WORKSHOP

"Caldera Volcanism: Analysis, Modelling and Response"

Parador de las Cañadas
Tenerife
Spain
16 - 22 October, 2005

Organisers:
Joan Martí
Joachim Gottsmann
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SESSION: GEOL

Geological, petrological, geochemical, stratigraphic and physical investigations of lithologies and deposits related to caldera forming eruptions
The Sierra Madre Occidental (SMO) is the largest continuous ignimbrite province in the world. It covers the NW portion of Mexico and has a minimum estimated rock volume of about 360,000 km$^3$. The southern part of the Basin and Range extensional province is in Mexico, and formed NW- to NE-trending normal faults that bound many large grabens, which are particularly long and deep in the southern SMO. Both graben formation and ignimbrite activity coincided in space and time, particularly for the 38-23 Ma period, which has been referred to as the *Ignimbrite Flare-up* event. Geologic observations at the southern SMO indicate that the vents of several large volume silicic ignimbrites were related with the graben's master faults, so these structures can be defined as volcano-tectonic depressions, or graben-calderas. The evidences include large pyroclastic dikes, co-ignimbrite lithic breccias, and post-ignimbrite aligned rhyolitic domes and rhyolitic lava dikes. All these features were found along the graben-caldera walls or on the graben's shoulders. Apparently, ignimbrite-forming eruptions occurred episodically with graben collapse through several vents along the graben-caldera walls. The downdropped blocks inside the graben-caldera have several distinct tilting directions, in many cases opposite to those caused by regular domino faulting (with tilting outwards from the graben’s axis), indicating a chaotic collapse of blocks. This is interpreted as blocks that collapsed during magma evacuation from the sub-graben magma chamber, similar to a piece-meal caldera collapse. Generally, the pyroclastic dikes occur as discontinuous elongated lenses rather than regular tabular bodies. Commonly, an elongated lava dome filled the vent and totally covered or destroyed the pre-existent pyroclastic dike, but in some cases the pyroclastic dike was preserved. Some of these vents are related to gold and silver hydrothermal mineralization. Faulting and subsidence continued for several millions of years after the ignimbrite emplacement, displacing the graben-caldera products downward into the tectonic depression, but preserving the chaotic arrangement of the collapsed blocks. Lacustrine deposits indicate that in most cases a paleo-lake filled the graben-caldera for a period of time, either during the ignimbrite emplacement or after it. Some of the graben-caldera ignimbrites were deposited in subaqueous environments.
USING THE STRATIGRAPHIC RECORD TO UNDERSTAND THE NATURE OF CALDERA COLLAPSE: THE 1.59 -PRESENT LAS CAÑADAS CALDERA COMPLEX, TENERIFE, SPAIN

GEOL

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The basaltic oceanic island shield volcano of Tenerife (12 Ma - present), Canary Islands, developed a phonolitic magma chamber about 4 Ma, which formed the central Las Cañadas edifice. The early history of this edifice is poorly known, but the last 2 Myr was marked by three major explosive eruption cycles, each caldera forming. The Ucanca Formation cycle (1.59 – 1.07 Ma) is relatively unknown. The Guajara Formation cycle (0.85 – 0.57 Ma) consists of the Arico Member (0.67 Ma widespread welded to non-welded ignimbrite, >2 km³), the Abades Member (~0.60 Ma, three ignimbrites and a fallout unit up to 4 m thick, (>1.6 km³), and the Granadilla Member (0.57 Ma, a fallout unit up to 10 m thick and a major ignimbrite up to 15 m thick; > 10 km³). The major eruption units in the Diego Hernández Formation cycle (0.37 – 0.17 Ma) are the Aldea Member (0.32 Ma, fallout and ignimbrites, >3 km³), the Fasnia Member (0.31 Ma, multiple fallout, flow and surge deposits, >13.5 km³), the Poris Member (0.27 Ma, multiple fallout, flow and surge deposits, > 2.7 km³) and the climactic Abrigo Member (0.188 Ma, multiple lithic rich ignimbrites, > 1.8 km³). Major eruptions occurred ~ every 30 ka, and lesser eruptions ~ every 2 ka. The last known phonolitic explosive event (Montaña Blanca, 2 ka) may herald a new explosive cycle.

Fellout deposit isopach maps and ignimbrite outcrop patterns and thicknesses, indicate that depocentres for major eruption units lay offshore, increasing known erupted volumes by an order of magnitude. Ignimbrites contain up to 50% accessory lithic clasts, indicating major vent collapse while large volumes of magma were erupted from phonolitic magma chambers at depths of 5 - 8 kms.

The Las Cañadas Caldera Complex consists of three coalesced calderas (Ucanca, Guajara and Diego Hernández), each about 6 – 8 kms in diameter. Each eruption unit could have produced a small caldera collapse but we believe that each major eruption cycle probably produced multiple or incremental stages.
of caldera collapse, which in aggregate produced the three coalesced calderas we see today. This model is consistent with the proposed existence of multiple magma chambers during the life of the Las Cañadas Caldera Complex.
The Guajara Formation, dated from 0.85 Ma to 0.57 Ma, is the second of three major explosive phonolitic phases of eruptive activity from the Las Cañadas Caldera system on Tenerife. The Guajara Formation includes a series of plinian fall deposits, ignimbrites, surge and ash fall units. Initial activity probably originated from fissure vents close to or along the caldera wall producing numerous welded, near-vent fall deposits. This early fall forming activity also produced more than 20 non-welded fall deposits away from vent on the Bandas del Sur, the southwest coastal plain of Tenerife. Eruptive activity then focused on one or two vents producing small volume ignimbrites (<1 km$^3$) with associated fall deposits of the Rio and Eras Members. Approximately 0.67 Ma ago, the eruption conditions changed again with the deposition of the Helecho Member, a lithic rich ignimbrite preserved locally on the Bandas del Sur. It appears to represent widening of the vent and it was probably the first erupted product related to a partial or incremental caldera collapse event during the Guajara phase. The ignimbrites deposited after this event (those within the Arico (>2 km$^3$ at 0.668 ± 0.004 Ma), Abades (>1.6 km$^3$ at 0.596 ± 0.014 Ma) and Granadilla (>10 km$^3$ at 0.569 ± 0.014 Ma) Members) are at least an order of magnitude greater in volume and more widespread with offshore depocentres. The Abades and Granadilla ignimbrites also display lithic rich concentration zones towards their tops, while in the Arico ignimbrite, a lithic rich zone appears directly beneath the top flow unit in the ignimbrite which is welded. Similar to the abundant vent derived lithic clasts within the Helecho ignimbrite, these lithic rich horizons may also imply partial/incremental caldera collapse during the eruptions of the Arico, Abades and Granadilla ignimbrites. The combination of these eruptions produced the Guajara sector of the nested Las Cañadas Caldera complex.

Over a 200,000 year period, only seven significant eruptions occurred which, in today’s world, would have resulted in severe damage to buildings and infrastructure and loss of life. These eruptions may have occurred on average every 28,500 years with each eruption gaining in strength and erupted volume as the volcano progressed from producing fall deposits from small fissure vents to intermediate volume ignimbrites from a collapsing caldera. Each new eruption was most likely initiated by a basaltic intrusion into the base of an evolved phonolitic magma chamber. The tephriphonolitic to phonolitic geochemical signature of the Guajara products suggests the presence of at least two magma chambers beneath the caldera that were involved in these eruptions. The eruption of each fall deposit and/or ignimbrite may have lasted tens, hundreds or even thousands of years, after which repose periods of perhaps hundreds to thousands of years must have occurred to allow soils to develop and new magma to refill the chambers and fractionate. Unfortunately, there is no point of reference to establish the timing and duration of these events as no large scale eruptions have occurred on Tenerife during recorded history. This makes modelling any future behaviour complicated.

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The Fasnia Member of the Diego Hernandez Formation, Tenerife, Spain, and the Bandelier Tuff, USA, are two examples of deposits related to caldera formation at Las Cañadas and Valles calderas, respectively. Traditional geochemical investigations have shown that multiple magmatic components are present in each system, and that the geochemical trends are dominated by mixing arrays. New LA-ICPMS microsampling within individual pumice clasts, however, have revealed fine scale incompatible trace element variations that differ significantly from the dominant whole-rock mixing trends observed in earlier datasets, and which are controlled by the behavior of late-crystallizing trace phases (titanite in the Fasnia; zircon and REE phases in the Bandelier Tuff). These mm-scale variations are not seen at the length scale of a whole pumice clast, and hence are interpreted to have been small-scale features in the magma chamber, consistent with extraction of liquid from a crystal pile.

The observed intraclast variability may be explained by thermal rejuvenation of shallow crystal-liquid mush or plutonic rock by intrusion of mafic liquid. Resultant temperature increases cause late-crystallized trace phases to be resorbed into the liquid to varying degrees, creating compositional variations that are distinct from whole-pumice data. Potentially, geochemical evidence for such a process could be obscured by subsequent magmatic processes unless the magma is rapidly erupted. Note that whole-rock analysis, even of single pumice clasts, was not sufficient to detect these variations.

The Fasnia Member is a mixture of two phonolitic and at least one mafic liquids. The less-evolved felsic endmember shows variations expected for dissolution of titanite. Similar relations are seen in the minimum-melt rhyolite of the Bandelier Tuff, where the pattern of variations in Zr, Hf, and REE is consistent with resorption of late-grown zircon + allanite ± chevkinite ± britholite. The similar evidence for heating from two very different sub-caldera minimum-melt magma bodies, one oceanic, moderate volume (~10 km³) and silica-undersaturated, and the other continental, large (~400 km³) and rhyolitic, suggests that thermal rejuvenation of multiply-saturated magmas at an advanced stage of crystallization, perhaps in the presence of an exsolved fluid phase, may be an important mechanism causing explosive eruption of highly differentiated magmas at calderas.
Spatial component variations within ignimbrite deposits can provide insight into dynamic caldera-forming eruptive processes. The 188 ka moderate volume (>>1.8 km³), lithic-rich (up to 40-50% of the deposit) phonolitic Abrigo Ignimbrite represents the last major explosive caldera-forming eruption from the Las Cañadas caldera complex on Tenerife. Apart from a rare, possible Abrigo plinian fall deposit, locally underlying the ignimbrite, the deposit consists of multiple depositional units of lithic-rich phonolitic ignimbrite exposed radially around the lower flanks of the Las Cañadas edifice. A correlative clast-supported lithic breccia occurs on the eastern caldera wall. Two major depositional units, separated by a fine ash layer, can be correlated across the southern coastal plain. A vent-derived lithic block concentration zone occurs within the upper part of the lower unit and represents a major partial collapse event during the eruption.

Qualitative and quantitative component studies, supported by juvenile clast geochemistry, within the stratigraphic framework of the Abrigo Ignimbrite were undertaken to establish spatial and temporal variations in magma chamber-vent processes. Accessory lithic clasts within the Abrigo Ignimbrite consist of basaltic to phonolitic coherent glassy to crystalline volcanic and equivalent hydrothermally altered clasts, and minor altered nepheline syenite (rare gabbroic clasts), welded or lava-like volcanic breccias and pyroclastic and epiclastic breccias. Juvenile glassy clasts, although predominantly highly vesicular phonolitic pumice, display variations in composition (basanitic to phonolitic), vesicularity (incipient to high) and crystallinity (near-aphyric to >60% crystals). Fresh juvenile nepheline syenite is also common.

Accessory lithic clast abundances suggest that soft pervasive hydrothermally altered roof rocks and a possible pre-Abrigo stratovolcanic complex existed above the Abrigo magma chamber. The wide diversity in juvenile clast types within the Abrigo Ignimbrite indicate that almost all parts of the magma chamber were erupted, facilitating/driven by caldera collapse. Spatial variations in the proportions of accessory lithic and juvenile clasts imply that the eruption occurred from multiple vents tapping a laterally variable subsurface geology and heterogenous magma chamber.

Major caldera-forming eruptions on Tenerife, although not common, modify large-scale intra-caldera landforms, produce widespread hazardous pyroclastic flows which destroy and bury the extracaldera landscape and generate tsunamis when they enter the sea. Hazard mitigation programs must account for caldera-forming eruptions in the volcanic history of the island.
CALDERA COLLAPSE AND CALDERA SYSTEM EVOLUTION

(1) Generation of source- and eruption-controlled flow units (FUs) is governed by intrinsic (magma reservoir) and extrinsic (crustal lithology, hydrology etc) controls and emplacement dynamics during single eruptions.

(a) Internal forcing: FUs contrasting in composition and degree of mixing reflect repetitive roof collapse above compositionally zoned reservoirs. Boundary conditions: physical parameters (density, viscosity, volatiles); chemical zonations and gradients in the reservoir; structural dynamic parameters (conduit collapse events and multiple roof subsidence). Magma mixing occurs in the reservoir/conduit system between eruptions of FUs possibly governed by repeated roof subsidence. Decrease in lithostatic pressure lowers the fragmentation level allowing deeper more mafic magma portions to be emptied explosively.

(b) External forcing: Eruption-controlled FUs in ignimbrites lacking compositional zonations are commonly governed by pervasive magma-water interaction caused by eruptions through caldera lakes. The step-wise evacuation of magma reservoirs during single eruptions has major implications for hazard assessment of voluminous caldera eruptions.

(2) Caldera system evolution
The growth and decay of a large caldera-ignimbrite system over ca. 5 million years is illustrated by the Gran Canaria Tejeda caldera system. Its longevity was sustained thermally and materially by mafic magmas, the melting anomaly changing in composition and decreasing in magma production rate.

