not have an effect on the deformation.

The sector sliding was characterized by a gently arcuate fault or straight fractures arranged in an arcuate pattern at the head of the slide and folds or thrusts at the front (**Fig.8**). Usually, the deformation started with the formation of arcuate fractures that were sometimes arranged en-échelon. Then, new fractures, generally arcuate, were formed behind the early formed fractures, i.e. opposite to the sliding direction. At the same time, folds started to form at the front of the sliding sector. Fractures appeared randomly and progressively at the surface of the creeping sector. In general, after 10-15 minutes, no new fractures were formed at the head of the sliding sector and the previously formed fractures simply widened. Along the side of the sector, en-échelon fractures were sometimes observed. The length and the width of the folded front increased while the deformation progressed. The folded zone eventually reached the head of the sliding sector (**Fig.8** at 26min). The model RPC1 ($h=5\text{cm}$; slope: 30°) even displayed a tectonic inversion where an extensional fracture at the edge of the creeping sector head was reused as a thrust. While the sector was sliding outwards, a few individual blocks were uplifted and tilted towards the direction of the sector displacement. The sector progressively lost its coherence and sometimes the blocks lost their cohesion entirely. Usually, the sector stopped creeping when an eruption occurred (i.e. when the golden syrup appeared at the surface).

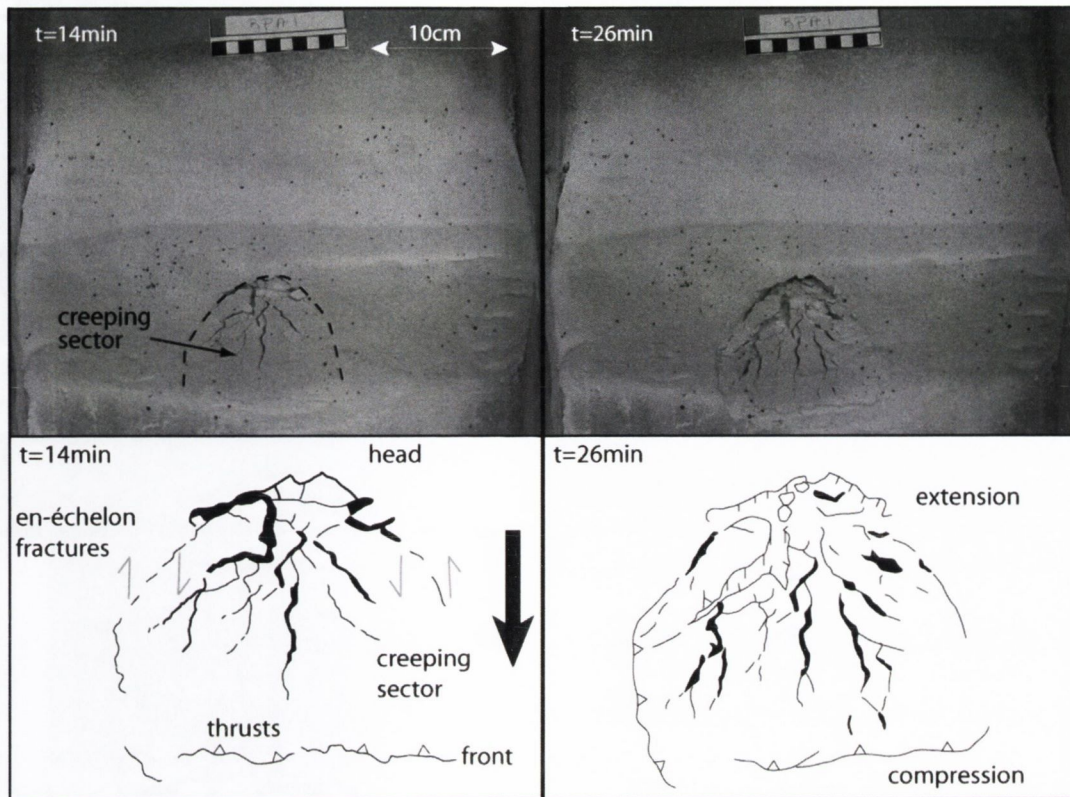


Figure 8: *RPA1 experiment ($h=5\text{cm}$; 30°), showing a sector creep with the associated structural features such as en-échelon fractures and thrusts.*

The summit bulges were characterized by the early formation of fractures concentrated above the intrusion (**Fig.9**). While the deformation progressed, the fractured area grew vertically, forming a bulge, and others fractures appeared at the surface of the uplifted zone. In the meantime, thrusts and folds formed around the bulge. The deformation stopped as soon as the golden syrup erupted.

Note that in two models, only fractures were observed without significant bulge formation or sector creep (RPC2 $h=5\text{cm}$, slope= 20° and RPC4 $h=6\text{cm}$, slope= 30°).

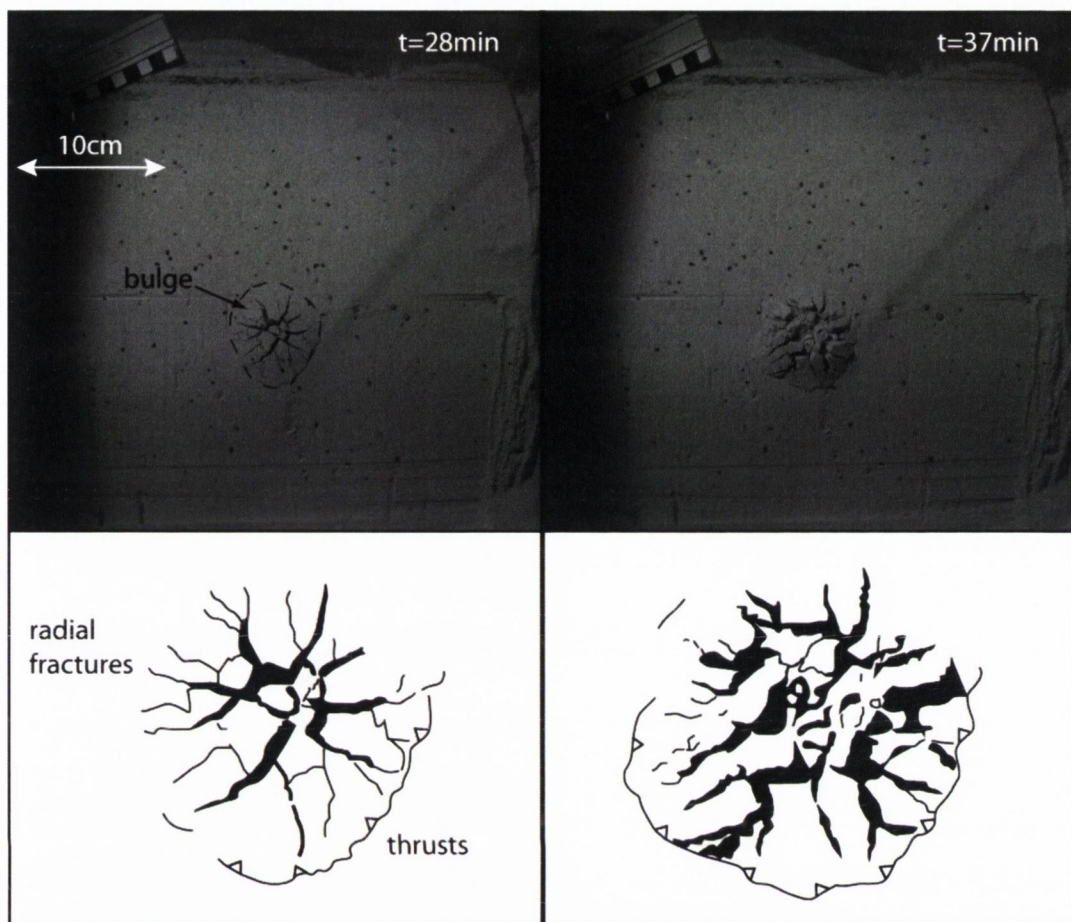


Figure 9: *RPC3 experiment ($h=3\text{cm}$; 10°) showing a bulge with the associated structures.*

2.6.1.2. Intrusion morphologies

Two types of intrusion shapes have been recognized (**Fig.10**). In general, an elongate cup-like intrusion (cf. *Mathieu et al., 2008*) was formed following sector creeping (**Fig.10A**), while the models with summit bulges formed a cylindrical cup-like intrusion (**Fig.10B**). Some en-échelon protrusions have been observed along the edge of the elongate cup-like intrusion.