(3) Where do mafic parent magmas of large alkalic caldera-ignimbrite systems hide?
Emplacement of ca. 30 felsic cooling units during ca. 5 million years on Gran Canaria was interrupted only twice by widespread evolved alkali basalts at 13.6 Ma and much younger local nephelinite, reflecting compositional changes in parent magma and source. Rare fresh sideromelane shards in felsic syn-ignimbrite turbidites widespread around Gran Canaria are likely derived from nearly coeval submarine hyaloclastites eroded by turbidites. The compositional change of sideromelane from tholeiitic to alkalic is paralleled by compositional changes from subalkalic to highly peralkaline in the rhyolite parent ignimbrites and derivative turbidites. Dense mafic magma is unable to penetrate thick low-density evolved magma reservoirs but can escape sideways.
STRATIGRAPHIC AND SEDIMENTOLOGICAL STUDIES WERE PERFORMED ON PYROCLASTIC DENSITY CURRENTS GENERATED DURING COMPLEX SUBPLINIAN ERUPTIONS: THE EXAMPLE OF THE AD 472 (POLLENA) ERUPTION OF SOMMA-VESUVIUS, ITALY

Session ID: GEOL

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Stratigraphic and sedimentological studies were performed on pyroclastic density currents generated during the AD 472 (Pollena) eruption of Somma-Vesuvius, Italy. Sedimentological data were interpreted in an innovative way, which considers together the models of granular flow motion, pyroclastic aggradation and en-mass deposition. The PDCs were generated by different eruptive mechanism, with explosive dynamic driven by both magmatic and phreatomagmatic fragmentation. Detailed lithofacies analysis shows that the studied deposits were emplaced under variable flow-boundary conditions, dominated by grain interaction, fluid escape or traction processes. Stratigraphic and lithofacies analysis suggests that the moving flows were density stratified with a basal underflow dominated by grain interaction. The underflows were organized as trains of self-organized granular flows of variable thickness and magnitude, depending on the contrasting effects of particle concentration and fluid turbulence. The change in slope between upper and lower slopes of the volcano promoted deposition and the different surges aggraded stepwise. Particle concentration, density, mean velocity, and flow height were assessed for the studied PDCs using two different methods for massive and stratified deposits. The calculated mobility of the flows is 0.2-0.3, in the range of small-scale PDCs. The variability of flow dynamics and depositional processes yield insights useful for the improvement of existing hazard mitigation plans in the circumvesuvian area.
The island of Terceira ( Açores, Portugal) is 400 km² in area and consists of four composite volcanoes with calderas (from youngest to oldest: Santa Barbara, Pico Alto, Guilherme Moniz, Cinco Picos) grouped along a basaltic fissure zone that transects the island from NW to SE. Three of the volcanoes are noteworthy for production of peralkaline felsic magmas, which have sometimes erupted explosively to produce ignimbrites and fall deposits. While little is known of the history of the older of these three (Guilherme Moniz), the two younger, active calderas appear to have formed incrementally: Pico Alto has had several ignimbrite-producing eruptions, with sub-plinian/lava dome-forming events occurring before, during, and after ignimbrite volcanism, and Santa Barbara has undergone numerous sub-plinian/lava dome-forming events. Ignimbrite eruptive volumes estimated from on-land deposits are small (~0.1-0.3 km³) but considerable amounts of material must have flowed into the sea.

The Lajes-Angra Ignimbrite, of comenditic trachyte composition and variously dated by 14C at 19-23 ka, is the youngest from Pico Alto caldera. This low-aspect-ratio ignimbrite is densely welded in places despite being only a few metres thick. A widespread ignimbrite veneer facies of this deposit attests that pyroclastic flows covered 2/3 of the island. At least five other ignimbrites have been identified so far in the sequences of deposits that underlie the Lajes-Angra deposit in sea cliffs around Terceira. It is not yet known whether these ignimbrites were all derived from Pico Alto, or whether any came from Guilherme Moniz caldera. New 40Ar/39Ar age determinations on anorthoclase crystals separated from ignimbrite pumice clasts, together with 14C ages for the younger units, support a rather narrow period of ignimbrite volcanism from 85 to 20 ka ago. The average time-interval between eruptions during this period is thus ~16 ka. The 20 ka elapsed since the most recent event suggests that ignimbrite volcanism from Pico Alto could still occur in the future, and may be a future threat to Terceira’s island-bound 60,000 population. Santa Barbara’s caldera is smaller (~2 km diameter) than Pico Alto’s (~4 km), and, based on stratigraphic and age relationships, this volcano could potentially enter an ignimbrite-forming phase.
OUTSTANDING ISSUES ABOUT RELATIONSHIPS BETWEEN LARGE-SCALE CALDERAS, IGNIMBRITE VOLUMES, AND MAGMA BODY SHAPE AND LONGEVITY

STRC

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Ideas on formation of large calderas by eruption of large-volume ignimbrites and relationships with subjacent magma bodies are rapidly changing. Rather than long-lived, blob-shaped magma chambers that are periodically tapped at the top, leaving a largely untapped, dominant volume of melt that then evolves until the next eruption, new ideas are being proposed. To test these ideas requires improved knowledge of caldera dimensions and form in relation to morphology and volume of the extracted magma body. This presentation examines some of these issues, based the 24 x 22 km-diameter Valles Caldera, New Mexico, USA, and two ignimbrites from this center, the upper (Tshirege) and lower (Otowi) Bandelier Tuffs, which were erupted ~ 500 ka apart at 1.2 and 1.7 Ma.

Estimating erupted volumes and relation to caldera size: Knowledge of products of major ignimbrite eruptions has shown that, in terms of volumes erupted, intra-caldera ignimbrite ≈ outflow ignimbrite sheet ≈ co-ignimbrite ash. Using the Bandelier – Valles Caldera examples, each ignimbrite-forming eruption yielded about 600 km$^3$ of magma and had estimated caldera collapse depths of 1.6 km (lower) and 2 km (upper). The latter is, interestingly, similar to the thickness of intra-caldera ignimbrite, which seems to be dominated by the upper Tuff.

Geometry of extracted magma volume: If the magma volume for each Bandelier ignimbrite was 600 km$^3$, by assuming that the area enclosed by the two caldera “ring fractures” reflects the area of the magma extraction zone, the thickness of extracted magma can be estimated in each case to be ~1.2 and 2 km. This suggests sill-like magma bodies, if a simple geometry is adopted. The magma volume was the major proportion of molten material available at the time of each eruption, suggested by the less-evolved composition and crystal-rich nature of late-extracted magma.

Evidence for transitory nature of erupted magma batches: Bandelier magmas were probably derived from melting of intrusions related earlier volcanic episodes. Evidence supports high silica rhyolite melt production and segregation just before the Otowi eruption. Other evidence indicates that a crystal pile left over from the Otowi eruption was rejuvenated to make the Tshirege magma. An intriguing possibility is that melt production and caldera formation may be linked?
If repeated today, phonolitic explosive eruptions such as occurred during the last complete magmatic cycle on Tenerife (the Diego Hernández Formation, DHF) would cause havoc to the island’s infrastructure and population and, at worst, could devastate the entire island. The growing recognition among petrologists that many of the disequilibrium microscale features of volcanic rocks are developed within a few years or decades prior to eruption leads us to hope that examination of the products of previous activity may yield valuable clues to the processes preceeding explosive venting of magma, and thus aid interpretations of medium – long term (1 – 10^2 years) eruption precursor symptoms that may be detectable at the surface.

DHF phonolitic magmas have near-uniform compositions very close to the 1 kbar water-saturated phonolitic minimum, but variations in trace element content require that at least two distinct phonolite types were independently and repeatedly generated throughout the 200 k.y. history of the DHF. Mixing with mafic magma has long been recognized as a potential eruption trigger and played a role in 5 of the 6 major DHF eruptions. New microanalyses of trace elements in glasses and evaluation of phenocryst-glass disequilibrium relationships using element partitioning theory reveals a more complex array of mixing and magma generation processes preceeding major phonolitic eruptions on Tenerife. In particular, variations within phonolitic glasses at the sub-mm scale are large and distinct from those produced by mingling with mafic liquid, while pyroxene phenocrysts are typically not in chemical equilibrium with their host glasses. These relationships require mixing between two phonolite melts in addition to the mafic component(s), and that generation of at least one phonolite type may be due to thermal rejuvenation of syenitic plutons and/or partly liquid “crystal mush” by mafic magma. We illustrate these relationships by reference to the 13 km^3, 309 ka Fasnia plinian + ignimbrite deposit, but they are also displayed by 3 other major DHF units including the climactic Abrigo ignimbrite. Similar trace element behaviour among minimum-melt rhyolitic glasses at Valles caldera, U.S.A., suggests that these processes may occur in a wide variety of highly differentiated, minimum-melt subcaldera magma bodies.
SESSION: MELT

Physicochemical properties of melts and magmas related to caldera volcanism
LI DIFFUSION IN PLAGIOCLASE PHENOCRYSTs OF THE NEA KAMENI DACITES: IS THE MAXIMUM RESIDENCE TIME OF PHENOCRYSTs REALLY THIS SHORT?

MELT

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Diffusion kinematics have been employed to determine residence times in magmatic reservoirs. The most recent and notable of these are attempts to calculate timespans using trace elements in plagioclase, a prevalent rock-forming mineral characterised by extremely slow CaAl-NaSi interchange. A caveat of this method is the selection of crystals that show primary trace element profiles, i.e. concentrations that have not yet been equilibrated by diffusion. As for the choice of trace element, one element known to diffuse rapidly, with low activation energies, is most ideal.

Within the past decade, modelling of Sr (Zellmer et. al, 1999) and Mg (Costa et. al, 2003) diffusion in plagioclase have been developed to ascertain maximum residence times for Nea Kameni dacites (Hellenic Arc) and gabbroic xenoliths of Volcan San Pedro dacitic lavas (Chilean Andes), respectively. This study is focused on the dacites of Nea Kameni as well, but using Li, owing to the much more rapid diffusion of this element in plagioclase (Giletti and Shanahan, 1997; La Tourette and Wasserburg, 1998).

Simple linear calculations on Li zonation profiles of plagioclase crystals yield residence times much shorter than previously derived or even expected. Plagioclase phenocrysts with accentuated Li profiles that imply rather unmodified concentrations, give maximum periods of ~4 hrs. It should be noted that these calculations involve conservative parameters, the lowest reasonable estimates of diffusivity: (1) a distance of 400 µm within the larger crystals, (2) a low temperature estimate of 800°C for the Nea Kameni dacites, and (3) an activation energy of 151 kJ/mol in anorthite, the composition where Li diffusivity is slowest in plagioclase.

Our results suggest that plagioclase crystallisation may have been driven by decompression and degassing during eruption, and that this process was undoubtedly fast. Although quite tentative, this technique may yet drastically change our knowledge of maximum magma residence times and their implication on volcanic hazards.

References:
DEALING WITH STRESS IN A CRISIS: LESSONS FROM MAGMA

MELT

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The volcanic events which lead to the creation of calderas involve the transport of large quantities of magma, rapidly, in P-T space. The stresses to which the magma is subjected are multiple, overlapping and ongoing during the entire eruptive event. The magma involved has, in theory, a number of response options to these stresses. In fact, the physico-chemical complexity of what can be gleaned from the eruptive products suggests that the magma endures a relatively complex path to the surface and responds and deals with these stresses accordingly.

The fundamental options for the magma are evolution of its state, relaxation of its properties, and/or failure of its integrity. All three of these options depend critically on the interplay between stress gradients in time and the rheology of the magma. The viscosity of the magma, about which we are obtaining an ever more detailed picture with time, is the most important control on it all.

The first option refers to the possible inducement of phase changes in the magma and/or the redistribution of chemical components between pre-existing phases. The primary avenue for such a response is the degassing-fed growth of new or pre-existing bubbles in a multiphase magma. The efficiency of this process is relatively well understood at the growth stage, less so at the nucleation stage. The degassing process is substantially accompanied by “parasitic” degassing that may influence significantly the geochemistry of what products are created by the eruption. The second state option is the crystallisation of condensed phases. Here the kinetic picture is very imperfect.

The second option, relaxation of its properties, is even more critically dependent on the rates of stress application, than the first. The relaxational response is an unavoidance and continuous response to stress changes that may be impeded by high viscosity or by enormous stress gradients in time. The polyphase nature of magma permits some statements about the strain history of erupted rocks and thus contains some information on the relaxational response, as do the resultant physical properties of the magma captured in eruptive products at the surface.

The third option, failure of integrity, is also set against criteria of stress application rates, once overcome, the integrity of the magma switches to the physical response of a particulate system and disappears from the physics of the eruptions. It may however return, in the post-eruptive emplacement of spatter-fed or pyroclastic fall deposits, where the initial contiguous phase is found as welded materials.

Eruptive products of caldera forming eruptions contain much more information on the conditions of their genesis than we have been traditionally used to exploring.
PRE-ERUPTIVE EXSOLVED GAS IN SILICIC MAGMAS: CONSTRAINTS FROM MAGMA DEGASSING DURING THE CA. 1340 A.D. ERUPTION OF MONO CRATERS, CALIFORNIA

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It is thought that many silicic magma systems contain a pre-eruptive, CO$_2$-rich vapor phase. For example, interpretation of fluid inclusion data from the 760-ka Bishop Tuff, California suggests approximately 25% by volume of pre-eruptive, exsolved CO$_2$ and H$_2$O. Eruption of the Bishop Tuff resulted in the formation of the Long Valley caldera. The Mono Craters magma system is spatially and temporally contiguous with the Long Valley caldera and CO$_2$/H$_2$O ratios in pyroclastic obsidian from the ca. 1340 A.D. Mono Craters eruption show increased concentrations of dissolved CO$_2$. This has been interpreted in terms of closed-system, equilibrium degassing during magma eruption. Equilibrium degassing requires a pre-eruptive volatile content of approximately 20%, consistent with results from Bishop Tuff fluid inclusions, but inconsistent with obsidian formation. We present results from a numerical conduit model of nonequilibrium magma degassing. We show that CO$_2$/H$_2$O concentrations in pyroclastic obsidian from Mono Craters may instead record nonequilibrium, open-system degassing during magma ascent to the surface. Our results indicate that permeability-controlled, open-system gas loss is consistent with obsidian formation and promotes nonequilibrium degassing at shallow depths. Because of the low diffusivity of CO$_2$ relative to H$_2$O, we find that CO$_2$ concentrations are above equilibrium during magma ascent. However, nonequilibrium is contingent upon low rates of bubble nucleation to shallow depths. We explore the feasibility and implications of these two contrasting models in terms of pre-eruptive magma degassing in silicic magma reservoirs.
SESSIONS: MONI+HARI

Multiparameter monitoring of restless caldera volcanoes and results from data modelling

And

Hazard assessment, risk mitigation and scenario planning at restless calderas
FLUID MIGRATION IS THE SOURCE OF DEFORMATION AT CAMPI FLEGREI CALDERA (ITALY).