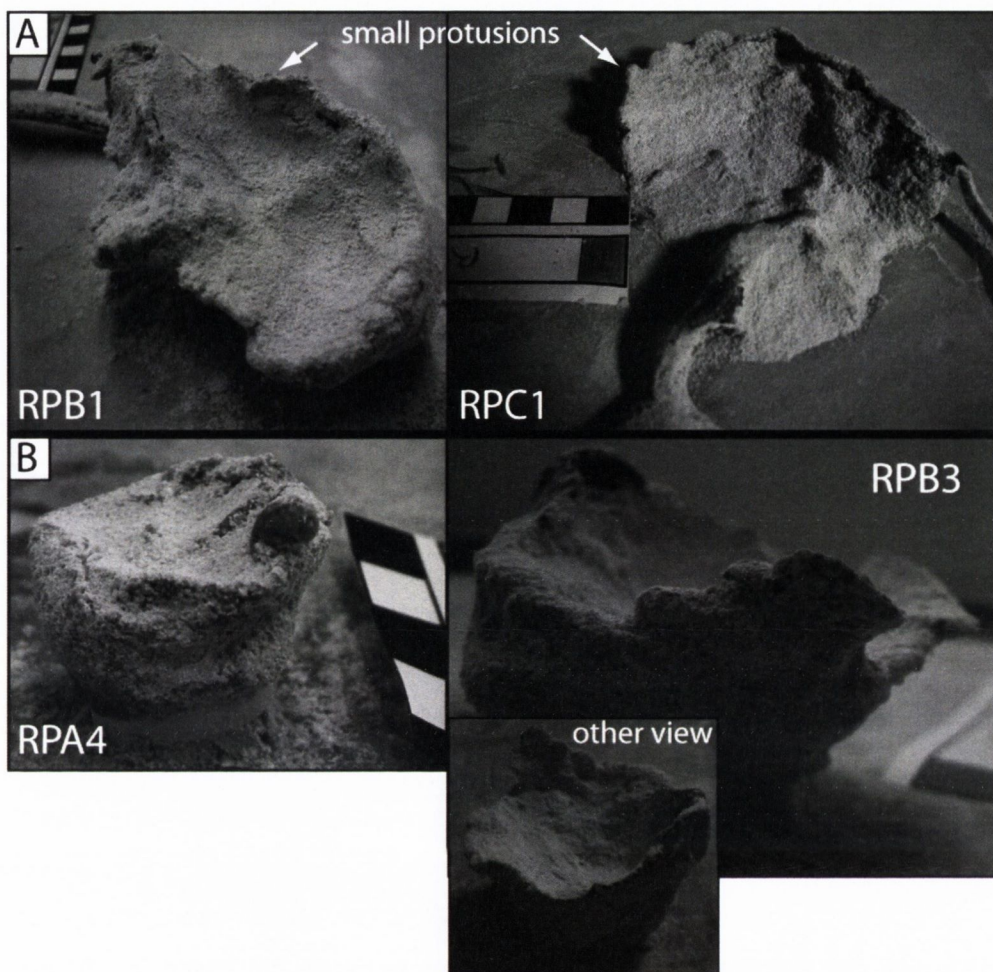


Figure 10: Typical intrusion morphologies obtained with the models of set up B. Note that intermediate shapes occurred, especially when a bulge was formed but started to slide outwards, resulting in a slightly elongated cup-like intrusion. **A-** Elongated cup-like intrusion that formed following sector creeping. **B-** Cylindrical cup-like intrusion that formed following summital bulge formation.

2.6.2. Horizontal displacement

Unfortunately, it was not possible to estimate the horizontal velocity for all the models, especially for the experiments leading to bulge formation as the displacements were mostly vertical. Some of the sliding experiments were also too rapid (model RPA2; $h=5\text{cm}$, $\text{slope}=30^\circ$) or the deformation was so intense that the black markers were mixed and rapidly lost within the sand and the plaster.

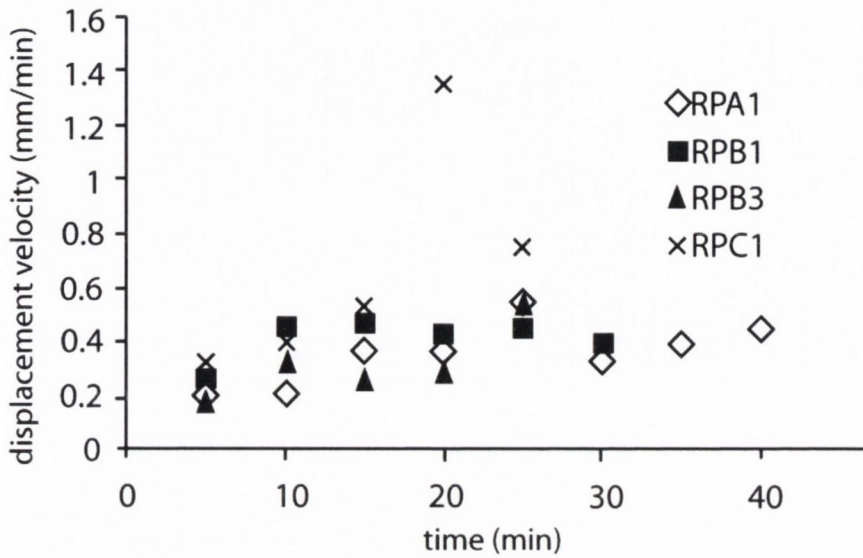


Figure 11: Horizontal displacement vs. time profiles several experiments of the set up B.

As mentioned in **section 2.2**, the horizontal velocities are not highly representative as during the same time interval, the displacement rates varied within the sliding sector. This is strongly reflected in the horizontal displacement vs. time profiles where the velocity instead increased or decreased through time (**Fig.11**). We sometimes note an acceleration at the end of the experiments that correspond to the collapse of the sector. The data can be used for the deformation field evaluations (**Fig.12**). Depending on the experiments, the displacements were sometimes greater at the head and sometimes greater at the front of the creeping sector. For all of the models, however, an outward spreading is observed while the sector is sliding. On **figure 12**, the position of the maximum horizontal displacement varied through time, but is localised around the head of the sector except between 20-25 minutes where the maximum is around the front of the sector. From 35 to 40 minutes, there are two maxima at the head and at the front of the sector.