MONI

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Campi Flegrei, a densely populated volcanic caldera immediately west of Naples (Southern Italy), has been the site of significant unrest for the past 2000 years. The caldera floor of Campi Flegrei rose 1.7 m between 1968 and 1972, and subsided 0.22 m between 1972 and 1974. From 1982 to 1984 the caldera floor rose 1.8 m, then subsided again at a rate of 1 to 6 cm/yr without an eruption or clear evidence of magma leaving the system. Such kind of ground movements, common to several calderas, can hardly be explained in terms of inflation and deflation of a magma body. Alternative models proposed include hot fluids migration or intense magma degassing. In this work, we bound the location, geometry and density of the intrusion by inverting levelling, trilateration and gravity measurements collected between 1980-84, and 1990-95. Given the difference between silicate melts (~ 2500 kg/m$^3$) and hydrothermal fluids (~ 1000 kg/m$^3$) density, this approach can help in distinguishing between possible sources of caldera unrest. We find that the best fitting inflation source is a penny shape crack, 2.8 to 4.0 km deep (95% bounds), radius between 1.1 and 2.7 km, and density 142 to 1115 kg/m$^3$. The deflation source is a vertical prolate ellipsoid, 1.8 to 2.4 km deep (95% bounds), aspect ratio of 0.35 to 0.62, and density between 849 and 1062 kg/m$^3$. In particular, the different shape of source during the subsidence phase, with respect to the starting uplift phase, can be interpreted in terms of deflation of fluids from the shallow aquifer. These results represent the first clear evidence that geothermal fluid migration plays a fundamental role in caldera unrests, and have important consequences for eruption forecast at these areas.
THE EFFECT OF FLUID FLOW ON CRUSTAL DEFORMATION IN LONG VALLEY CALDERA

Session ID: MONI

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Long Valley caldera in east central California has been an area of volcanic unrest for over two decades. Four M = 6 earthquakes occurred in 1980 and unrest has continued to the present day in the forms of seismic swarms, crustal uplift, and emissions of magmatic gas. The dominant mode of contemporary deformation in Long Valley caldera is uplift that is approximately radially symmetric. Previous interpretations invoke either elastic or viscoelastic response to inflation of the magma chamber. However, hydrothermal fluids may play more of a role than commonly acknowledged. We use numerical modeling to examine the role of aqueous fluids and CO₂ in crustal deformation. We couple a groundwater flow model (TOUGH2) with a mechanical deformation code (B IOT) to examine how fluids could affect surface deformation, and use a sensitivity analysis to quantify the likely volumes, rates, and fluid compositions required to create and maintain the observed ground-surface displacement in Long Valley caldera. The combination of forward numerical modeling and inversion techniques applied to match relevant geophysical data can help constrain the mode of caldera deformation. The approach is transferable to other large caldera systems.
We test ten volcanic areas in the western United States for annual periodic signals (seasonality) in their seismic catalogs, focusing on large calderas (Long Valley caldera and Yellowstone) and stratovolcanoes (Cascade Range). We use a combination of six statistical tests to analyze each area for seasonal seismicity during twenty-year periods. In four of the ten regions, statistically significant seasonality occurs (> 90% probability in five out of six tests), such that there is an increase in the monthly seismicity during a given interval of the year. In five regions, significant seasonal seismicity occurs in the shallow portions of the crust (the upper 3 km). Of the regions that show seasonality, Yellowstone Lake is the most statistically significant and increased seismicity occurs over a shorter interval. Peak seismicity occurs in the summer and autumn at Mt. St. Helens, Yellowstone Lake, and Mammoth Mountain. In Hebgen Lake/Madison Valley and the eastern south moat of Long Valley caldera (LVC) peak seismicity occurs in the winter and spring. We quantify the possible external forcing mechanisms that could modulate seasonal seismicity. Both snow unloading and groundwater recharge can generate large stress changes of > 5 kPa at seismogenic depths and may thus contribute to seasonal modulation of seismicity.
The Somma-Vesuvius volcanic complex consists of an older volcano dissected by a summit caldera, Mount Somma, and a recent cone, Vesuvius, which grew within the caldera of Mount Somma in the last 2000 years. At present, the northern rim of the caldera is a well-defined steep wall that progressively lower in the eastern sector and disappears in the southern and western sectors. This morphology is the result of the main explosive eruptions occurred in the last 18 ka. The present day caldera wall represents an obstacle for dispersion of pyroclastic density currents in case of renewal of explosive activity at Somma-Vesuvius. Because the maximum expected event (MEE) at Somma-Vesuvius is in the range of a Subplinian eruption, and the eruptive scenario comprises generation of pyroclastic density currents (PDCs), the effects of the morphological obstacle must be investigated. Geological data from past explosive eruptions yield valuable insights for assessing the influence of caldera wall in PDC behavior. In particular, the dispersion and physical characteristics of PDCs from three Plinian eruptions (Mercato, 8 ka; Avellino, 3.8 ka; Pompeii, AD 79) and two Subplinian eruptions (Pollena, AD 472; AD 1631) were investigated in detail. The collected data demonstrated that the capability to surmount the caldera wall is mainly function of position of the eruptive vent and eruptive mechanism. However, the caldera wall influenced the physical characteristics (e.g. grain size, sedimentology) of PDCs even when they are able to surmount the obstacle. The velocity vs. distance from the vent of selected diluted PDCs were also examined in open and sheltered areas. The collected data represent a starting point for improvement of hazard mitigation plans in the Somma-Vesuvius area.
THE NEED TO CONSTRAIN RHEOLOGY WHEN ASSESSING MAGMATIC SOURCE PROCESSES FROM GEODESY: THE CASE OF LONG VALLEY CALDERA, CALIFORNIA, USA

Session ID: STRC

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The massively explosive eruptions that produce large silicic calderas such as Long Valley in California are among the most violent geological phenomena on Earth. With ever increasing populations and infrastructure being built near such volcanoes, the hazards posed by devastating caldera eruptions are ever increasing. After at least 100 years of calm, Long Valley caldera has, in the past 25 years, actively begun uplifting, generating tens of thousands recordable earthquakes and significantly altering it's geothermal system. Though comparisons with activity at other volcanoes suggest that an eruption is not imminent, our understanding of the physical processes underlying these volcanoes is incomplete, and the potential for future eruptions remains significant.

Because of the recent activity, and abundant and diverse geodetic, seismic and geologic datasets available at Long Valley, the system remains a near-ideal study area to learn about the magmatic plumbing system before an eruption occurs. Unfortunately, geodetic modeling of volcanic activity generally averages data over long intervals and uses homogeneous Poisson solids to simplify fits to data. However, since well developed silicic systems generally have complex compositions, have a strong heat gradient near the source, and rapidly varying activity, these models can lead to inadequate results.

Here, I will use the available geodetic data (including EDM, GPS and InSAR) available from Long Valley's recent activity along with seismic (tomography and microearthquake locations) and geologic (caldera structure and strength) data to show how local rheology, including weakened and time-dependent viscoelastic crust, and volume significantly changes the estimates of pressure, shape and depth of the source of activity. These are important parameters for understanding the physical processes in a shallow magmatic system prior to eruption, and are necessary to better constrain in order to improve eruption forecasting and prediction.
The analysis of the seismic activity of the island of Tenerife from 2001 shows a clear interaction between the regional tectonic activity, the magmatic camera and the walls of the Caldera.

In this work the seismic activity has been classified in areas, using as tool the GIS, and Markov's matriz were calculated, showing the transition probabilities from a region to another. In these systems, what is important is the last seismic event happened who conditions the possibilities of the following one and, therefore, conditions the pattern of space flow.

The objective is to establish a guideline of volcanic system’s behaviour which permits, simultaneously, explain its dynamics. This analysis applied in different temporary intervals, allowing to define the diverse activity phases.

The Data belong to the seismic catalogue of the National Geographical Institute. The Data were filtered according to the space areas defined for the island of Tenerife. The initial definition of the matriz was carried out with a high number of areas and then they were reduced until obtaining a stable group of matriz of 5 cells.

The temporary evolution of the system is also another factor to keep in mind. From May of 2004 onwards an important seismicity increment inside the island begins to be registered. Nevertheless, its spatial distribution and their characteristics have not always been equal, having as points of change two dates: 17th October and 7th December 2004. These dates, an important increment of the activity culminated with the apparition of an important flow of gas on Mountain Teide’s crater and the opening of a fracture in the Valley of Orotava, also with an emission of constant gases.

The results show that the increase of volcanic activity provokes superficial seismicity in north and west face of Caldera’s walls, as a consequence of regional seismicity. Probably, regional activity has destabilized the magmatic system, causing a fracture of caldera’s walls. The study of the temporary evolution of the magnitudes of seismic events next to the surface, suggests an increment of the size of the fractures.

The application of this methodology (that is, the distribution of seismic events by means of Markov’s matriz) has shown to be a powerful tool to characterize the volcanic seismicity in a complex volcanic system.
The Phlegraean Fields caldera is an active volcanic system where important episodes of ground deformation have been accompanied by significant changes in geochemical and geophysical parameters monitored at the surface. Volcanic unrests, even if not accompanied by a renewal of eruptive activity, bear important consequences in a densely populated region. Early warning and a proper evaluation of the state of evolution of the volcanic system are therefore essential to ensure a proper management of possible crises and to mitigate volcanic hazard. To this regard, geochemical and geophysical monitoring play an essential role. A proper interpretation of monitoring data, however, is only achieved within the framework of a robust conceptual model of the system. Recent research work carried out at the Phlegrean Fields points out the relevant role played by hydrothermal fluids in the evolution of the caldera. A pulsating magma degassing, periodically discharging greater amounts of CO$_2$-enriched fluids into the shallow hydrothermal system, was shown to be consistent with the observed compositional variations. Periods of increased degassing were also shown to induce a significant amount of ground deformation. Changes in fluid density also generate gravity changes which are detectable at the surface. In this work, numerical modeling of heat and fluid flow through porous media have been applied to study the role of a pulsating magma degassing in generating compositional and gravity changes, as a function of fluid composition and temporal evolution of degassing rate. Modeling results allow to define possible scenarios corresponding to different degassing histories. Simulations confirm that hydrothermal fluid circulation play an important role in the evolution of this caldera.
THE CAMPI FLEGREI CALDERA (SOUTHERN ITALY): MODELING, INTERPRETATION AND HAZARD ESTIMATION

HARI

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This work reviews the main geophysical observations, modelling interpretations and hazard estimation at Campi Flegrei caldera, both in normal periods and during unrests involving very high uplift and seismicity rates. Campi Flegrei caldera is a volcanic area partially including the city of Naples (Southern Italy), Explosive activity, mainly hydromagmatic, characterises this area, in which huge ground deformations, well known since ancient Roman times, occur. In the last 30 years, two episodes at least of spectacular ground deformation and seismicity have occurred. This work reviews the main issues and problems in the interpretation of Campi Flegrei activity, presenting a coherent model which involves both elastic and thermal fluid-dynamical effects. The basic concepts, which are thought to be distinctive features of activity at all the calderas, involve the close interplay among magmatic stress, bordering caldera faults and shallow geothermal system, with an important contribution of background regional stress. The work presents new insight for interpreting pre-eruptive sequences at calderas, and new methodologies for the estimation of hazard in such a densely populated area.
SESSION: MODE

Analogue and numerical modelling of magma chamber and eruption processes leading up to and triggering caldera collapse
THE DEEP STRUCTURE OF CALDERAS

MODE

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Understanding the structure of calderas is crucial to try to predict their behavior during an unrest episode. However, field observations on the best exposed active calderas do not permit to get insights on their deep structure. Similarly, outcrops of Neogene and Paleozoic eroded calderas do not allow a comprehensive understanding of their deeper architecture. Therefore, analogue experiments and numerical models remain among the most useful tools in trying to understand the deeper structure of calderas.

Several sets of analogue experiments, with different materials and apparatus, have been performed in the last years to understand the structure of calderas as a result of overpressure within the magma chamber. In these experiments, the deep caldera structure usually consists of a pair of concentric ring structures. An outward dipping reverse ring fault forms first as a result of a differential uplift, and constitutes the structural border of the caldera. When the vertical displacement exceeds a certain threshold (corresponding to few hundreds of meters in nature), this is followed by an inward dipping normal ring fault, developed as a result of gravitational instability. The diameter and the depth of nucleation of the inner and outer calderas show constant proportions. Several pit craters and collapse calderas in nature have similar geometries and proportions.

Various numerical and analogue models also show that, in case of overpressure within a (a) sill-like chamber subject to doming and (b) laccolith generating apical tensile stresses, a caldera, bordered by inward dipping normal faults, may form. In these cases, the total amount of subsidence is usually more limited and the diameter of the caldera is much smaller than that of the underlying reservoir.

Therefore, the results of analogue and numerical models suggest that the deep structure of calderas mostly depends from (1) the underpressure or overpressure conditions within the underlying reservoir and (2) the amount of subsidence.
MODELING OF PYROCLASTIC FLOW PROPAGATION IN A CALDERA SETTING: THE CASE OF PHLEGREAN FIELDS CALDERA (ITALY)

Session: MODE

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The Phlegrean Fields caldera represents a typical example of complex caldera setting, where large and small volcanic features, such as cones and craters, punctuate an otherwise flat topography. Future volcanic activity at the Phlegrean Fields is expected to be explosive, and likely involving the generation of pyroclastic flows. The dynamics of propagation of these flows bears important consequences in terms of hazard evaluation in a densely urbanized area. In particular, a key question is whether or not pyroclastic flows could eventually overcome the large hills bounding the eastern and northern sides of the caldera, and directly threaten the town of Naples. To investigate the dynamics of flow propagation we simulated magma ascent along the volcanic conduit and its dispersal into the atmosphere along selected topographic profiles. The magma properties employed in the simulations are representative of the 4400 BP Agnano Monte Spina eruption. Grain size and particle properties, representative of pumice, ash and lithic fragments, have been also explicitly accounted for into the model. Three different topographies have been considered, to study the effects of obstacles at different distances from the vent. Modeling results showed that flows generated by large scale eruptions can easily overcome the caldera walls. Obstacles on flat terrains induce back-flows that move toward the vent region and interact with newly discharged material. Effects may propagate backwards and affect fountain dynamics and even eruptive style. Obstacles are also responsible for particle sorting which, in turn, reflects on the run-out of the flow. Particle sorting is also expected to affect grain size distribution in tephra deposits that form at different locations. Particle sorting, back-flows and their effects on overall flow dynamics all depend on number, geometry and location of obstacles.
EVENT FORECAST WITH FFM IN THE REACTIVATION OF A CALDERA: 
APPLICATION TO THE 2004 SEISMIC CRISIS OF TENERIFE.