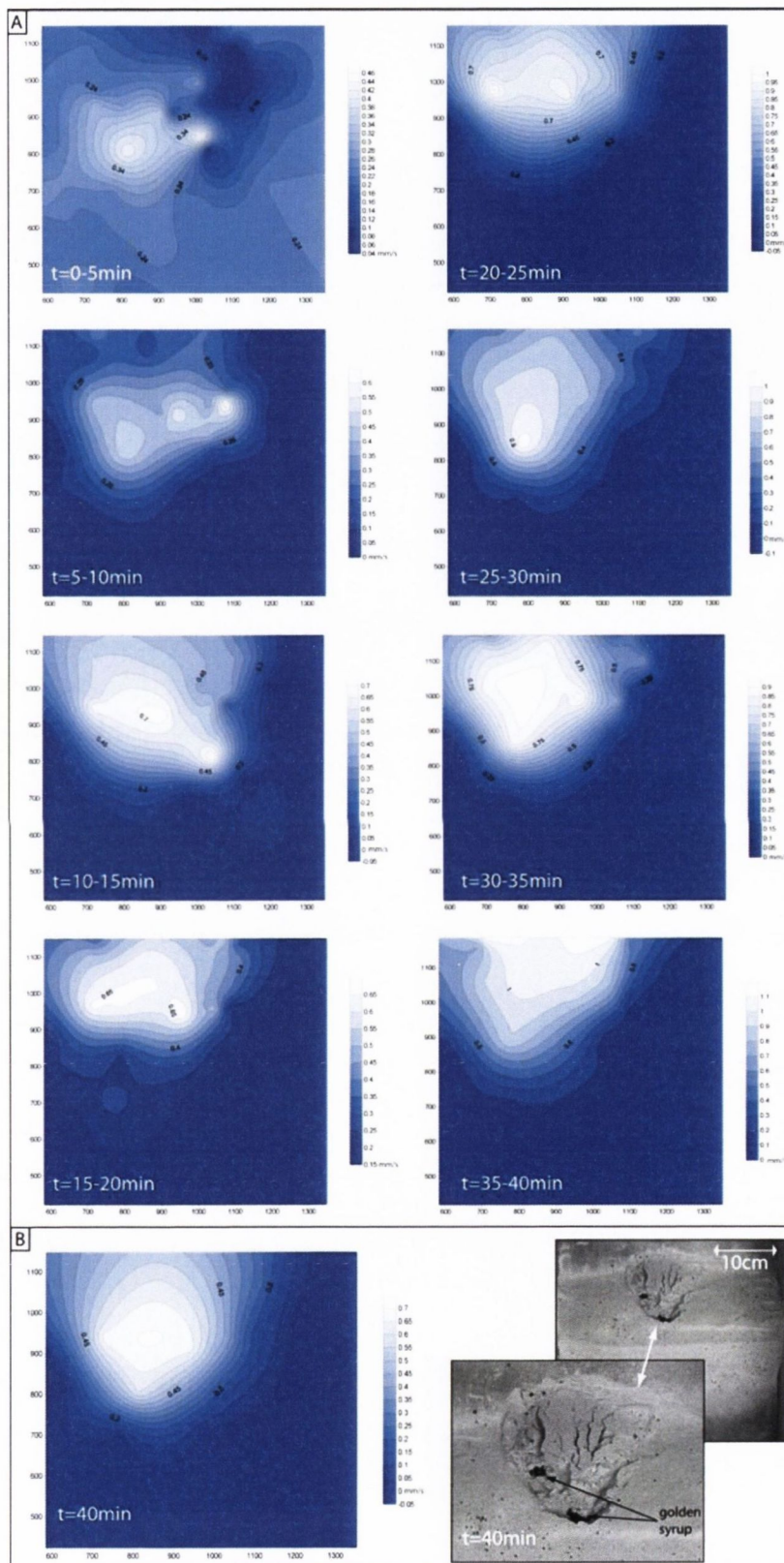


Figure 12: Horizontal displacement fields for the model RPA1 ($h=3\text{cm}$; 10°). **A:** The figures represent the variation of the displacement during regular time interval (5 minutes step); **B:** This figure shows the average of the displacements during 40 minutes with the associated displacement vectors. The scale bars displayed on the right are in mm/s.

2.6.3. Interpretation

The main factor determining the structures formed, i.e. either a bulge or a sliding sector, depends on the position of the intrusion within the ridge. An intrusion growing underneath the crest favours bulge formation, while a sector of the flank will creep when the Golden Syrup intrudes within the flank. The lithostatic load is maximum at the ridge crest, therefore an intrusion growing underneath the summit is partially blocked, allowing a bulge and a cylindrical cup-like intrusion to be formed. When the intrusion is emplaced within the flank, the height of the material above and thus the lithostatic load is lower. In consequence, the system spreads outwards, i.e towards the un-buttressed side (**Fig.8**). Another solution to explain bulge formation or sector creep is to consider a symmetric stress field at the ridge crest vs. an asymmetric stress field along the flank. At the ridge crest the stress field is symmetric relative to the ridge axis. Consequently, the intrusion rising through the model propagates upward and pushes the flank (more or less symmetrically) perpendicular to the ridge axis. On the other hand, an intrusion rising through the flank will follow the easiest direction of propagation determined by the asymmetric stress field and will thus propagate outwards, perpendicular to the ridge axis favouring sector creep.

Sector flank sliding is triggered here by the outward propagation of the intrusion. The progression of the deformation observed at the surface reflects the progression of the intrusion within the model. The first stage of intrusion emplacement is accommodated by extensional structures, typically normal faults formed above the golden syrup pipe. As the intrusion propagates outwards, the extension increases at the head and one sector of the flank slides outwards to accommodate the growth of the intrusion. Consequently, the push of the intrusion and the sector creep are responsible for fold formation at the front. The en-échelon fractures which sometimes formed along the sliding sector show that strike-slip movements also accommodate the deformation. The small en-échelon protrusions observed along the elongate cup-like intrusion may reflect the en-échelon fractures and the strike-slip movements observed at the surface.

The Golden Syrup shows a period of inflation with bulge and uplifted block formation and also a period of deflation regarding the intrusion shapes (cup-like). The horizontal

displacement variations through time reflect these periods where an increase of displacement rate may reflect an inflation period and a decrease may reflect a deflation episode. The inflation period may be related to the ascent of Golden Syrup within the model and the deflation period may correspond to lateral propagation of the intrusion. The variation of the displacement rate within the sliding sector is probably linked to the preferential path followed by the intrusion and controlled by the surrounding local stress field.

2.7. Type C-models: results

2.7.1. Deformation

The structures formed in these models are similar to those obtained with the B models, where sector flank creep and bulges were reproduced. In general, low slope models (10°) lead to bulge formation while flank creep was observed for all the other experiments. Bulges did not appear all along the slit but rather formed within a limited area (**Fig.13**). At the beginning of the experiments, one main fracture was formed just above the slit and parallel to it. Sometimes, secondary fractures (less long and less wide) were formed simultaneously or few seconds later, more or less parallel to the main one (**Fig.13A**). Then, a bulge developed while the intrusion grew (**Fig.13B**). As observed for the B models, bulges were characterized by radial fracture formation within the uplifted part and by a thrust belt at the foot. The formed bulges were either circular (**Fig.13B**) or elongated. The intrusion shapes are generally similar between the experiments and are defined by two crests parallel to the slit that delimit a central depression (**Fig.13C**). One of the crests can develop a protusion that corresponds to the location where the bulge was formed at the surface. These small extremities could represent small dyke-like intrusions.

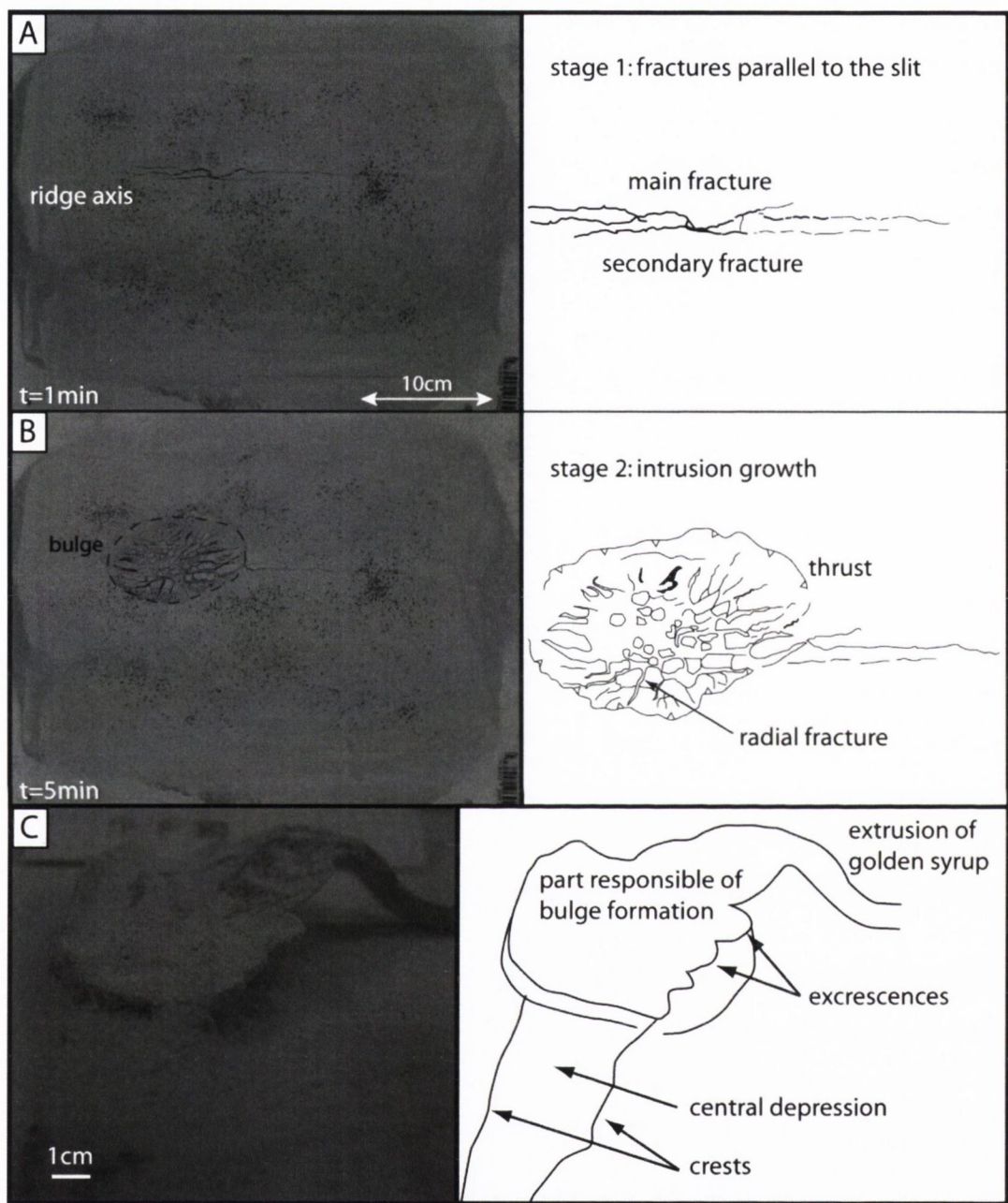


Figure 13: Bulge formation during the experiment RF10 ($h=3.5\text{cm}$; 20°) with associated structures. **A-** First stages of the experiment ($t=1\text{min}$). **B-** Bulge growth ($t=5\text{min}$). **C-** resulting intrusion morphology.