MONI

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The Materials Failure Forecast Method (FFM) is described by the differential equation
\[
\frac{d^2 \Omega}{dt^2} = A \left( \frac{d\Omega}{dt} \right)^n,
\]
where \( \Omega \) is an observable linked to the deformation of the system. When an experimental data time series (a potential precursor) can be regarded as a solution to this differential equation, one can compute the time when an event is likely to occur by the extrapolation of the descending parts in the time evolution of the inverse of the given observable. This event corresponds to a significant perturbation of the magmatic system (eruption or seismic event).

From the end of the 1980's the FFM has been applied to several volcanoes with good results when applied to the case of explosive volcanism. On the contrary, no reference exists to a successful application of the method to the case of effusive eruptions, typical e.g. of a basaltic volcanism.

The experience obtained during the seismic crisis of Tenerife with the application of the FFM on the observable RSEM (Real-time Seismic Energy Measurement) shows a sufficiently good correlation between the obtained forecasts and the occurrence of tectonic seismic events. To avoid the problem of human subjectivity, an automatic program has been developed for this study to generate the forecasts. Moreover, Bayes' theorem has been applied to validate the results. During the most energetic phases a promising percentage of successful forecasts was obtained.
Regional Stress Changes Triggering an Eruption: Properties of Viscoelastic Media Surrounding the Magma

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Evidence of coupling between tectonic events and eruptions has been reported in different volcanoes around the world. Maintained stress or changes in the stress fields in active volcanoes can trigger an eruption even in a system remaining below a critical condition. Rock holding a magma may undergo a material failure if a subcritical stress is set and maintained for a long enough time. The material failure process produces a characteristic release of strain that may be detected as an accelerating rate of cumulative seismic energy in nearby monitoring stations. The shape of the accelerating seismic energy release provides information about the stress history leading to an eruption. Although these precursory accelerating patterns do not always appear, at times they can be successfully applied to predicting the time of an eruption. When the triggering stress remains constant, the characteristic accelerating patterns follow a hyperbolic law. A linearly growing stress produces a logarithmic increase of the seismic energy release. A real-time monitoring device set at Tungurahua volcano in 2003, permitted to observe in 20 August a characteristic accelerating pattern conducting to an eruption. The onset of the pattern is marked by a M4.5 regional earthquake. The analysis of the cumulative seismic energy indicates that the regional earthquake induced a linearly increasing stress on the volcano for about 320 minutes, producing a logarithmic increase of the seismic rate, and then remained constant continuing with a hyperbolic increase until the holding material failed causing the eruption. The duration of the increasing stress is interpreted in terms of the relaxation time of the viscoelastic media surrounding the magma. Assuming a material rigidity of \(~30\) GPa, this corresponds to a bulk viscosity component for the viscoelastic material of order \(10^5\) GPa.s.
CALDERA RING FRACTURE REACTIVATION
CONTROLLED BY EXTRINSIC STRESS CHANGES

Session ID: MONI

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There is increasing evidence that volcanoes are influenced by very small perturbations of the stress field. In this project we elaborate extrinsic processes that cause reactivation of caldera ring fractures. We consider mechanisms common at caldera volcanoes, specifically (i) short-term mechanisms (earthquakes, landslides, or volcanic activity), and (ii) long-term mechanisms (regional tectonic deformation, rifting and lateral spreading).

In theoretical models we simulate stress changes associated with magmatic and tectonic events and calculate traction and stress changes at concentric and dipping ring fractures. Our models suggest that ring fractures may locally slip or open as magma pathways, depending on the type and locality of the preceding extrinsic mechanism. An increase in Coulomb failure stress shows that caldera faults are reactivated, whereas tensile traction shows that dike intrusion is facilitated (unclamping). Following a systematic study of the stress field changes, we apply the modeling technique to Fernandina Volcano, Galapagos:

In 1995 a southwest flank eruption occurred in Fernandina Volcano, for which InSAR ground deformation suggest a dipping dike oriented radial and outside of the ~5 km wide caldera. We apply the 1995 dike geometry and calculate associated stress changes at the ring fractures. Our model calculations suggest that the 1995 dike significantly unclamped the caldera ring fractures; unclamping is most in the southwest and decreases towards the southeast. Indeed, on May 2005 a concentric fissure eruption occurred at the ring fracture. The first lava flows emanated from the western part of the fissure and the latest flows from the eastern part. An exciting and promising result is that the location and development of the 2005 Fernandina fissure eruption almost perfectly matches the zone unclamped by the previous dike.
SESSION: STRC

Implications for mechanisms of caldera formation from structural investigations
The Las Cañadas caldera (LCC) is one of the most important geological structures of Tenerife whose origin is still a matter of debate. The analysis of its whole internal structure is an additional key for understanding the complex sequence of vertical and/or flank collapse events which have occurred there. Moreover, being a part of the main groundwater reservoir of the island, this survey aims at providing new elements for understanding the hydrogeological behaviour of this system and relationships between presumably different aquifers.

Our AMT measurements show that at short periods (less than 1s) the data show a uniform, close to 1D behaviour on most of the caldera, in agreement with Pous & al. (2002). The shape of the apparent resistivity and phase curves reveals a conductive layer at shallow depth, underlying a more resistive cover. The strong electrical resistivity contrast makes AMT an efficient method in this context.

A very dense mapping (270 AMT sites) of the conductive layer in the eastern part of the LCC is obtained from one-dimensional modelling. Very low frequency (VLF) measurements crossing the presumably buried western caldera wall argue for its presence through a vertical conductive anomaly.

We interpretate the conductive layer as the hydrothermal alteration of pyroclastic deposits and paleosols or breccias underlying landslides. From its morphology, we infer a close correlation between main structural features such as ring faults, with vents and volcanic edifices location. LCC morphology may be divided into three sectors corresponding to the structural models of Marti & al. (1994, 1997), with a deepening and opening toward the northeast. These results are in very good agreement with the age of the giant structural events, their morphologies and available groundwater flow rates around the LCC.

Morphology of the roots of the Icod landslide seems to follow from multiple northward directed lateral collapses. A groundwater flow axis runs from the southwest to the northeast limited by the Pico Viejo - Teide complex and by the southern caldera wall.
EFFECTS OF STRESS FIELDS, STRUCTURES, AND MATERIAL PROPERTIES OF VOLCANOES ON THE FORMATION AND SLIP OF RING FAULTS

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The ring faults of collapse calderas are rock fractures; the magma chambers from which they develop are rock cavities. The initiation and development of any rock fracture depends on the state of stress in the host rock. The state of stress in a volcano is controlled by the material properties of its rock units and structures (existing contacts, faults, joints, etc.), as well as by the loading conditions. The loading conditions are determined by the tectonic environment and, in particular, the geometry and magma pressure of the associated chamber.

To understand how and when ring faults develop, and why existing ring faults slip so infrequently, one must know the state of stress in the associated volcanoes. This implies the knowledge of the rock properties and structures of the volcanoes. Furthermore, to forecast whether a ring fault is likely to form, or slip to occur on an existing one, during a particular unrest period we must have a rough idea of the geometry of the associated magma chamber. Ring-fault formation and slip are mechanical processes that cannot be forecasted solely on the basis of empirical criteria; if we are to develop viable models to assess the probability of ring-fault formation or slip these processes must be understood in mechanical terms.

Most ring faults are either circular or slightly elliptical in plan view. The faults are vertical or steeply dipping, and have vertical displacements from several hundred metres to a few kilometres. For example, in active composite (central) volcanoes in Iceland, the average maximum ring-fault diameter is 8 km (18 km being the largest), the average minimum diameter 3 km, and the maximum vertical displacements reach 500-700 m; similar ring faults occur in the extinct Tertiary and Pleistocene volcanoes. Traditionally, ring faults have been modelled using analytical, analogue, and numerical methods. While analytical methods are now largely obsolete, analogue and numerical models, combined with rigorous testing using field data, have in recent years provided many new ideas and improved understanding of the conditions of ring-fault formation and slip.

In this talk I present numerical models on ring-fault formation in isotropic (non-layered) and anisotropic (layered) host rocks. The main results may be summarised as follows: (1) Magmatic underpressure (below lithostatic) or excess pressure (above lithostatic) as the only loading normally generates a stress field which favours dyke injections rather than ring-fault formation. (2) A spherical (circular) chamber in a volcanic field subject to doming, tension, or both is unlikely to trigger ring-fault formation, unless the chamber is located in a very soft (low Young’s modulus) layer or has recently injected dykes. (3) The numerical models indicate that a ring fault (and a ring dyke) is most likely to form, in layered as well as non-layered host rocks, in a local stress field generated by a shallow sill-like chamber in a volcanic field subject to doming, tension, or both. (4) For a 20-km wide layered volcanic field, either tension or tension combined with doming may result in ring-fault formation. (5) For a 40-km wide layered volcanic field, tension is not necessary; doming alone is sufficient to trigger ring-fault formation.
Our understanding of the lithospheric structure beneath active volcanic calderas is far from complete. This is largely due to the resolution limit of geophysical imaging techniques, as well as to the ambiguities inherent in interpreting geophysical data. Information on the subsurface structure not only serves economic interests, such as hydrothermal and ore exploration, but also provides valuable data for the validation of results from analogue, analytical or numerical modeling on caldera processes. Furthermore, data interpretation from dynamic geophysical investigations, such as geodetic or microgravimetric studies, can be facilitated by and critically assessed against coherent images of subsurface structures.

This talk explores our current knowledge on the subsurface beneath selected active caldera centres in arc, intracontinental and oceanic island settings including Rabaul, Campi Flegrei, Valles, Taupo, Toba, Las Canadas and Long Valley.

Data from magnetic, seismic, gravimetric and electric investigations as well as borehole data are employed in order to explore where results from the different techniques converge to provide a coherent picture of the structural arrangement beneath these volcanoes. This includes the identification of faults and their inclination, of magmatic and hydrothermal reservoirs and their geometry, in addition to magmatic rejuvenation processes.
A REVIEW ON COLLAPSE CALDERAS MODELLING

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A meaningful study of caldera collapse formation should ideally involve multiple aspects like regional tectonics, system geometry, magma and host rock properties, fluid-structure interaction, pre-existing structural discontinuities, deformation history, etc. Due to the impossibility of a such comprehensive analysis, studies have centred just on relevant but atomized topics. From a methodological point of view and in addition to the essential field and petrological studies, collapse calderas have also been investigated through analogue and theoretical models. Each approach presents advantages and drawbacks. We review the most significant contributions, summarize the relevant outcomes, and highlight the strong points and weaknesses of each approach. Analogue models allow a qualitative study of the structural evolution of a caldera collapse process and suggest which geometric factors play a relevant role. Differences among these models lie behind the applied experimental devices, the host rock analogue materials (dry quartz sand, flour, etc.), and the magma chamber analogue (water or air-filled balloons, silicone reservoirs, etc.). However, the results obtained are not substantially different when the same input parameters are considered. Discrepancies between results mainly come from the restrictions of each experimental design. Theoretical models have grown in importance during the last decades in parallel with the development of computational resources. Nowadays, these models constitute a significant source of information on caldera-forming processes and can predict semi-quantitative general conditions for fault formation and propagation. Theoretical studies can be classified in two groups according to their objectives. One group focuses on the evolution of pressure within the magmatic reservoir during a caldera-forming eruption, whereas the second looks more into the structural conditions of a collapse and hence interacts with the analogue models.
RELATIONSHIP BETWEEN CALDERA COLLAPSE AND MAGMA CHAMBER WITHDRAWAL: AN EXPERIMENTAL APPROACH

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Collapse calderas have received considerable attention due to their link to Earth’s ore deposits and geothermal energy resources, but also because their tremendous destructive potential. Although calderas have been investigated through fieldwork, numerical models and experimental studies, some important aspects still remain poorly understood. Following several published experimental models addressed to characterize the structural features and controlling parameters of caldera collapse processes, this work presents a correlation of the different stages of caldera formation with the erupted volumes of magma from the underlying reservoir, a subject addressed only partially by some analytical models. In order to correlate the structural evolution of a collapse caldera with the erupted magma chamber volume fraction $f$, we perform analogue experiments within a transparent box (60 x 60 x 40 cm) filled with dry quartz sand and using a water-filled latex balloon as a magma chamber analogue. Evacuation of water through a pipe causes a progressive deflation of the balloon that leads to a collapse of the overlying structure. The experimental device allows to record the temporal evolution of the collapse and to track the evolution of fractures and faults. We distinguish up to six different steps or stages characterized by the appearance of new structural features or by a distinctive evolution of previous ones, and correlate each different step with the corresponding removed volume fraction $f'$. We also determine the critical conditions for caldera onset. Experimental results show that for any step, the experimental relationship between $f'$ and the chamber roof aspect ratio $r$ ($r = \text{roof thickness/chamber width}$) fits into a logarithmic curve. This implies that the erupted magma chamber volume fraction $f$ required to trigger a caldera collapse is higher for chambers with low aspect ratios (shallow and wide) than for those with high values of $r$ (deep and small). These new results are in agreement with natural examples and previous analytical studies.
Archean subaqueous calderas in the Superior Province, Canada are primary loci for volcanic-hosted massive sulfide (VHMS) deposits and have a highly variable physical volcanology. The 2-6 km-thick volcanic edifices of predominantly felsic composition feature 25-400 m-thick rhyolite flow units, including dome-flow-hyaloclastite complexes, and 5-650 m-thick (reworked) pyroclastic and eruption-fed density current deposits. The >2.7 Ga calderas can be divided into 1) explosive-types, e.g. the Mattabi caldera and 2) effusive-types, e.g. Blake River, Hunter Mine and Normetal calderas. The shallow-marine Mattabi caldera is dominated by 15-650 m-thick composite volcaniclastic units, with local VHMS deposits. Mattabi subaqueous pyroclastic deposits are divided into a) 10-155 m-thick massive lapilli tuff, b) 6-48 m-thick massive to graded lapilli tuff, and c) 1-13 m-thick graded tuff, and display a similar internal organization to deep-water, fire-fountain deposits in the Hunter Mine caldera or the classic, pyroclastic Dalembert tuff, Blake River Group. The Blake River caldera formed in a deep - to shallow-water, whereas the Hunter Mine and Normetal calderas are consistent with deep-water conditions, as shale/mudstone or basalts/komatiites cap the sequence. Deep-water conditions are consistently characterized by effusive volcanism, rather than magma-draining ash-flow eruptions that are common to modern-day calderas in shallow-water (e.g. Myojin Knoll, Izu-Bonin arc). The explosivity is controlled by a balance of the hydrostatic pressure column, which dampens eruptions, with rapid degassing of volatile-rich magma that enhances paroxysmal eruptions. Such magmatically driven eruptions produced the hot to cold Hunter Mine fire-fountains, with ingestion of water into the eruption column and during transport. Deep-marine, flow-dominated calderas may be formed by constant draining of the magma chamber, and shifting of magma to satellite reservoirs, which explains incremental and slow subsidence. Flow-dominated calderas are favourable for VHMS formation because hydrothermal systems are not destroyed by large-scale explosions. Synvolcanic ring and graben fractures facilitate dyke injection and magma ascent, and are used for hydrothermal venting. Felsic dyke swarms, 5-7 km wide, indicate extension and facilitate caldera subsidence. Archean calderas are significant because the vertically dipping strata display a rare cross-sectional view of the volcano rather than a surface morphology inherent to modern volcanoes.
FIELD GUIDE TO