Sector flank spreading was characterized by three main phases (**Fig.14**). In a first phase, similarly to the bulges, fractures parallel and above the slit were formed (**Fig.14A**). Two

cases occurred: either only one major fracture was formed, or the main fracture (wider and longer) was associated with minor fracture. This second fracture was arcuate or almost straight. Sometimes, this second minor fracture was in fact composed of several fractures that progressively joined. In between these two fractures (the main and the minor), other smaller ones were sometimes formed while the deformation progressed.

The second phase was characterized by the creeping of one flank (**Fig.14B**). When the intrusion was offset from the ridge crest, i.e. along one flank, the creeping affected this side. When the intrusion was below the ridge summit, the creeping affected the flank where the main fracture was formed. For example, on **figure 14**, as the main fracture is just to the south of the crest, the flank will slide southward. This fracture delimits the upper part of the creeping block. As the intrusion rises, the flank is uplifted. Radial fractures developed from the main fracture at the upper part of the sliding block and thrusts formed at the foot. Behind the creeping block the extension increases while the block is sliding. The sliding does not occur all along the ridge and the slit but affects only a sector.

The third phase is characterized by the break-up of one part of the sliding block (**Fig.14C**). This stage occurs just before an “eruption” (i.e. the golden syrup appears at the surface). During the experiment, a narrowing of the deformation is observed from the first stage where fracture(s) is/are formed all along the slit to the last stage where only one part of the sliding block is totally broken. Note that in contrast to the B models, the boundaries of the sliding blocks are not delimited by fractures or en-échelon fractures, but only by thrusts. Similar to the experiments where bulges were formed, the intrusions are made of a base composed by two crests that delimit a central depression (**Fig.14D**). The crest(s) (only one or the two) are higher and thicker in one area, and only one of them has forms a protusion characterized by a depression that mimics the slide shape.

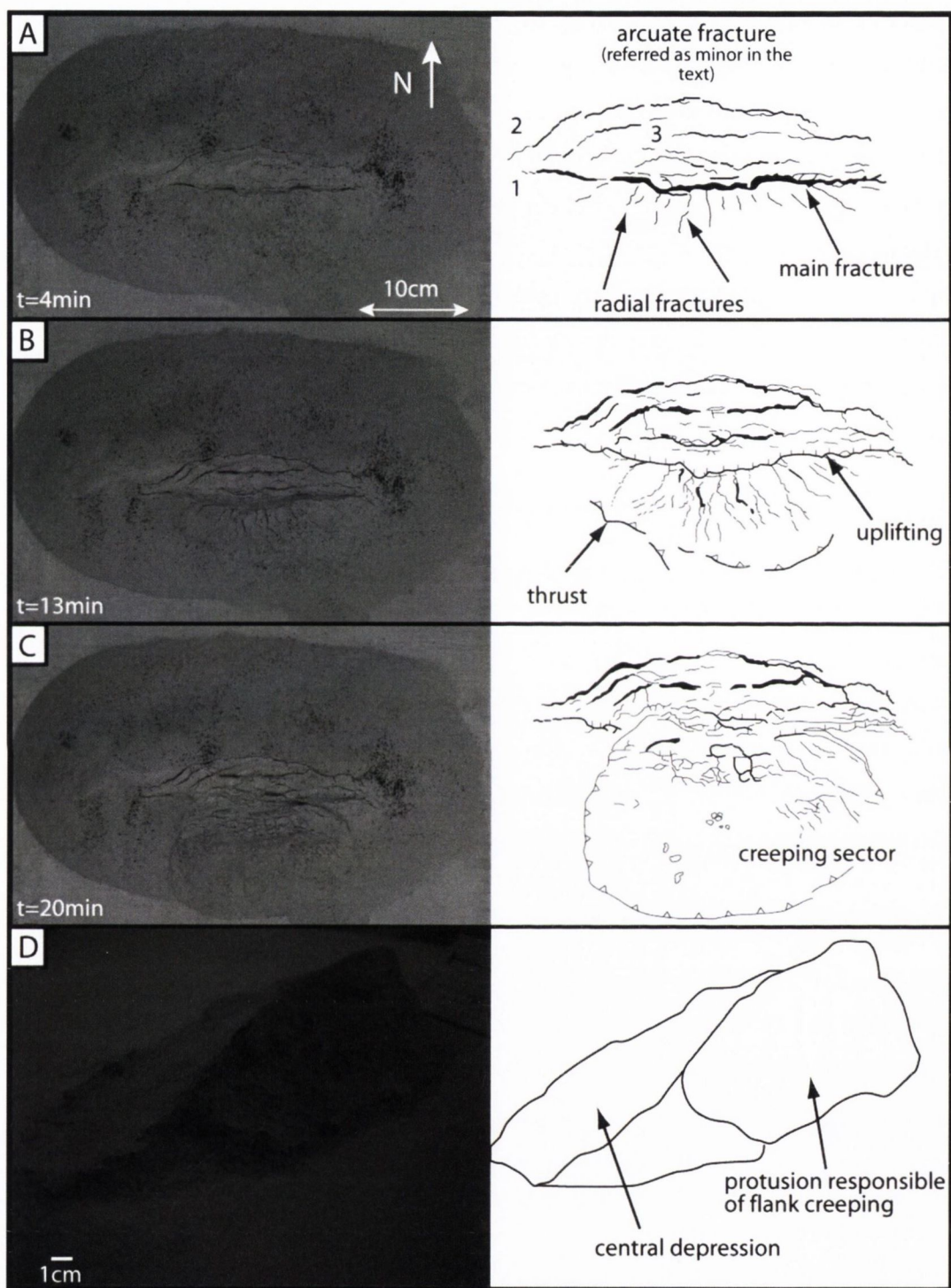


Figure 14: Sector creeping formed during experiment RF18 ($h=7\text{cm}$; 30°) and associated structures. **A-** First stage. **B-** Second stage. **C-** Last stage. **D-** Resulting intrusion shape.

2.7.2. Horizontal displacement

Similar to the B models, the displacement rates are not homogeneous along the ridge and within the sliding/bulging area. The calculated displacement velocities are thus given for information (**section 2.2**). It is preferable to use the displacement fields to better assess the displacement pattern during the experiments.

The models leading to bulge formation are characterized by intense vertical displacement with limited horizontal component. The displacement fields can thus not be determined as only the horizontal displacements were tracked.

The velocity profiles (**Fig.15**) usually show an increase of the velocity during the first minutes of the experiments. Then the velocity slightly decreases but remains almost constant. Finally, the velocity increased towards the end of the experiment.

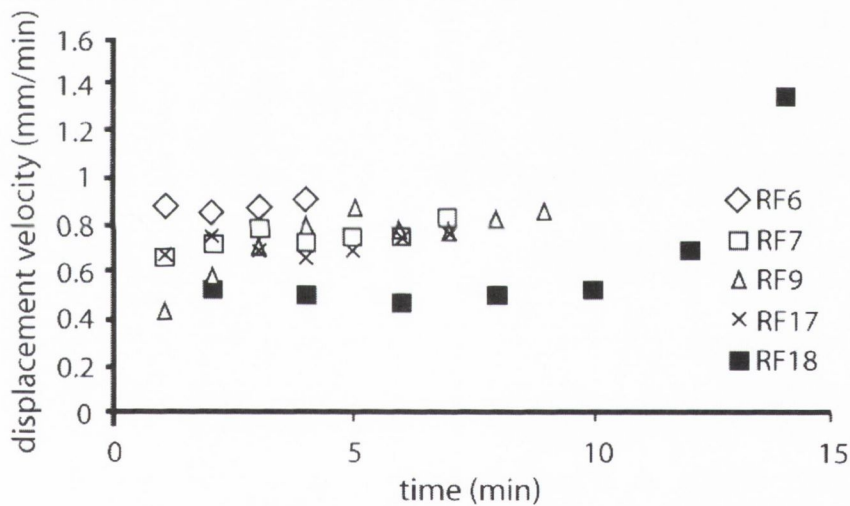


Figure 15: Horizontal displacement vs. time for several experiments using set up C.

Figure 16 shows the displacement pattern for an experiment with sector creep (RF18; **Fig.14**). During the first 4 to 6 minutes, the horizontal displacement field contours are elongated (slightly following a fan shape) and concentrated at the head of the sliding block. Outside of the sliding flank, the ridge is totally stable. Then, the zone of maximum displacement shifted towards the south (corresponding to the body of the sliding sector) and the contours are more rounded. Some displacement fields display an asymmetric fan shape. The three phases previously described which correspond to the narrowing of the

deformation with time (**section 2.7.1**) are reflected in the deformation fields. We clearly see that the last phase of deformation (12-14 minutes) corresponds to a fast slide of the sector (see scale values and **Fig. 14C**).