THE LAS CAÑADAS CALDERA VOLCANISM

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INTRODUCTION

THE CANARY ISLANDS

The Canary Islands lie in the eastern Atlantic Ocean adjacent to the African passive continental margin (Fig. 1). The age sequence and spatial distribution of the Canaries and nearby seamounts is somewhat complex compared to other intraplate island groups such as the Hawaiian chain, and a variety of models has been proposed for Canarian magmatism. The models fall into two groups: (i) dominant lithospheric control of magmatism (e.g. Anguita & Hernán, 1975; Araña & Ortiz, 1991); (ii) a mantle plume origin (e.g. Morgan, 1972; Hoernle et al., 1991; Hoernle & Schmincke, 1993; Carracedo et al., 1998). There are two basic reasons for the controversy. Firstly, the proximity of the Canaries to a regional belt of active deformation in the Atlas Mountains of Morocco allows construction of plausible scenarios for lithospheric decompression and consequent mantle melting (Araña & Ortiz, 1991). Secondly, in variants of the plume model, the sluggish motion of the African plate and regional variations in lithospheric thickness are expected to result in deviations from the classic 'Hawaiian' model of a narrow plume feeding a long island chain with orderly age progression of volcanism (e.g. Hoernle & Schmincke, 1993). Recent workers have tended to favour the plume model (Carracedo et al. 1998; Geldmacher et al., 2001), which is consistent with geochemical evidence. In particular, the temporal pattern of isotopic variation in lavas of the tiny Selvagen Islands and nearby seamounts north of the Canaries is similar to that seen in the archipelago, which may therefore all be part of a single hotspot track originating in the late Cretaceous (Geldmacher et al., 2001; Fig. 1). The Canaries plume apparently sourced young HIMU mantle that contains a component of oceanic lithosphere subducted at about 1 Ga (Thirlwall, 1997; Simonsen et al., 2000). During construction of the islands, temperature and depth of partial melting of the mantle plume was variable in space and time, and the rising magmas and associated fluids experienced a range of complex interactions with regional lithosphere (Hoernle & Schmincke, 1993; Neumann et al., 1995, 2000, 2002, 2004; Simonsen et al., 2000; Thirlwall et al., 2000).

![Figure 1. The Canary and Selvagen Islands in the eastern Atlantic Ocean. Also shown in grey are the axes of two possible hotspot tracks (each assuming a different model of African plate motion) beginning at 60 Ma in the vicinity of the Lars Seamount (Geldmacher et al., 2001), and with present locations near the westernmost Canary Islands. The actual zone of melting around the hotspot axis is envisaged as ~450 km broad in each case.](image-url)

THE VOLCANIC EVOLUTION OF TENERIFE

Tenerife is an extremely complex volcanic system, having been at various times in its history an oceanic basaltic shield volcano complex, a stratovolcanic edifice and a highly explosive caldera system, sometimes with more than one of these personas active simultaneously. Tenerife ranks as the second largest shield oceanic island on Earth after the island of Hawaii (Mauna Loa, Kilauea and Mauna Kea). It has been constructed by numerous phases of volcanism, encompassing a diverse range of volcanic processes and spanning...
more than 12 million years (Table 1; Fig. 2). This history can be subdivided into four major stages:

- **Stage I (>12 – 3 Ma)** involved the growth of several emergent alkali basaltic shields (Old Basaltic Series).
- **Stage II (3.8 – 2 Ma)** saw the construction of a central volcanic complex (Lower Group of the Las Cañadas Edifice).
- **Stage III (2 – 0.2 Ma)** includes three extended cycles of highly explosive phonolitic activity and associated caldera formation (Upper Group of the Las Cañadas Edifice) with ongoing shield-building by basaltic flank eruptions (Dorsal Rift Zone, Santiago Rift Zone, Southern Volcanic Zone) and major debris avalanche events (Güímar, La Orotava and Icod valleys).
- **Stage IV (0.2 – 0 Ma)** witnessed the growth of a new central stratovolcanic complex within the older caldera (Teide-Pico Viejo Formation), accompanied by continued shield-type flank activity (Recent Basalts). Volcanic activity has continued into historical times, with the most recent eruption occurring in 1909.

![Figure 2. Generalised geology of Tenerife, depicting the major stages in the volcanic evolution of the island. From Pittari (2004), after Ablay and Kearey (2000).](image)

There is considerable debate on the origin of the Las Cañadas Caldera, the large central summit depression. One model attributes the caldera to multiple, debris avalanche mass wasting processes and the other to major caldera collapse associated with large-scale explosive eruptions. Independent work on the stratigraphy and geochronology have led to competing stratigraphic schemes, and interpretations of the island’s evolution (e.g. Martí et al., 1994; cf. Ancochea et al., 1999; Huertas et al., 2002, Bryan 1998, Edgar 2003, Middleton in prep.).

In this review the stratigraphy developed by Martí et al. (1994, 1997), Bryan (1998), Edgar (2003) and Middleton (in prep.) will be employed. The estimated volumes of the pyroclastic products of individual plinian eruptions as well as several other lines of evidence are consistent with caldera collapse, although large-scale landslide events were also clearly important in the volcanic history of the island (see below), and at times probably linked to major caldera forming eruptions.

### THE OLD BASALTIC SERIES (OBS)

The oldest exposed rocks on Tenerife are known collectively as the Old Basaltic Series, OBS (Fuster et al., 1968). The OBS outcrops in several disconnected and highly dissected massifs – Roque del Conde in the southwest, Teno in the northwest and Anaga in the northeast (Fig. 2). These massifs comprise the products of millions of years of fissure-fed
eruptions of alkali basalts, ankaramites and basanites, including lavas, pyroclastic and epiclastic deposits and shallow intrusive rocks (Table 1; Fúster et al., 1968). Minor felsic products occur near the tops of the Teno and Roque del Conde successions and throughout the Anaga sequence (Ancochea et al., 1990). A detailed account of stratigraphic data and K/Ar age determinations for these rocks is given by Ancochea et al. (1990) and more recent $^{40}Ar/^{39}Ar$ dates and geochemical data have been presented by Thirlwall et al. (2000).

The OBS represents the construction of several Hawaiian-type alkali basaltic shields, which emerged above sea level at least 12 Ma ago and remained active until 3.3 Ma (Ancochea et al., 1990). The intensity of volcanism varied between massifs; most of Teno was rapidly constructed between 6.4 and 6.0 Ma, whereas activity at the Anaga and Roque del Conde involved multiple magmatic cycles and lasted several million years (Ancochea et al., 1990; Thirlwall et al., 2000). This old basaltic shield complex accounts for most of the volume of the island, even though its remnants only occupy a fraction of the present surface area due to extensive cover by more recent volcanic products (Martí et al., 1995). The original structure and morphology of the shield complex are unclear. Basaltic shield building volcanism however continues to the present day through activity from the flanking rift zones.

OBS products are alkali basalts and basanites and include primitive compositions with ca. 10% MgO. Radiogenic isotope ratios vary significantly from massif to massif among broadly contemporaneous lavas, indicating either a spatially variable mantle plume source and/or varying degrees of magma-lithosphere interaction (Simonsen et al., 2000; see also Hoernle et al. 1991, Thirlwall et al., 1997, and Widom et al., 1999 for discussion of the magma sources of other Canary Islands). The Teno and Anaga lavas lie at either extreme of the isotopic range, with the Roque del Conde lavas intermediate between these two. All three massifs contain limited proportions of more fractionated lavas; these are noteworthy because, with rare exceptions, they follow a weakly alkaline differentiation trend of hawaiite-mugearite-benmoreite, rather than the strongly silica-undersaturated phonotephrite-tephriphonolite-phonolite trend that characterises products of the Las Cañadas edifice (Fig. 3). Thirlwall et al. (2000) suggest that this could be due to high oxygen fugacity and consequently early crystallization of magnetite during differentiation of the OBS magmas, however magnetite precedes plagioclase on the liquidus in later basanites of the Upper Group (Ablay et al., 1998; Olin, 2003) that are intimately associated with strongly undersaturated differentiated magmas, hence some other reason for the contrast in differentiation trends must be sought.

Neumann et al. (1999) invoked fractional crystallization in periodically refilled magma chambers (O'Hara, 1977) to explain large variations in incompatible trace element contents of OBS and younger basaltic lavas at constant Mg-number; many of these lavas have pyroxene phenocrysts with cores that grew from highly differentiated liquids. This model has been challenged by Thirlwall et al. (2000), who ascribed the trace element features instead to variation in depth and extent of mantle melting followed by fractionation and accumulation of magma. At issue here are the relative roles of deep and shallow level processes in controlling the compositions of mafic magmas that make up the bulk of the island and may represent parents to the rest of the Tenerife suite; these issues await resolution.
Figure 3. Total alkali-silica classification diagram (LeBas et al., 1986; pb = picrobasalt, h = hawaiite) for approximately 1,000 volcanic rocks from Tenerife. Data sources are cited in the text. Only rocks that could be assigned to one of the formations in the stratigraphy employed in this guide are plotted; plutonic xenoliths and lithic fragments are also excluded. Note the negative 1:1 slope of the most silica-rich compositions among phonolites and apparent trachytes; this population consists mostly of non-welded pyroclastic rocks with high LOI contents and the trend of the boundary is thought to be dominated by post-eruptive alkali loss from pumice and glass shards, shown by the bold arrow. We suggest that there are, in fact, no true trachytes in the Upper Group. Among the Old Basaltic Series, however, most of the analyses plotting in the benmoreite and trachyte fields have low LOI contents and their relatively alkali-poor character is considered to be a magmatic feature; the same may be true of a few Lower Group samples (not distinguished) that plot as trachytes.

THE LAS CAÑADAS EDIFICE: LOWER GROUP (LG)

The Las Cañadas Edifice is the major central volcanic complex of Tenerife (Hausen, 1956; Füster et al., 1968; Araña & Brändle, 1969; Ridley, 1970a, 1970b; Araña, 1971; Ancochea et al., 1990, 1999; Martí et al., 1994) and it was built on the remnants of the Old Basaltic Series. Two major stratigraphic intervals have been identified by Martí et al. (1994): a Lower Group (LG) representing a complex constructive phase and an Upper Group (UG) characterized by cycles of highly explosive, caldera forming volcanism and partial destruction of the edifice (Table 1). Only the upper part of the Lower Group is exposed at the surface. The best-studied outcrops occur in the bounding scarps of the Las Cañadas Caldera (Fig. 4). The Roques de García form a spur of LG rocks which projects out into the Las Cañadas Caldera. Lower Group lavas are also widely exposed beneath Upper Group deposits on the outer flanks of the edifice, particularly in the west and in the deeper barrancos of the southeast (Fig. 5).

Early radiometric dating pointed to a hiatus in volcanic activity between the end of the OBS construction (3.3 Ma) and the beginning of the Las Cañadas activity (1.9 Ma) and this was characterized as a period of quiescence and erosion (Ancochea et al., 1990). More recent dating (Martí et al., 1994), however, has revealed that Lower Group activity began as far back as 3.8 Ma with the construction of a mafic edifice known as the Boca Tauce volcano, whose presence below the surface has been interpreted from gravity data (Ablay & Kearrey, 2000). There was thus no significant interruption in volcanic activity between Stages I and II. There was, however, a significant period of repose between the end of the Lower Group
The complex stratigraphy of the Lower Group (LG) has not been studied in detail but Martí et al. (1994) subdivided it into seven informal "sequences" on the basis of palaeosols and erosional unconformities. As exposed in the caldera wall, these are (in clockwise order): El Cabezón, Las Pilas, Las Angosturas, Montón de Trigo, Roques de García, Boca de Tauce and El Cedro (Fig. 4; Table 1). The geochemistry of these sequences ranges from basaltic and trachybasaltic to phonolitic, and deposits include lavas, welded rocks and non-welded pyroclastic deposits (Martí et al., 1994). These authors attributed the stratigraphic complexity of the preserved succession to the existence of a number of overlapping vents rather than a centralized system.

The Lower Group represents the transition from an early shield-building phase (represented by the Old Basaltic Series) to post-shield volcanic activity, as observed on other islands in the Canary group. The greater proportion of evolved compositions in the eruptive products of Lower Group age signifies the development of relatively shallow magma chambers in which crystal fractionation and crustal assimilation could occur. These eruptions gradually constructed a composite stratovolcanic complex. Morphological reconstructions suggest that this edifice was elongate along a NNE-SSW axis and attained an altitude of 2700-3000 m (Araña, 1971).

Of all the major stratigraphic divisions on Tenerife, the Lower Group is currently the least well petrologically and geochemically characterised. Overall it is more highly evolved and silica-undersaturated than the differentiated flows within the OBS; Zafrilla (2001) reports feldspathoid-bearing phonolites at El Cedro. However trachyte flows occur within the Lower Group SE of Las Cañadas. Considerably more work is required before the petrogenesis of the Lower Group is understood.
Table 1. Stratigraphic scheme for the island of Tenerife, showing the major constructive and destructive episodes as presented in the literature (see text for references). From Edgar (2003).
Figure 4. (a) Schematic illustration of the stratigraphy of the Las Cañadas caldera wall, showing the seven sequences of the Lower Group and three formations of the Upper Group (From Edgar, 2003, after Martí et al., 1994). (b) Map showing the caldera wall sectors and outlines of the three major vertical collapse calderas as proposed by Martí et al. (1994): 1 - Ucanca, 2 - Guajara, 3 - Diego Hernández (i.e. younging to the NE). The inferred positions of the La Orotava (LO) and Icod (I) landslide scarp are also shown (dashed lines); compare with the landslide model in Fig. 2.9a. PG: Pico de Guajara, T: Teide, PV: Pico Viejo.