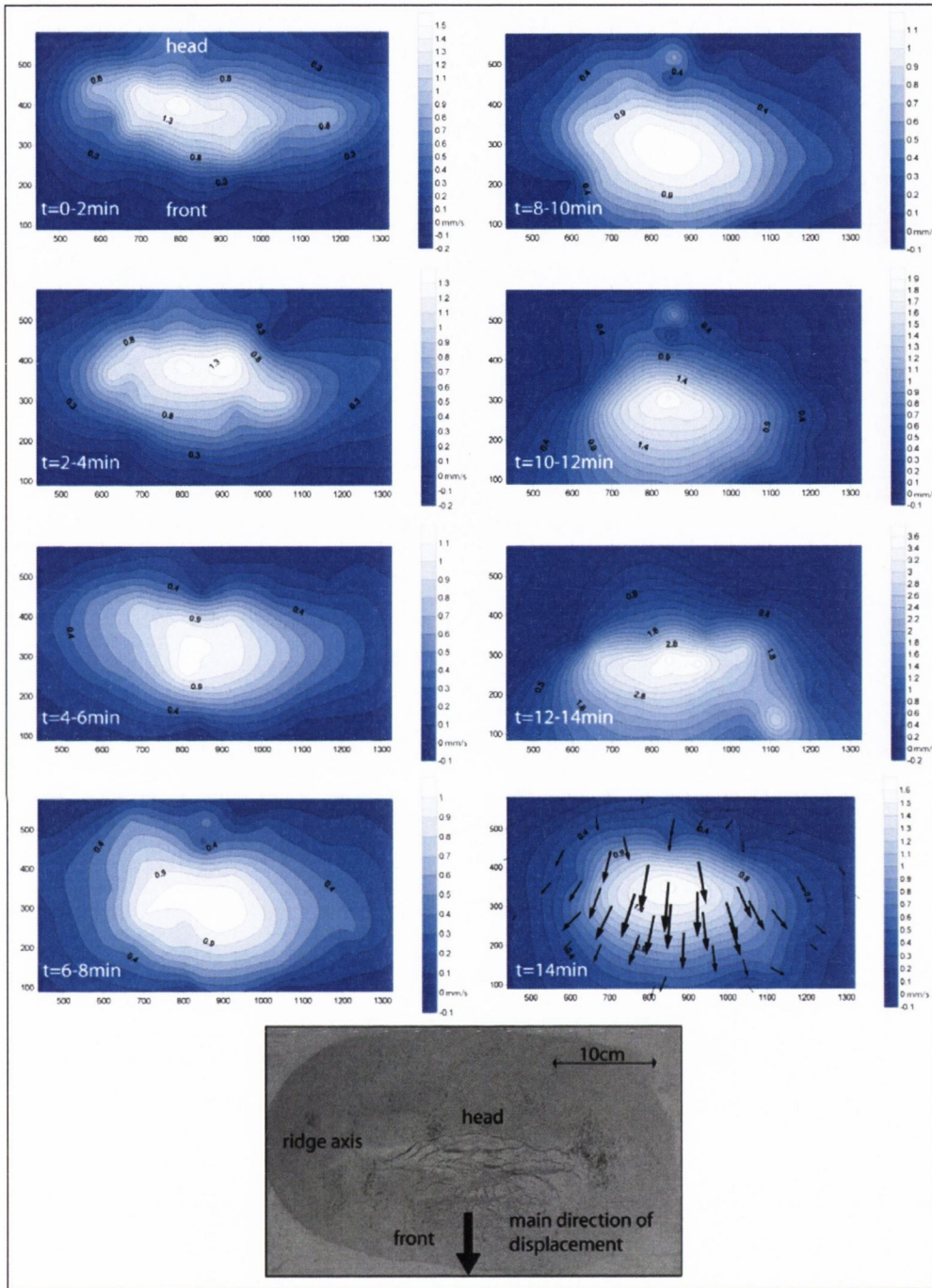


Figure 16: Horizontal displacement fields for the model RF18 ($h=7\text{cm}$; 30). The figures represent the variation of the displacement during regular time intervals (2 minutes per step), except the last one that shows the average of the displacements during 14 minutes with the associated vector displacements.

2.7.3. Interpretation

The structures formed using this set up are similar to the ones obtained with the B models (i.e. bulges and sector sliding). However, for the B experiments, the occurrence of bulge or sliding was related to the model height and to the offset of the ridge crest from the intrusion pipe. A high ridge with a central intrusion favours bulge formation and a smaller ridge with an offset intrusion favours sector creep. It has also been suggested earlier that the occurrence of bulge or flank sliding was due to the stress field state. The ridge crest is characterized by a uniform stress field that favours bulge formation while the flank is characterized by an asymmetric stress field favouring flank creep. In the C-experiments, bulges were observed for all the low slope (10°) experiments and in two experiments with a 20° slope, independently of the position of the ridge related to the slit. It is thus different from the results obtained with the B models. The bulges are in general formed for $\Pi I < 0.4$, i.e. a relatively small ridge. The position of the ridge crest relative to the slit is not the parameter controlling the deformation. In this set up, the factor that determines the types of structures formed seems to be only the stress field pattern - almost symmetric in the case of the low slope models and asymmetric for high slope models. The shape of the feeding system, i.e. elongated or circular (slit or pipe), thus changes the stress conditions radically and with it the deformation style.

The two intrusive crests rising from the slit and parallel to it following bulge formation are responsible for the early fracture(s) formation. The limited protusions that sometimes stick out through the crests are responsible for surface bulge formation and correspond to the golden syrup that appeared at the surface.

Three phases observed during flank creeping are recognized through the study of the intrusion shape, where the basal crest(s) is/are responsible for the early fracture formation. Then one sector of the flank slides due to the propagation of the intrusion whose morphology reproduces the creeping block shape. The last phase of deformation is sometimes reflected on the intrusion body when its summit forms a smaller protusion.

The displacement fields clearly reproduce the narrowing of the deformation through space and time. Some models display an asymmetric displacement field due to a preferential path followed by the intrusion within the model. The horizontal displacement velocities are higher early in the model run as the ridges still have their maximum height

and have thus more potential to be destabilised when the intrusion is rising. Then, the velocities remain constant or slightly decrease as the system reached a semi-stable configuration by flank creeping. The last stage of deformation is characterized by an increase of horizontal displacement velocity, corresponding to a rapid flank sliding, that can be regarded as a collapse and is due to the propagation of the intrusion that is getting closer to the surface.

3. Discussion

3.1. Comparison with previous models

Gelatin ridge models intruded by ink have been previously used to model dyke emplacement within rift zones (*Fiske and Jackson, 1972; Walter and Troll, 2003*). *Fiske and Jackson (1972)* modelled ridges where dyke emplacement was controlled by gravitational load (**Fig.1A**). They concluded that dyke orientation depends mainly on the rift zone geometry, with a straight ridge allowing the formation of a straight rift zone and, similarly, and a curved ridge develop around a curved rift zone. When intruding the gelatine ridge, the ink propagates easily and extends widely along the rift axis. In contrast, in our models, the intrusions within the ridge do not propagate extensively along the long axis. Consequently, no rift curvature was observed. More recently, *Walter and Troll (2003)* modelled dyke emplacement in an unstable conical edifice (**Fig.1B**). The destabilisation was triggered by a sector basal weak layer, contrasting to our models where sector creep was self-generated by the intrusion propagation (set ups B and C). In their models, the dyke orientation pattern depended on the position of the injection point related to the unstable sector. When the ink was intruded away from the creeping sector, radial fractures were produced. An injection within the unstable sector favoured the formation of collinear rift zone. Triple-armed rift zones were produced by ink injection at the boundary between stable and unstable sectors. In that case, the dykes wrapped the collapse scar and a weak third arm formed perpendicularly at the head of the collapse embayment. The three arms are about 120° from each other. In this study, only the A models with intrusion reproduced a kind of “triple-geometry”. The part parallel to the ridge axis can be considered as two collinear arms (on each side of the pipe) with a third perpendicular arm. The angles between the arms are however different from the ones

obtained by *Walter and Troll (2003)*. Another difference in our experiments is that the branch perpendicular to the rift axis is here rather more developed. Note that *Walter and Troll (2003)* used a conical gelatine model and it is thus difficult to compare their models with the ones produced during this study as the stress field is different within a cone than within a ridge. Furthermore, our models represent a long-lived system with the growth of a basal intrusive complex, responsible of long-term deformation and not punctuated system associated with short-term deformation, related to dyke emplacement.