Figure 5. Las Cañadas Edifice, Lower Group (highly simplified). From Edgar (2003). Data sources: Martí et al. (1994); Füster et al. (1994); Bryan (1998); Ablay & Hürlimann (2000, Boca Tauce Volcano).
THE LAS CAÑADAS EDIFICE: UPPER GROUP (UG)

The Upper Group was defined by Martí et al. (1994) in their stratigraphic and geochronological study of the Las Cañadas Caldera wall (Fig. 4). The contact between the Upper and Lower Groups was identified as a regional erosional unconformity, representing a hiatus in central volcanic activity of up to 0.4 Ma. The Upper Group itself was subdivided into three major formations – Ucanca (1.57-1.07 Ma), Guajara (0.85-0.57 Ma) and Diego Hernández (0.37-0.188 Ma) – composed of phonolitic pyroclastic deposits and subordinate basaltic rocks (Table 1; Martí et al., 1994, 1997). The age span of the Diego Hernández formation has since been revised to 0.57-0.188 Ma. These formations were interpreted to represent three distinct cycles of explosive phonolitic volcanism sourced from vents within the area now occupied by the Las Cañadas Caldera. Intervening repose periods (lasting up to 0.23 Ma each) were characterized by widespread mafic volcanism sourced by both central and flank vents. Each cycle culminated in caldera collapse and the migration of the locus of volcanic activity and caldera formation to the northeast, resulting in the progressive enlargement of the Las Cañadas nested caldera complex (Martí et al., 1994).

In contrast to the areally restricted Lower Group sequences, the Upper Group formations are extensively exposed along the caldera wall (Fig. 4). Furthermore, a distal succession of phonolitic pyroclastic deposits – primarily non-welded ignimbrites and plinian fall deposits – accumulated on the lower slopes of the island (Booth, 1973; Wolff, 1985; Alonso, 1989; Bryan, 1998; Bryan et al., 1998a, 1998b, 2000; Brown et al., 2003; Brown and Branney 2004; Edgar 2003; Edgar et al. 2002; Pittari 2004; Pittari and Cas, 2004; Pittari et al., 2005, in review). Due to the stratigraphic complexity of the pyroclastic succession, the proximal-distal facies variations and the sparsity of medial exposure, only a few units, principally those in the uppermost Diego Hernández Fm., have been correlated between distal and proximal areas (Martí et al., 1994; Bryan et al., 1998a; Edgar et al., in review). Quaternary distal ash deposits found in Ocean Drilling Project (ODP) drill holes up to 156 km east and southeast of Tenerife have been correlated with the Upper Group by geochronological (Van den Bogaard, 1998) and geochemical (Rodehurst et al., 1998) criteria. Ashes that may correlate with the Diego Hernández Formation have also been found in the North Canary basin (Moreno et al., 2001). Although detailed (member-level) correlations have not yet been made, these studies may provide important constraints on the dispersal of tephra fallout and subaqueous pyroclastic density currents. Van den Bogaard (1998) found that plinian activity from the Las Cañadas Edifice during the Upper Group period was more or less continuous and suggested that the recognition of distinct formations by Martí et al. (1994) was an artefact of an incomplete onshore tephra record. Sumita et al. (2000) reported that 10-20% of the Pleistocene volcaniclastic layers (up to 450 total) in these drill holes could be interpreted as products of flank collapses and major plinian eruptions.

The petrology and geochemistry of the Upper Group phonolites and their plutonic xenoliths have been the subject of numerous studies (Wolff & Storey, 1983, 1984; Wolff, 1984, 1985a, b, 1987; Wolff & Palacz, 1989; Palacz & Wolff, 1989; Wolff & Toney, 1993; Wolff et al., 2000; Zafirilla, 2001; Bryan et al., 2002; Edgar et al., 2002 and in review, plus ongoing unpublished work) as have the contemporaneous rift and flank basalts (Neumann et al., 1999; Simonsen et al., 2000; Olin, 2003). These studies have shown that:

(a) The Upper Group is bimodal, with a paucity of compositions lying in the range 50 - 55% SiO₂ (phonotephrite to tephriphonolite).
(b) Phonolites have compositions near the 0.1 GPa water-saturated minimum, suggesting final evolution in magma chambers at 4 - 6 km depth, a finding supported by phase relations in cognate syenitic blocks and experimental petrology data (Andujar et al., 2005). Several of the larger phonolitic pyroclastic deposits are compositionally zoned in a manner consistent with fractionation of the observed phenocryst assemblage; however the same variations could be produced by mixing of phonolitic liquids that had evolved to different degrees, or been independently derived through melting of different protoliths, or different degrees of melting of a single protolith. Oxygen and lead isotope variations require at least a minor role for assimilation of country rocks into phonolitic magma.
(c) All UG phonolitic liquids appear to have been saturated with titanite (sphene), and their trace element inventories are strongly influenced by separation of titanite from melt. The geochemical hallmark of titanite crystallization is strong fractionation of Nb from Ta and of trivalent middle rare earths from both the heavy and light elements of the REE series, causing high Nb/Ta and (La,Yb)/(Sm,Gd) in the separated liquids. This chemical signature is not,

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however, diagnostic of any particular mechanism of crystal-liquid separation; it could arise through fractional crystallization, partial melting leaving a titanite-bearing residue, or partial to complete melting of a protolith that that experienced titanite fractionation during its own petrogenesis. Abundant syenitic lithics in many UG ignimbrites indicate that such protoliths exist within the Tenerife volcanic edifice.

(d) Two phonolite types have been repeatedly erupted during Upper Group time; one type, in which the titanite separation signature is strong, dominates the Diego Hernández Fm. but also occurs in the Guajara Fm. The second type, with a weak titanite signature, dominates the Ucanca and Guajara Fms. Both magma types were available for simultaneous eruption during Diego Hernández time. The two types may be derived from melting of different syenite protoliths, or be derived through fractional crystallization, involving different amounts of titanite, from a single parental magma type.

(e) Many eruptions, especially those of the Diego Hernández Fm., appear to have been triggered by mixing of phonolitic and mafic magmas. In some cases, three or four distinct magmas can be identified within the products of a single eruption. It is in these mingled rocks that the scarce phonotephrite to tephriphonolite compositions appear, rather than as discrete lavas or tephra layers.

(f) Within the Diego Hernández Fm., there are geochemical affinities between the most mafic component in a particular mingled pyroclastic deposit and the immediately underlying basaltic lava. This suggests that major explosive phonolitic eruptions are heralded by extrusion of basalt related to the eruption-triggering replenishment event.

(g) The isotopic compositions of Upper Group lavas and pyroclastics and contemporaneous basalts overlap those of all three OBS centers, indicating derivation from the same mantle sources during the whole history of Tenerife, and/or remelting of rock within the Tenerife edifice during Upper Group magma production.

It should be noted that most of this work has been focussed on the Guajara and Diego Hernández Formations, and that relatively little is known about the Ucanca Formation.

**Ucanca Formation (UF)**

The oldest subdivision of the Upper Group, the Ucanca Formation (UF) has a known age range of 1.57 to 1.07 Ma (Martí et al., 1997). It extends along the southern and southeastern sectors of the caldera wall and is thickest in the southwestern sector (Figs 2, 4). It comprises a sequence of peralkaline phonolitic lavas, welded and non-welded fall units, welded and non-welded ignimbrites, intraformational lithic breccias, pyroclastic surges and clastogenic lavas, with minor interbedded lavas and pyroclastics of basaltic composition. These deposits were produced by a succession of highly explosive plinian eruptions sourced from vents within the present-day caldera depression and often close to the caldera wall (Martí et al., 1994).

**Distal deposits:** Welded and non-welded ignimbrites and plinian fall deposits of Ucanca age have been found on the southwestern, southeastern and northern flanks of the island (Fig. 6; Martí et al., 1994; Füster et al., 1994; Huertas et al., 2002; Edgar 2003). Due to surface erosion and burial by younger deposits, the distal Ucanca Formation is poorly exposed except in the southwest (Füster et al., 1994). The most detailed stratigraphy for these deposits has been proposed by Huertas et al. (2002) but this work does not recognize the stratigraphic subdivisions (such as the Ucanca and Guajara formations) developed by Martí et al. (1994).
Guajara Formation (GF)

Extending along the southeastern sector of the caldera wall and unconformably overlying the Ucanca Formation is the Guajara Formation (GF), with age data spread between 0.85 and 0.57 Ma (Martí et al., 1994; Bryan et al., 1998a; Bryan, 1998). With a maximum thickness of 250 m at Pico de Guajara, the Guajara sequence represents a second cycle of highly explosive and largely phonolitic eruptions. It produced a similar range of deposits as the preceding Ucanca cycle. Preservation and outcrop of the stratigraphy of the Guajara Formation is however better than for the Ucanca Formation. Booth (1973) and Booth and Walker (unpublished data) were the first to attempt a stratigraphic subdivision of the Guajara Formation, followed by Alonso (1989). More recently Bryan (Bryan et al., 1998a; Bryan, 1998) and Middleton (in prep.) have considerably advanced understanding of the stratigraphy and volcanology of the Guajara Formation.

Distal deposits: The Guajara Formation is also distributed widely on the middle and lower southeastern slopes (Fig. 7; Martí et al., 1994; Bryan et al., 1998a; Edgar 2003), on the Tigaiga massif (Ibarrola et al., 1993) and offshore (Van den Bogaard, 1998), consistent with the magnitude and style of the eruptions. Although likely Guajara Formation correlates have been mapped in the north and northwest (in this study), by far the most extensive and highest quality exposure of this pyroclastic succession occurs in the south and southeast of the island, where it underlies the Diego Hernández Formation. Deposits away from the caldera are generally non-welded, although the Arico ignimbite is the only welded ignimbrite away from the caldera wall succession, and it is widely welded (Schmincke and Swanson 1967; Middleton in prep.). The Guajara Formation in the Bandas del Sur consists of large volume pyroclastic fallout, flow (ignimbrite) and minor surge deposits. The stratigraphic record contains fewer units than the overlying Diego Hernández Formation, and it is unclear if this truly reflects fewer explosive events, or whether, due its older age, erosion has stripped away more of the stratigraphy.

A major problem with understanding the Guajara Formation remains the difficulty of correlation of Guajara Formation age rocks preserved in the caldera wall with deposits on the flanks of Tenerife, especially on the Bandas del Sur, on the arid south side of Tenerife, where the exposure of the Guajara Formation stratigraphic units is good, but discontinuous. Thick sequences of both non-welded and welded rocks in the caldera wall in the section up to Pico de Guajara have not been uniquely correlated to specific units on the Bandas del Sur using geochemistry, facies or geochronology.
**Diego Hernández Formation (DHF)**

The third and youngest cycle of Upper Group phonolitic volcanism produced the Diego Hernández Formation, DHF (Araña, 1971; Mitjavila, 1990; Mitjavila & Villa, 1993; Martí et al., 1989, 1990, 1994; Edgar, 2003). The DHF is defined to include the products of all central phonolitic volcanism on Tenerife between 0.57 Ma (i.e. the end of the Guajara cycle) and 0.188 Ma. The mapped extent of the DHF is shown in Figure 6. Unlike the two older formations, the proximal DHF is restricted to the eastern sector of the caldera wall, where it infills a broad depression, the DH palaeovalley (Fig. 4). In terms of proximal facies, the DHF succession is distinguished from the two older formations by the scarcity of welded rocks; it is dominated by non-welded ignimbrites and minor plinian fall deposits (Martí et al., 1990, 1994; Edgar, 2003). The succession is capped by a coarse lithic lag breccia, rich in syenite fragments, which has been correlated with a widespread lithic-rich ignimbrite exposed on the lower slopes, the Abrigo ignimbrite (Alonso, 1989; Martí, 1994; Pittari et al., 2004; Pittari et al., 2005, in review). This deposit is thought to record a climactic caldera-forming eruption at 0.188 Ma (average of several $^{40}$Ar/$^{39}$Ar dates of Mitjavila & Villa, 1993; Nichols, 2001; Nichols et al., 2001; and Brown et al., 2003), completing the incremental growth of the Las Cañadas Caldera and bringing the Diego Hernández cycle to an end. With the exception of the Abrigo ignimbrite, the proximal and distal deposits of the DHF have until recently remained poorly correlated due to the lack of medial exposure, the complexity of the stratigraphy and unpredictable lateral facies variations (e.g. Bryan et al., 1998a). Detailed mapping, section logging and deposit characterisation by Edgar (2003) have led to detailed correlations becoming possible between the caldera wall succession and the Bandas del Sur DHF succession. On the Bandas del Sur, mapping by Edgar (2003) and detailed volcanological studies by Edgar (2003), Edgar et al. (2002, and in review), Brown et al. (2003), Brown and Branney (2004), Pittari (2004), Pittari and Cas (2004) and Pittari et al. (2005) have led to significantly improved understanding of the volcanic eruptions styles, dispersal processes and deposit characteristics on Tenerife.

**Distal deposits:** The Diego Hernandez Formation is also best preserved on the Bandas del Sur, where it consists of widespread, well-preserved successions of non-welded pyroclastic fallout, flow (ignimbrite) and co-ignimbrite surge deposits. Although some eruptions were relatively simple, with initial plinian fallout deposits followed by one or more ignimbrite units, others were extremely complex (e.g. Fasnia Member, Poris Member; Edgar et al. 2002; in review; Brown et al. 2003, Brown and Branney 2004), with numerous intra-
plinian ignimbrites interrupting thick plinian fallout deposits. The Fasnia Member contains seven such intra-plinian ignimbrites.

**Figure 8.** Distribution of the Diego Hernández Formation (From Edgar 2003).

**THE TEIDE-PICO VIEJO FORMATION (TPVF)**

The Teide-Pico Viejo Formation (TPVF) consists of the twin stratocones of Teide and Pico Viejo, numerous satellite vents and outflow lavas (Hausen, 1956; Araña, 1971; Ablay, 1997; Ablay et al., 1998; Ablay & Martí, 2000). With an altitude of 3718 m, the summit of Teide forms the highest point on the island and, indeed, in all of Spain. The TPVF postdates the Upper Group and the formation of the Las Cañadas Caldera (0.19 Ma), which it partially infills. It comprises a thick sequence of basanite lavas overlain by more evolved lava sequences, ranging from phonotephrite to phonolite (Table 1; Ablay & Martí, 2000). Drilling in the floor of the Las Cañadas Caldera has revealed TPVF lava sequences exceeding 500 m in thickness and the pre-TPVF caldera floor has not been intersected (Martí et al., 1994; Ablay & Martí, 2000). Many TPVF lavas also flowed north and west down the flanks of the Las Cañadas Edifice (Fig. 9). TPV activity is thought to have been sourced by two distinct phonolitic magma chambers with different depths and physical conditions, with activity alternating between Teide and Pico Viejo in the recent history of the complex (Ablay & Martí, 2000). Historic eruptions have occurred at Pico Viejo (1798), Teide (1492) and numerous satellite vents (Table 1). The largest explosive phonolitic eruption on Tenerife since the end of the DHF cycle occurred at Montaña Blanca (Fig. 20), a major satellite system of the TPV system, approximately 2000 years ago (Ablay et al., 1995). This subplinian eruption dispersed 0.08 km$^3$ of tephra (Ablay & Martí, 2000) over the eastern sector of the caldera and the proximal Dorsal Ridge. However, no TPVF eruptions have produced fall deposits of plinian dispersal and the youngest phonolitic pyroclastic deposits exposed on the lower flanks of the island belong to the DHF.

**THE DORSAL & SANTIAGO RIFTS**

A detailed study on the stratigraphy and structure of the Tenerife rift systems is found in Galindo (2005). The Dorsal Ridge (or Dorsal rift) is a major basaltic fissure-vent system extending along a SW-NE lineament from the eastern sector of the Las Cañadas Edifice towards the Anaga Massif (Fig. 20). Morphologically, it forms a high-relief crest or spine which becomes higher (2400 m, Montaña de Abreo) and broader to the southwest (here referred to as the “proximal” Dorsal Ridge). Despite its inclusion in several general studies on the geology and geochronology of Tenerife (e.g. Fuster et al., 1968; Ancochea et al., 1990), it remains poorly understood. The products of this system, known collectively as the Dorsal Series, include scoria cone complexes and thick lava sequences. The age data of Ancochea...
et al. (1990) were interpreted to indicate a period of rapid growth between 0.9 and 0.78 Ma, although activity on the Dorsal Ridge stretches back to at least 1.8 Ma (Feraud et al. 1985) and has continued to the present day. Historic activity includes the 1704-5 basaltic eruption of Volcán de Fasnia which occurred along an 11 km-long fissure. The interbedding of phonolitic pyroclastic deposits of the Diego Hernández Formation within the complex basaltic stratigraphy of the proximal Dorsal Ridge (described in this study) indicates the contemporaneous activity of the two adjacent magma systems (Table 1). Alternative names for the Dorsal Ridge include the Cordillera Dorsal (Fúster et al., 1968; Ancochea et al., 1990) and the Cumbre de Pedro Gil (Ridley, 1970a, 1971; Feraud et al., 1985).

Figure 9. Teide-Pico Viejo Formation (From Edgar 2003, adapted from Ablay & Martí, 2000).

The smaller Santiago rift (Ablay & Martí, 2000) forms a NW-trending zone of intense mafic to intermediate volcanism extending between Pico Viejo and the Teno Massif (Fig. 20). Like the Dorsal rift, it is a fissure-vent system formed by a large number of monogenetic cones and lava flows. Its age is poorly constrained (>0.3 Ma, Ablay & Martí, 2000) but it remains active to the present day; the eruption of Chinüero in 1909 is the most recent volcanic eruption on Tenerife. The Dorsal and Santiago rift zones have been interpreted as two arms of a three-branched rift system, a model which was used to support a hotspot origin for the Canary Islands (Carracedo, 1994). The third branch was identified as the broad zone of flank volcanism extending southward into the Bandas del Sur (see below) – here referred to as the Southern volcanic zone (after Ablay & Martí, 2000). However, this model was questioned by Martí et al. (1996) and a more recent interpretation views the Dorsal and Santiago rifts as part of a continuous axis of extension and volcanism which also underlies the active TPV stratovolcanic complex (Ablay & Martí, 2000).

Olin (2003) studied the petrology and geochemistry of Dorsal Ridge basalts emplaced as scoria cones, lavas and dykes at the intersection of the Dorsal Ridge with the Diego Hernández paleovalley. They are variable in composition and fall into several fields on the TAS diagram (Fig. 3). Of the 116 samples analyzed by Olin (2003), the majority (51%) are basanites with normative olivine ($ol$) >10%, 15% are tephrites ($ol$ <10%), 13% are alkali basalts, 11% hawaiites, 2% microbasalts and 1 nephelinite sample (<41% SiO$_2$). Major element variations are consistent with fractional crystallization of clinopyroxene (titanaugite), olivine and Fe-Ti oxides from basaltic magma, but with little role for plagioclase. Although some samples contain plagioclase phenocrysts, disequilibrium textures (e.g. sieve texture, embayments, zoned coronas with irregular cores and reaction rims) suggest that many were introduced into basaltic magma via mixing or contamination processes involving plagioclase-phyric magmas.
The trace element character of the Dorsal basalts is typical of HIMU-type ocean island basalts, with large ion lithophile elements such as Ba, Rb and K less enriched than Nb and Ta (Weaver, 1991). Nb/Ta (15-17) and Pb/Ce (0.02-0.03) values for most samples are generally consistent with other ocean island basalt (OIB) data (Hofmann et al., 1986; Weaver, 1991; Thirwall et al., 1997 and references therein). Higher values (>17 for Nb/Ta and > 0.035 for Pb/Ce) probably indicate contamination by phonolite or syenite, perhaps at extremely shallow levels; the country rock for last few hundreds of meters of magma transport to the surface was the glassy phonolitic pumice deposits of the Diego Hernández Formation. A general lack of heavy rare earth element depletion may reflect an ultimate origin in garnet-free shallow mantle.

FLANK VENTS

Flank Basalts

A large number of small basaltic volcanoes, mainly monogenetic strombolian cones, dot the flanks of the Las Cañadas Edifice (Fig. 20). These have been classified in various ways by different researchers. The nomenclature used here derives from Fúster et al. (1968), who identified three "series": Trachytic and Trachybasaltic Series, Basaltic Series III and Recent Basic Series (their Recent Acidic Series corresponds to the Teide-Pico Viejo Formation). The Trachytic and Trachybasaltic Series includes the Montaña de Guaza and Caldera del Rey (discussed separately below) as well as a number of lava flows of variable composition scattered around the island.

Figure 10. Distribution of flank basalts and Tijuela landslide. DRZ = Dorsal Ridge Rift Zone; SRZ = Santiago Rift Zone; SVZ = Southern Volcanic Zone. From Edgar (2003). Data sources: Fúster et al. (1968); García Moral (1989); Bryan (1998).

The more extensive Basaltic Series III was originally defined as postdating the Las Cañadas Edifice and predating the TPV Formation. However, they are now considered to predate or be contemporaneous with the DHF cycle of the Upper Group of the Las Cañadas Edifice (Table 1; Cycle 3 basalts of Bryan et al., 1998a). Series III basaltic scoria deposits and lava flows were erupted from a large number of strombolian vents, many of which were aligned along structural lineaments (rift systems). They are concentrated in the south, northwest and northeast of the island and have been interpreted as a major basaltic phase separating different cycles of central phonolitic volcanism (Martí et al., 1994, 1997; Bryan et al., 1998a). Studies of individual cones include that of the Montaña de Taco in the Isla Baja region by Alonso et al. (1992). Finally, the Recent Basic Series or Basaltic Series IV refers to the products of a number of widespread basaltic eruptions which occurred in prehistoric or historic times (Fúster et al., 1968; García Moral, 1989). Major volcanic fields arose in the Valle de San Lorenzo in the south, part of the Southern volcanic zone and on the Santiago Workshop on Calderas 16-22 October, 2005
ridge in the northwest. Geophysical surveys have also revealed ongoing submarine volcanism on the seafloor surrounding Tenerife (e.g. Ablay & Hürlimann, 2000; Krastel & Schmincke, 2002).

Montaña Guaza
Montaña Guaza (Buch, 1825; Fritsch & Reiss, 1868; Hausen, 1956; Fernández Santín & Nafría López, 1978; Fúster et al., 1994) is a steep-sided, trachytic-phonolitic lava dome complex situated in the western Bandas del Sur adjacent to the coastline (Fig. 7). Ancochea et al. (1990) obtained a K/Ar age of 0.67 Ma for one of the trachytic lava flows, indicating contemporaneity with the Guajara Formation of the Las Cañadas Edifice.

Caldera del Rey
Located 4 km NW of Montaña Guaza in the shadow of the Roque del Conde OBS massif is the Caldera del Rey (Fig. 6), a double-cratered phonolitic maar. The products of this eruption centre consist mainly of successions of pyroclastic surge and fall deposits (some rich in accretionary lapilli), characteristic of a phreatomagmatic eruption style (Paradas Herrero & Fernández Santín, 1984). They extend to a radius of approximately 4 km (Fúster et al., 1994), although Bryan et al. (1988a) attributed widespread ignimbrite and plinian fall deposits of the Diego Hernández Formation to this centre (the Caldera del Rey Pumice and Ignimbrite Members dispersed up to 20 km to the east). Paradas Herrero & Fernández Santín (1984) suggested that activity of the Caldera del Rey was contemporaneous with Basaltic Series III (i.e. post-Guajara Formation, pre- to syn-DHF) based on the stratigraphic relationships of plinian fall deposits which they believed were produced by the maar. This scheme was followed by Bryan et al. (1998a). However, Fúster et al. (1994) obtained a K/Ar date of 1.54 ± 0.28 Ma and Huertas et al. (2002) presented a more precise 40Ar/39Ar date of 1.13 ± 0.03 Ma. This is far older than Series III and is consistent with stratigraphic data which indicate contemporaneity with the Ucanca Formation (Table 1.1). These relationships place important constraints on the interpretation of the DHF stratigraphy.

MAJOR LANDSLIDES AND CALDERA FORMATION: ARE THEY RELATED?

Types of degradational processes
In addition to the usual sedimentary processes which slowly contribute to the erosion of landscapes, Tenerife retains evidence of having been subject to numerous gravitational collapse events which generated various mass flow phenomena. Furthermore, a long history of highly explosive and relatively large-volume pyroclastic eruptions must have resulted in some form of caldera collapse. Several large depressions bounded by scarps have generated long-running controversy among scientists as to their origin and evolution. The most notable of these are the Gúímar, La Orotava and Icod valleys, and the central Las Cañadas Caldera.

Gúímar & La Orotava valleys
The Gúímar and La Orotava valleys are 6-10 km-wide depressions open to the sea, situated on either side of the Dorsal Ridge in central Tenerife (Fig. 7). They are bounded by steep, linear scarps up to 600 m high and have relatively flat, gently sloping floors. Their genesis has been variously attributed to horst-and-graben style extensional faulting (Hausen, 1956), “trap-door” caldera collapse due to magma withdrawal (Ridley, 1971), progressive lateral expansion via erosional processes (Palacios, 1994), and giant landslides or debris avalanches (Bravo, 1962; Coello, 1973; Ancochea et al., 1990; Carracedo, 1994, 1999; Martí et al., 1997; Cantagrel et al., 1999; Ablay & Hürlimann, 2000; Masson et al., 2002). There is now a general consensus that large-scale landslide processes were instrumental in the formation of these major morphological features. Sonar surveys of the sea floor off the north coast of Tenerife have revealed debris avalanche deposits extending 100 km offshore with a total estimated volume of 1000 km³ (Watts & Masson, 1995, 2001).

The head region of the La Orotava valley is thought to be exposed in the eastern sector of the Las Cañadas caldera wall (Fig. 7; Bravo Bethencourt & Bravo, 1989; Ancochea et al., 1990; Martí et al., 1997; Cantagrel et al., 1999). The steep head scarp truncates the Las Pilas sector and the topographic depression produced (the DH palaeovalley) was subsequently infilled by successive eruptions of the Diego Hernández Formation and the proximal Dorsal Ridge. Radiometric dating of the youngest lava flows truncated by the valley
walls (at the top of Las Pilas and the Tigaiga Massif) and the oldest infill deposits (in the DH palaeovalley) have constrained the age of valley formation to between 0.69 and 0.54 Ma (Ancochea et al., 1990; Cantagrel et al., 1999). Since the climactic eruption of the Granadilla Member at the end of the Guajara Formation (0.57 Ma; Bryan et al., 1998a) did not leave pyroclastic deposits in the DH sector of the caldera wall or elsewhere in the La Orotava infill, valley formation is further constrained to the period 0.57-0.54 Ma (Table 1). This supports the recent proposal of a genetic link between climactic caldera collapse at the end of major cycles of explosive phonolitic volcanism and major landslide events (Martí et al., 1996, 1997). The volume of the submarine La Orotava landslide deposit has been estimated as 80 km$^3$ by Ablay & Hürlimann (2000), who proposed two closely spaced lateral collapse events to account for the seaward narrowing of the valley.

The Güímar valley is younger than 0.83 Ma (Ancochea et al., 1990) but is older than the La Orotava valley, since its infill includes Guajara Formation deposits (Martí et al., 1997). The timing of this event corresponds to a period of rapid growth of the Dorsal Ridge (between 0.9 and 0.78 Ma, Ancochea et al., 1990), which may have oversteepened the slopes and promoted gravitational instability. The submarine debris avalanche deposit associated with the Güímar valley has been mapped in sonar surveys in the channel which separates Tenerife and Gran Canaria (Teide Group, 1997; Krastel et al., 2001, Krastel & Schmincke, 2002). It has also been identified in Ocean Drilling Program (ODP) drill holes with a thickness of 4 m measured 156 km off the coast (Sumita et al., 2000) and has an estimated minimum volume of ~120 km$^3$ (Krastel et al., 2001).

**Icod valley**

The Icod valley is another scarp-bounded depression open to the sea and thought to have been formed by landslide processes (Ancochea et al., 1990; Martí et al., 1997; Carracedo, 1999; Cantagrel et al., 1999; Watts & Masson, 2001). It is located on the northern slopes, separated from the La Orotava valley by the Tigaiga Massif, a high-relief remnant of the Las Cañadas Edifice (Fig. 20). The Icod valley has been significantly infilled by lavas of the TPV Formation and the position of its headwall is uncertain. No Upper Group deposits have been found within it, and its formation therefore postdates the DHF. Martí et al. (1997) proposed that climactic caldera collapse associated with eruption of the Abrigo Member at the end of the Diego Hernández Formation (0.20 Ma) triggered the Icod landslide event. On the basis of sonar surveys of the submarine Icod landslide deposit, Watts & Masson (2001) identified two flow lobes, Icod I and Icod II, thought to have been generated by a single, complex lateral collapse event. Ablay & Hürlimann (2000) inferred a deposit volume of 80 km$^3$ based on a separate geophysical survey of the seafloor.