The structures and intrusion shapes obtained using sand-plaster mixed models intruded with Golden Syrup may also be different from the ones obtained by *Fiske and Jackson (1972)* and *Walter and Troll (2003)* due to the different material used. Gelatine is an elastic material while sand and plaster are brittle. The intrusion rising through the various models will thus break the material differently. Ink intrusion within gelatine models (elastic behaviour) will create hydraulic fractures and thus form dykes. In contrast, in the sand-plaster models the Golden syrup can not create hydraulic fractures within the brittle material (note that however small extremities can be interpreted as dykes of small extent). During inflation and deflation periods, the Hawaiian volcano edifices show an elastic behaviour (*Fiske and Kinoshita, 1969*). The volcanoes deform elastically also before fracturing and prior to dyke emplacement (*Rubin, 1995*). Consequently, for dyke emplacement studies, gelatine models are appropriate, while for fault and collapse studies associated with basal complex emplacement (brittle deformation) sand-plaster models are preferred. Also, dyke (ink) and intrusive complex (Golden Syrup) rheologies are different. A dyke, being less viscous, will propagate as a blade-like intrusion while the intrusive complex, that is probably more viscous, will form a massive body. This is the reason why we form diffuse rift-zones rather than well-defined rift-zones in our experiments. Once the instability and the sector creep is triggered by viscous intrusive complex emplacement (type B & C models), dyke propagation will depend on the topography, but also on the stress field induced by the previously emplaced intrusive complex. Note that we consider, in that case, that no regional stress field acts on the system.

3.2. Model implications

The field data (see **chapter 1**) show that the north east rift-zone is rather diffuse and not well-defined, in contrast to the various models and general accepted concepts of rifts (*Fiske and Jackson, 1972; Dieterich, 1988; Ancochea et al., 1990; Carracedo, 1994; Walter and Troll, 2003; Walter et al., 2006*). The different sand-plaster models reproduce this character well, where the intrusive complex and its associated deformation is localised, i.e. the syrup intrudes only one sector of the flank or the ridge summit and does not propagate parallel but rather perpendicularly to the ridge axis. The sand-plaster models with a basal ductile layer (type A-models) allow the whole ridge to spread with, however, more deformation above the intrusion path. For high models, the intrusive complex sometimes propagates parallel to the ridge axis, but most of the Golden Syrup propagates perpendicular to the ridge axis. For the other sand and plaster models (type B and C), the deformation is localized and restricted to the immediate surroundings of the creeping block. The resulting stress field, after basal complex emplacement, will be modified. In consequence, the orientation of later dyke intrusion rising from the basal complex will be influenced by the new stress field. This phenomenon can also explain the diffuse character of the rift zone observed in the field. In the models studied here, only one injection was performed, leading to the formation of one collapse. It can be deduced that several basal intrusive complex emplacement episodes will complicate the rift configuration, favouring the geologically undefined character of the rift, as well as trigger several collapses.

Flank creeping seems to be the direct answer for the ridge to intrusive complex emplacement (flank push). As previously discussed in field and paleomagnetism sections (**chapter 1 & 2**), flank creeping is a process observed on oceanic islands that is thought to accommodate intense dyke intrusion and basal complex emplacement. It has been observed in Hawaii (*Swanson et al., 1976; Dieterich, 1988; Cayol et al., 2000; Brooks et al., 2008*) and in La Réunion Island (*Merle and Lénat, 2003*). Flank creeping can be encouraged by basal weak layers such as sediments, hydrothermal layers, breccia, cumulates, or even by the intrusive complex itself (*Borgia, 1994; Clague and Denlinger, 1994; van Wyk de Vries and Francis, 1997; Merle and Lénat, 2003; Leslie et al., 2004; Cecchi et al., 2005; Oehler et al., 2005; Márquez et al., 2008*). In our experiments, flank

creeping was triggered by a basal ductile layer (type A-models), by the intrusive complex (type B and C models) or both (type A-models with intrusion). The main observed features are compressive folds at the front of the sliding area (models A, B and C), en-échelon fractures along the edges (type B-models) and sub-vertical fractures within the creeping zone (type B and C models). In the field, no compressive structures have been observed. As the ridge continues under the sea, it is conceivable that we may find such structures under water. *Watts and Masson (2001)* identified “boulder ridges” that may be interpreted as compressive structures due to the Icod debris avalanche emplacement. These ridges may also correspond to the compressive folds formed at the front of the creeping area. On the other side, if such structures exist, they may have been buried underneath several debris avalanches and turbidite deposits. En-échelon fractures linked to strike-slip movements have been deduced from dyke orientation patterns in the field. Such en-échelon fractures have been previously observed in Madeira and were associated with strike-slip movements that accommodate flank creeping (*Klügel et al., 2005*). The B-models are thus consistent with the field observations and also the paleomagnetism results (see **chapter 1 and 2**). The sub-vertical fractures formed within the creeping area have also been observed on the field. In the C models, they are directly at the head of the sliding block as observed in La Orotava where the lava units are cut by several vertical to sub-vertical dipping faults and fractures. The typical horse-shoe shape of the creeping sector is also reproduced in the type B and C models. The different set ups (A, B & C) represent end-members and the natural cases are probably a combination of these three extremities. Note that during creeping, the flank is displaced as a coherent block (*Le Corvec and Walter, 2009*) until a certain limit, where the flank breaks-up and collapses. A clockwise vertical axis rotation of the upper units of the rift zone has been deduced from the paleomagnetic data (**chapter 2**). The deformation fields obtained from the analogue models confirm that rotation of the sliding block is possible (see **Fig.12&16**). The other possibility to explain the discrepancy between the expected magnetic field direction and the magnetic field direction obtained with the NERZ samples was a 28° WNW side down tilt along a N009 horizontal tilt axis. A major tilt of the head of the sliding block was observed in the type B and C models. Note that during sliding events, the tilt is supposed to be inwards, but as the intrusive complex pushes, an outward tilt is

favoured (**Fig. 17**). It is thus possible that a tilt occurred also at the head of La Orotava and Guimar collapses as well as a combination between a tilt and a rotation. However, as previously mentioned, no field evidence demonstrating a tilt have been found (**chapter 1 & 2**).

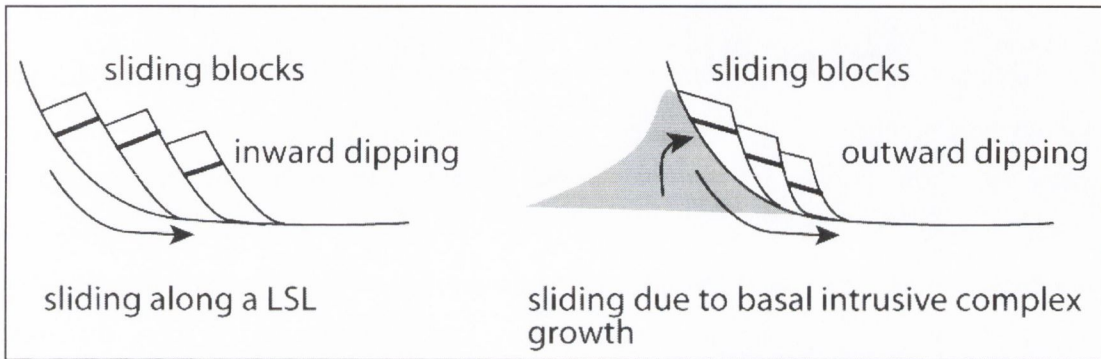


Figure 17: *Cartoon illustrating the expected tilt orientation occurring during a slide event and the orientation of the expected tilt occurring during the emplacement of an intrusive basal complex.*

A question arises when looking at the saucer-shape of several intrusive complexes formed following the experiments. Sometimes, the basal extension of the intrusive complex erupted at the base of the ridge. It is possible that few volcanic edifices may have developed from the basal lateral extension of the intrusive complex - obviously after the collapse event- at the base of Tenerife similarly to our experiments. However, no studies report any cinder cones at the base of the island (*Watts and Masson, 1995; 2001*). Note that they could have been buried by the several debris avalanche deposits.

4. Conclusion

The emplacement of an intrusive complex favours flank creeping under certain geometric conditions. For a circular point of injection, flank creeping occurred when the ridge is not too high and when the ridge axis is offset from the injection point. For an elongated point of injection, flank creeping is favoured when the ridge is high. If there is a basal ductile layer, the deformation is not localised around the injection point and the whole ridge will spread outwards. However, the most intense deformation occurred above the intrusion propagation path.

Flank creeping is characterized by compressive folds at the front of the sliding area, en-échelon fractures along the edges and sub-vertical fractures within the creeping zone. The structural features observed in the field –except folds that have not been observed- have been successfully reproduced using analogue models.

A tilt and a rotation of the creeping area have been observed within the models. The rotation inferred from the combination of paleomagnetic and field data is shown to be realistic by the models.

The models are characterized by a localised deformation due to the basal intrusive complex emplacement. This complex does not extend widely along the ridge axis and rather propagates perpendicularly to the ridge, forming a scattered rift zone. The NERZ is thus probably not a simple one-stage rift zone and is composed by more than one intrusive complex, i.e by several intrusive batch events, consistent with the field and paleomagnetic conclusions (**chapter 1&2**). The emplacement of an intrusive complex modifies the stress field that will, in turn, influence the orientation of later dyke intrusions reinforcing the “scattered” character of the rift zone.

The emplacement of an intrusive complex and associated dykes is responsible for flank creeping. The rift accommodates the deformation with en-échelon and sub-vertical fractures. At some stage, however, the rift cannot accommodate any further intrusions and becomes unstable. Further intrusive events will then trigger collapse.

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APPENDIX: list of experiments and Π numbers

The list of the parameters of the experiments and Π numbers are displayed below. The length dimensions are in cm but the intrusive flux is given in $\text{m}^3.\text{s}^{-1}$. For the set up B - as mentioned in the text – the intrusion was injected from the base or from a tube slightly sticking out from the rigid table.

NAME		height (h)	1/2 width (wr)	1/2 width (ws)	length (l)	angle	proportion sand/plaster	brittle sub.thick.(tb)	ductile sub.thick. (td)	frame dimension	intrusive flux (f)	hole/slit dimension	hole/slit position
SET UP A	ridge30_1	4.3	8.5	9.5	30	30	90/10	0.6	0.6	19*30	NA	NA	NA
	ridge30_2	4.8	9	12.5	30	30	90/10	0.6	0.6	25*30	NA	NA	NA
	ridge30_3	5.4	10	15	30	30	90/10	0.7	0.7	30*38.5	NA	NA	NA
	ridge30_4	5.1	8.5	9.5	53.5	30	90/10	0.4	0.4	19*53.5	NA	NA	NA
	ridge30_5	4.7	8	9.5	30	30	90/10	0.4	0.4	19*30	NA	NA	NA
	ridge30_6	4.7	10	11	37	30	90/10	0	3.8	22*37	NA	NA	NA
	ridge30_7	5	9.5	14	45	30	90/10	0	1.9	28*45	NA	NA	NA
	ridge30_9	4.8	8	9.5	30	30	90/10	0.6	0.6	19*30	NA	NA	NA
	ridge30_10	5.2	7.5	12.5	30	30	90/10	0.9	0.9	25*30	NA	NA	NA
	ridge30_11	5	6.5	13	35	30	90/10	0.5	0.5	26*35	NA	NA	NA
	ridge30_12	4	7	12	34	30	90/10	1	0.5	24*34	NA	NA	NA
	ridge30_13	3.8	7	13	37	30	90/10	0.6	0.6	26*37	NA	NA	NA
	ridge30_14	5.1	7.75	15	46	30	90/10	0.7	0.7	30*46	NA	NA	NA
	ridge30_15	4	8	12	40	30	90/10	0	3	24*40	NA	NA	NA
	ridge30_16	3	6.5	12	40	30	90/10	0	3	24*40	NA	NA	NA
	ridge30_17	3.6	7	20	31	30	90/10	0.6	0.6	31*40	NA	NA	NA
	ridge30_18	5	9.5	15	38	30/20	90/10	0.5	0.5	30*38	NA	NA	NA
	ridge30_19	7	10	19.5	28	30/20	90/10	0.5	0.5	28*39	NA	NA	NA
	ridge_in30_1	5	9.5	9.5	30	30	90/10	0.5	0.5	19*30	NA	2.7	centre/base
	ridge_in30_2	4.8	9.5	9.5	30	30	90/10	0.6	0.5	19*30	NA	2.7	centre/base
SET UP B	ridge_in30_4	4.5	6.75	14	30	30	90/10	0.5	0.5	28*30	1.68E-08	2.7	centre/base
	ridge_in30_5	4.5	6.5	18.5	30	30	90/10	0.5	0.5	30*37	3.14E-08	2.7	centre/base
	ridge_in30_6	4.5	7.2	13.5	42	30	90/10	0.5	0.5	27*42	3.18E-08	2.7	centre/base
	ridge_in30_7	4.2	6.5	13	36	30	90/10	0.4	0.4	26*36	2.01E-08	2.7	centre/base
	ridge_in30_8	6	12	13.5	30	30	90/10	0.9	0.9	27*30	3.57E-08	2.7	centre/base
	RPA1	5	9	NA	NA	30	50/50	NA	NA	NA	NA	1.7	flank/tube
	RPA2	5	10.5	NA	NA	30	50/50	NA	NA	NA	NA	1.7	flank/tube
	RPA3	4	10	NA	NA	30	50/50	NA	NA	NA	1.62E-06	1.7	centre/tube
	RPA4	4	11.5	NA	NA	20	50/50	NA	NA	NA	2.84E-06	1.7	centre/tube

SET UP C	RPB1	8	14	NA	NA	30	50/50	NA	NA	NA	9.86E-07	1.7	flank/tube
	RPB2	6.5	16	NA	NA	20	50/50	NA	NA	NA	1.79E-06	1.7	centre/tube
	RPB3	6	17	NA	NA	20	50/50	NA	NA	NA	2.27E-06	1.7	centre/tube
	RPC1	5	10	NA	NA	30	50/50	NA	NA	NA	1.49E-06	1	flank/base
	RPC2	5	14	NA	NA	20	50/50	NA	NA	NA	1.06E-06	1	centre/base
	RPC3	3	14.5	NA	NA	10	50/50	NA	NA	NA	1.01E-06	1	centre/base
	RPC4	6	10.5	NA	NA	30	50/50	NA	NA	NA	NA	1	centre/base
	RPD1	5	10.5	NA	NA	30	50/50	NA	NA	NA	9.67E-09	2.7	centre/base
	RPD2	4.2	11.5	NA	NA	20	50/50	NA	NA	NA	2.09E-08	2.7	flank/base
	RPD3	6.5	15	NA	NA	30	50/50	NA	NA	NA	1.15E-08	2.7	flank/base
	RPD4	6	10.5	NA	NA	30	50/50	NA	NA	NA	NA	2.7	flank/base
	RPD5	4	10	NA	NA	20	50/50	NA	NA	NA	NA	2.7	flank/base
	RPD6	3.2	11	NA	NA	20	42/58	NA	NA	NA	NA	2.7	flank/base
	RF6	5.5	18	NA	27\39	30	42/58	NA	NA	NA	9.31E-08	18*0.7	centre
	RF7	6	20	NA	21\41	30	42/58	NA	NA	NA	1.06E-07	18*0.7	centre
	RF8	5.5	18	NA	19\29	30	42/58	NA	NA	NA	6.70E-08	15.7*1.4	centre
	RF9	5	24	NA	30\42	20	42/58	NA	NA	NA	1.13E-07	18*0.7	centre
	RF10	3.5	21	NA	32\42	20	42/58	NA	NA	NA	5.98E-08	18*0.7	centre
	RF11	2.5	19	NA	20\29	10	42/58	NA	NA	NA	1.05E-07	15.7*1.4	centre
	RF12	3	23	NA	35	12	42/58	NA	NA	NA	8.38E-08	18*0.7	centre
	RF13	3.7	16	NA	21	30	42/58	NA	NA	NA	5.24E-08	15.7*1.4	flank
	RF14	4	19	NA	28	20	42/58	NA	NA	NA	1.68E-07	18*0.7	flank
	RF15	2.3	20	NA	30	10	42/58	NA	NA	NA	8.38E-08	18*0.7	flank
	RF16	6	20	NA	14\30	30	42/58	NA	NA	NA	8.38E-08	15.7*1.4	centre
	RF17	5	9\10	NA	19\30	30\20	42/58	NA	NA	NA	9.57E-08	15.7*1.4	centre
	RF18	7	22.5	NA	26\41	30	42/58	NA	NA	NA	7.96E-08	18*0.7	centre
	RF19	5.5	9.5\7	NA	17\27	30\40	42/58	NA	NA	NA	9.57E-08	15.7*1.4	centre

	Π1	Π2	Π3	Π4
ridge30_1	0.506	10.17	0.895	1
ridge30_2	0.533	11.00	0.720	1
ridge30_3	0.540	10.71	0.667	1
ridge30_4	0.600	15.75	0.895	1
ridge30_5	0.588	14.75	0.842	1
ridge30_6	0.470	2.24	0.897	NA
ridge30_7	0.526	3.63	0.667	NA
ridge30_9	0.600	11.00	0.842	1
ridge30_10	0.693	8.78	0.600	1
ridge30_11	0.769	13.00	0.500	1
ridge30_12	0.571	13.00	0.583	2
ridge30_13	0.543	9.33	0.538	1
ridge30_14	0.658	10.29	0.517	1
ridge30_15	0.500	2.33	0.667	NA
ridge30_16	0.462	2.00	0.542	NA
ridge30_17	0.514	9.00	0.350	1
ridge30_18	NA	13.00	0.633	1
ridge30_19	NA	17.00	0.513	1
ridge in 30_1	0.526	13.00	1	1
ridge in 30_2	0.505	11.00	1	1
ridge in 30_4	0.667	12.00	0.482	1
ridge in 30_5	0.692	12.00	0.351	1
ridge in 30_6	0.625	12.00	0.533	1
ridge in 30_7	0.646	13.50	0.500	1
ridge in 30_8	0.500	9.67	0.889	1
RPA1	0.556	NA	NA	NA
RPA2	0.476	NA	NA	NA
RPA3	0.400	NA	NA	NA
RPA4	0.348	NA	NA	NA
RPB1	0.571	NA	NA	NA
RPB2	0.406	NA	NA	NA
RPB3	0.353	NA	NA	NA
RPC1	0.500	NA	NA	NA
RPC2	0.357	NA	NA	NA
RPC3	0.207	NA	NA	NA
RPC4	0.571	NA	NA	NA
RPD1	0.476	NA	NA	NA
RPD2	0.365	NA	NA	NA
RPD3	0.433	NA	NA	NA
RPD4	0.571	NA	NA	NA
RPD5	0.400	NA	NA	NA
RPD6	0.291	NA	NA	NA
RF6	0.611	NA	NA	NA
RF7	0.600	NA	NA	NA

RF8	0.611	NA	NA	NA
RF9	0.417	NA	NA	NA
RF10	0.333	NA	NA	NA
RF11	0.263	NA	NA	NA
RF12	0.261	NA	NA	NA
RF13	0.463	NA	NA	NA
RF14	0.421	NA	NA	NA
RF15	0.230	NA	NA	NA
RF16	0.600	NA	NA	NA
RF17	NA	NA	NA	NA
RF18	0.622	NA	NA	NA
RF19	NA	NA	NA	NA

CONCLUSION

Tenerife, Canary Islands, is constructed of three rift-zones organized in a triple-armed pattern that enclose major collapse scars. Using a multi-disciplinary approach, we attempted to better understand the mechanism(s) of the formation and evolution of the North-East Rift-Zone on Tenerife, and its intimate relationship with lateral collapses. The different methodologies bring complementary information that allow us to derive an overall model of rift zone evolution (**see figure**).

Initial rift development may be largely influenced by the regional stress field or by pre-existing structures. With proceeding eruptions, however, the developing intrusive complex(es) and the growing ridge take control on the near-surface stress field. Once the ridge has reached a critical height, depending largely on shape, dynamics and position of the intrusive complex(es) below the rift, the flanks will start to creep to accommodate the growth of the intrusive complex and the emplacement of dykes and sills near the surface. This creep mechanism is characterised by a slow displacement of the flank along a basal décollement and lateral strike-slip faulting, as indicated by dyke orientation patterns in the field, paleomagnetic data of dykes, and by the analogue models presented. Flank creep has to occur along a basal décollement, such as breccias, or above an intrusive complex likely sited within the edifice.

Incoming magma into the rift is thus accommodated by flank creep, but brings the rift gradually (i.e. with each new magma batch) towards a terminal unstable state. If the intrusive activity stops, or slows down, the rift will reach a stable position again by creeping. However, if new or continued intrusions are emplaced, the flank will collapse. This episodic intrusion-creeping-collapse generates a positive feedback as the destabilization of the ridge flanks creates fractures and space for eventual incoming new magma batch.

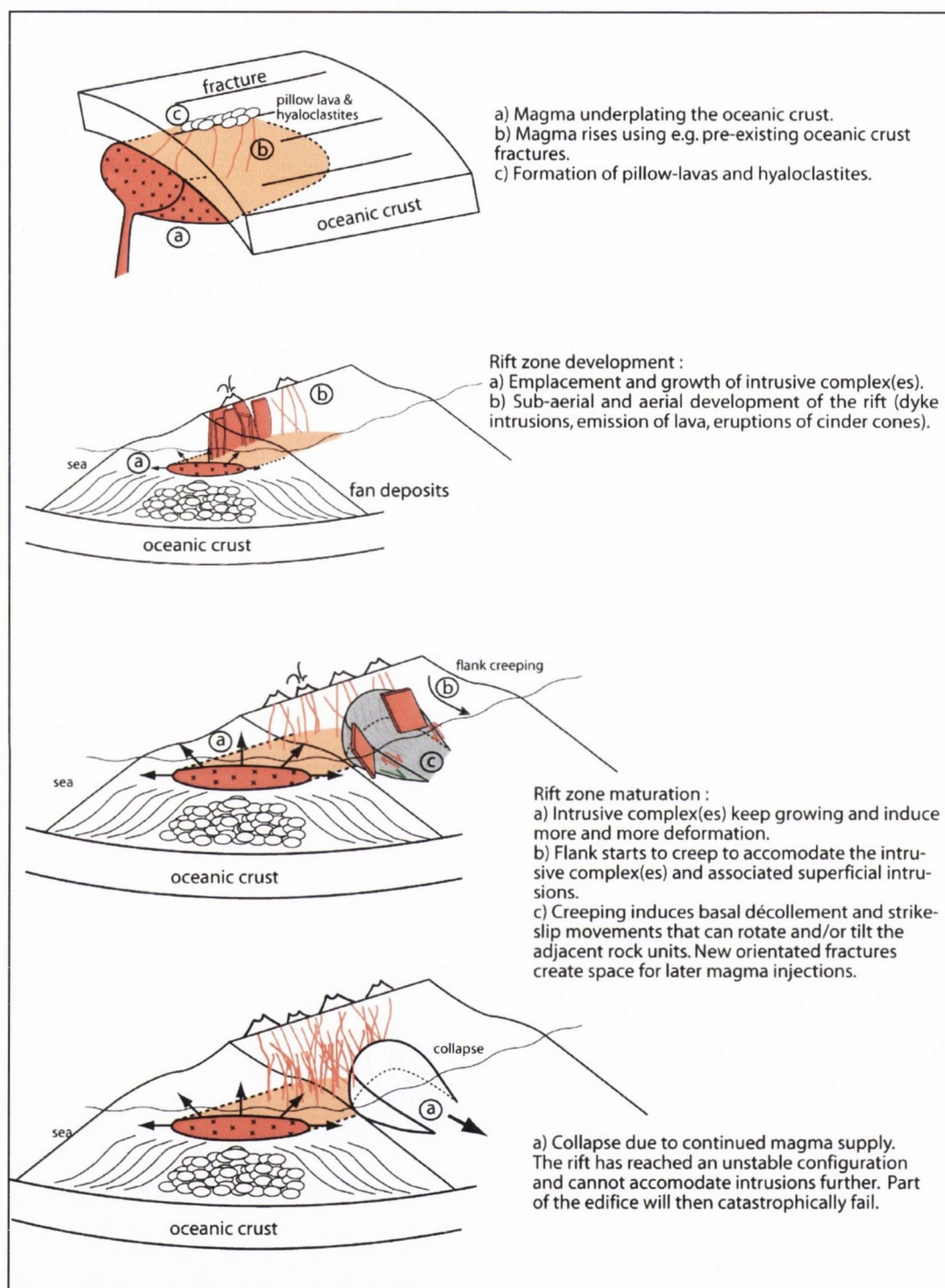


Figure: Model of the development and growth of the North-East Rift-Zone that is largely controlled by the intrusive complex emplacement and its associated dykes. The model is divided in four main stages: (1) submarine stage with formation of pillow-lavas and

hyaloclastites; (2) rift-zone subaerial and aerial development stage with the growth of intrusive complex(es), the emplacement of dykes and lava flows, and the eruption of cinder cones; (3) rift-zone maturation stage where the mechanism of flank creeping is initiated; (4) final stage with catastrophic lateral collapse.

This evolutionary sequence contrasts the widely held idea that rift-zones are generally considered as narrow and well-defined linear structures, with a consistent dyke and fracture orientation trend. The different approaches used here show that a rift-zone can be rather diffuse and probably composed by several events, with several structural complexities. The emplacement and growth of intrusive complex(es) affect the stress field that, in turn, controls the orientation of new dyke injection, and thus promotes the diffuse character of the rift-zone with its lifespan (structural maturation). The North-East Rift-Zone represents a long-lived and dynamic system with the development of various successive geological and geo-morphological environments, largely linked to the responses of the ridge to the emplacement and growth of the intrusive complex(es) and its associated dyke intrusions.