**Other landslide events**

Recent studies have also presented evidence that landslides also occurred during the growth of the Old Basaltic shield complex, affecting the north flanks of the Anaga and Teno massifs at ~6 and 5.6 Ma respectively (Cantagrel et al., 1999; Ablay & Hürlimann, 2000; Masson et al., 2002; Walter & Schmincke, 2002). Another generation of "old post-shield landslides" was identified by Ablay & Hürlimann (2000) and interpreted as products of major north-directed failure of the LCE and proto-Dorsal Ridge at ~3 Ma to form a large amphitheatre, now buried. These authors also identified a distinct landslide event affecting the northeastern flank of the Dorsal Ridge. The age of this East Dorsal landslide is poorly constrained (<0.56 Ma) but the submarine deposit has been mapped to the north of Tenerife and has an inferred volume of 100 km$^3$. Finally, a poorly documented submarine debris avalanche deposit has been mapped off the southern coast of Tenerife. The Las Bandas del Sur deposit (Krstel et al., 2001) is thought to be younger than 2 Ma. It is considerably smaller than the other known landslide deposits (<25 km$^3$) and no onshore head scarp has been identified (Krstel et al., 2001).

**Las Cañadas Caldera**

The Las Cañadas Caldera (LCC) has often been described as the most spectacular volcanic landform in the Canary Islands and it has aroused the interest of geologists since the early nineteenth century (e.g. Humboldt, 1814; Buch, 1825). This major summit depression of the Las Cañadas Edifice is roughly elliptical and measures 17 x 9 km, elongated along a NE-SW-oriented major axis (Fig. 4). The bounding escarpment is continuously exposed along
the southern and eastern margins, with a maximum rim elevation of 2717 m at Pico de Guajara and a maximum (present-day) depth of 600 m. The isolated northern sector of La Fortaleza (part of the Tigaiga massif) is situated between the northeastern gap of El Portillo and a much larger northwestern gap (Fig. 4). El Portillo is thought to have formed during the La Orotava landslide event and has since been partially infilled by Dorsal Ridge strombolian activity and by lava flows of the TPV Formation. The “missing” northwestern wall has been variously explained, as discussed below. The post-caldera growth of the TPV stratovolcano complex has led to the burial of the caldera floor by thick (>500 m) lava sequences (Martí et al., 1994; Ablay and Martí, 2000), with the present surface at a mean altitude of 2000 m. The original depth of the caldera must therefore have exceeded 1000 m and the missing volume has been estimated at >140 km² (Martí et al., 1994). The spur of the Roques de García is conventionally used to divide the caldera into eastern and western parts (Hausen, 1956; Martí et al., 1994).

The origin of the LCC has been the most contentious issue in the geological investigation of Tenerife. Early ideas expressed by Buch (1825) envisioned upheaval followed by collapse of the summit (his “crater of elevation” concept). Charles Lyell visited the island in 1853-54 and proposed an erosional origin by the action of running water (Lyell, 1855). Fritsch & Reiss (1868) proposed a combination of erosion and great explosions. Gagel (1910) likened the explosive origin of the caldera to the 1883 eruption of Krakatoa (Indonesia), while Friedländer (1915) and Fernández Navarro (1916) suggested collapse similar to that of Somma Vesuvius. In the second half of the twentieth century, two competing models developed and are yet to be reconciled, one emphasizing landslide processes (Bravo, 1962; MacFarlane & Ridley, 1968; Navarro & Coello, 1989; Ancochea et al., 1990, 1999; Carracedo, 1994, 1999; Cantagrel et al., 1999) and the other emphasizing caldera collapse (Hausen, 1956, 1961; Füster et al., 1968; Araña, 1971; Ridley, 1971; Booth, 1973; Martí et al., 1994, 1996, 1997; Bryan et al., 1998a, 2000; Martí & Gudmundsson, 2000). These two models are summarized briefly here.

The landslide model proposes a similar genesis for the LCC as is generally accepted for the Güímar, La Orotava and Icod valleys: one or more large-scale seaward lateral collapse events generating debris avalanches and remnant scarp-bounded depressions. The main arguments put forward to support this model are as follows:

- Numerous tunnels (Spanish: galerías) have been excavated into the flanks of the island for the purpose of extracting drinking water from aquifers. Investigations of the subsurface geology exposed in the galerías on the northern slopes (Bravo, 1962; Coello, 1973; Navarro & Coello, 1989; Coello & Bravo, 1989) encountered a widespread chaotic breccia up to 300 m thick, dipping gently towards the coast. This “fanglomerado” deposit (Bravo, 1962) consists of subangular to rounded clasts of variably altered basalt, trachyte, phonolite and minor gabbro and carbonized wood fragments in a finer matrix. It was originally interpreted as the widely dispersed, explosive product of an early destructive phase of the Las Cañadas Edifice, which was buried by lavas and subsequently acted as a detachment for large-scale landslide events (Bravo, 1962; Coello, 1973; Carracedo, 1994). Later workers reinterpreted the “fanglomerado” facies as the deposits of major debris avalanches (Navarro & Coello, 1989; Ancochea et al., 1999; Cantagrel et al., 1999).

- The absence of a caldera wall in the northwest is assumed to support the idea that there never was an enclosed summit depression, as would be expected to result from caldera collapse into a magma chamber. Rather, the existing “caldera” wall is the head wall for one or more major north-directed flank failures (Coello, 1973; Ancochea et al., 1999; Cantagrel et al., 1999). Remnants of a northern caldera wall have not been encountered in the galerías of the upper Icod valley (Coello, 1973) and therefore they have not simply been buried by the subsequent eruptions of the TPVF. Furthermore, the scalloped horseshoe shape of the existing caldera wall has been claimed to bear a striking resemblance to the La Orotava and Güímar valleys (Carracedo, 1994) and to the head scars of major landslides (Cantagrel et al., 1999; Watts & Masson, 2001).

- Finally, sonar surveys of the sea floor north of Tenerife (Watts & Masson, 1995, 2001) have yielded evidence for multiple north-directed landslide events related to the formation of the La Orotava and Icod valleys. Many authors have equated the Las Cañadas caldera wall with the head scarp of the Icod landslide event, thus denying the significance of vertical caldera collapse in forming the summit depression. They argue that the estimated volume of submarine landslide deposits (1000 km³) cannot be accounted for by the La
Orotava and Icod valleys alone and thus a greater portion of the Las Cañadas Edifice must have been involved (Cantagrel et al., 1999).

The caldera collapse model identifies the primary agent in the origin of the LCC as vertical collapse of the roof of a magma chamber following the evacuation of magma during large-scale explosive eruptions. The growth of the caldera is assumed to have been incremental, progressing from southwest to northeast through time (Martí et al., 1994). Proponents of this model cite several lines of supporting evidence:

- The most obvious evidence for caldera collapse is the thick Upper Group pyroclastic succession, representing over one million years of cyclical explosive phonolitic volcanism. Upper Group deposits display dispersal patterns and facies characteristics indicating repeated large plinian eruptions from source vents located within the Las Cañadas Caldera (Booth, 1973; Alonso, 1989; Martí et al., 1994; Bryan et al., 1998a, 2000). The total volume of Upper Group pyroclastic material has been estimated to exceed 130 km$^3$ (DRE) by Martí et al. (1994), who noted that this compares well with the inferred volume of the original (pre-TPVF) depression (>140 km$^3$). Upper Group deposits exposed in the caldera wall, including thick welded fall deposits, clastogenic lavas and co-ignimbrite lag breccias, are typical of proximal facies of caldera-forming eruptions (e.g. Druitt & Sparks, 1982; Walker, 1985; Cas and Wright, 1987). The abundance of ignimbrites, many of which are lithic-rich, in the Upper Group stratigraphy is also favorable to a caldera collapse hypothesis (Araña, 1971).

- The morphology of the surviving caldera wall is similar to that of other collapse calderas, its scalloped form being attributed to the overlap of multiple circular vertical collapse structures (Martí et al., 1994). The isolated northern sector of La Fortaleza, with its south-facing escarpment, is difficult to reconcile with north-directed sector collapse (Bryan, 1998a).

- Various structures observed in the caldera wall, including radial phonolitic dykes, cone sheets, sills, reverse radial faults and concentric normal faults (Fúster et al., 1968; Araña, 1971; Martí et al., 1994; Martí & Gudmundsson, 2000), support repeated episodes of magma chamber deflation and caldera collapse. Experimental and numerical modelling of this process explains the migration of the magma chamber following climactic caldera collapse episodes of each cycle by changes in the local stress field (Martí & Gudmundsson, 2000).

- Detailed volcanological studies of the Granadilla Member (0.57 Ma), the youngest eruptive unit in the Guajara Formation, have demonstrated a significant eruptive volume (>10.2 km$^3$ according to Bryan et al., 2000) and lithic breccia facies consistent with caldera collapse (Booth, 1973; Bryan et al., 1998a, 2000).

- Proponents of the caldera collapse model have never denied the importance of landslide processes in modifying the northern flanks of the Las Cañadas Edifice. Indeed, a post-caldera landslide event (associated with formation of the Icod valley) is necessary to account for the absence of the inferred northwestern caldera wall (Booth, 1973; Martí et al., 1994, 1997). Furthermore, Martí et al. (1997) suggested that major vertical collapse events may have triggered the landslides which formed the La Orotava and Icod valleys. The available age data indicate that the formation of these valleys coincides with the final eruptions of the Guajara and Diego Hernández phonolitic cycles respectively. Thus, the various lines of evidence used to support northward lateral collapse events (the “fanglomerado” deposit, the sonar evidence of offshore debris avalanche deposits, the recognized instability of oceanic island volcanoes, etc.) are not inconsistent with the caldera collapse model.

In summary, the key point is whether the present wall of the Las Cañadas Caldera is the head scarp produced by north-directed landslides or the surviving portion of an originally enclosed depression formed by multiple caldera collapse events. In the latter case, the missing northwestern wall was removed by a post-caldera northward landslide (Icod), perhaps triggered by the final phase of caldera collapse (Martí et al., 1997). Alternatively, the Icod landslide event may have triggered the climactic caldera-forming eruption by decompression of the magma chamber (e.g. Ancochea et al., 1999; Scarth & Tanguy, 2001; Huertas et al., 2002). Major episodes of sea-level fall also coincide with post-shield landslide events, suggesting another possible trigger for lateral collapse (Ablay & Hürliemann, 2000). Other points of contention relate to the timing of constructive and destructive events in the history of the Las Cañadas Edifice and to the most appropriate volcanostratigraphic framework.

In the past decade, the two competing models have been updated and elaborated. To a certain extent, there is growing recognition that the complex history of the Las Cañadas...
Edifice and its summit caldera is the result of a variety of processes which include constructional phases, caldera collapse events, landslide events and ongoing surface erosion. For example, the most recent versions of the landslide model (Ancochea et al., 1999; Cantagrel et al., 1999; Huertas et al., 2002) now recognize multiple caldera collapse events associated with the eruption of large volumes of pyroclastic material. However, they claim that these vertical collapse episodes were relatively small and left no surviving structures. The caldera wall is still interpreted as the head scarp of at least three landslide events – Tigaiga (>2.3 Ma), Roques de García (0.6-0.7 Ma) and Icod (<0.15 Ma) – which generated separate breccia horizons observed in the galerías. In a separate geophysical study, Aubert & Kieffer (1998) Pous et al (2002) provide piezometric data which they claim supports the notion of vertical caldera collapse for the southwestern and central sectors of the caldera but a landslide (“graben sector slipping”) origin for the northeastern sector.

**Stratigraphy of the Upper Group**

Significant advances in understanding the stratigraphy of the Upper Group have occurred over the last 15 years, especially with regard to the Guajara and Diego Hernandez formations, which are the best exposed (Table 2). Combined with field litho-stratigraphic mapping, geochronological studies have provided significant clarification of the chronology of eruption events, although the common presence of excess argon in samples continues to produce discrepant ages and relationships between lithostratigraphic and geochronological ages. This is exacerbated when samples are collected randomly in the field without accompanying detailed understanding of the lithostratigraphy, rendering some interpretations nonsense.

The evolving chronstratigraphy is, however, a good basis to begin understanding eruption frequency, eruption trends, magma chamber dynamics, timescales for magma genesis, etc.
Table 2. Comparison of the stratigraphic schemes that have evolved for the Upper Group. In this field guide we adopt the scheme in the right hand column, based on the most recent stratigraphic studies of Edgar (2003) and Middleton (in prep).

**VOLCANOLOGY OF THE UPPER GROUP**

The occurrence of dual magma systems (basalt, phonolite) has produced a complex array of eruption activity and styles, from basaltic fire fountaining, lava effusion and explosive Strombolian style activity to phonolitic fire fountaining, lava effusion and highly explosive plinian scale explosive eruptions. Phreatomagmatic activity fuelled by both surface water (crater lakes?) and hydrothermal systems have been common. Explosive eruptions have therefore been variously triggered by magmatic exsolution and vesiculation, phreatomagmatic explosive interaction and hydrothermal explosions. Injection of basalt magmas into phonolite-syenite magma chambers have also been significant eruption triggers.

Dispersal processes have included widespread plinian scale fallout, pyroclastic flow and pyroclastic surge (e.g. Bryan 1998, Edgar 2003, Edgar et al., 2002). Many large fallout deposits have been mostly dispersed off-shore, making eruption volume calculations more...
difficult. Nonetheless, some of the biggest fallout events have preserved thicknesses of over 50 metres near vent, and almost 10 metres, even at 15 kms from vent (e.g. Granadilla Member fall deposit, Bryan et al. 2000; Fasnia Member fall deposits, Edgar 2003; in prep). Many near vent fallout deposits are welded, suggesting low fountain style eruptions (e.g. Soriano et al. 2002). Some of these welded fall deposits became rheomorphic. Many near vent fall deposits, especially the welded ones, may have no distal equivalent, and conversely, many of the distal fall units, even very thick ones, may have no preserved near vent deposits because of the effects of erosion on unwelded fall deposits on the steep upper slopes. Although many pyroclastic fall deposits are known on the vegetated north side of the island, it appears that perhaps the majority were dispersed to the south by strong prevailing northerly winds. Attempts to calculate the volumes of fallout deposits using the Pyle (1989), Fierstein and Nathenson (1992), Carey et al. (1995), Bonadonna et al. (1998) and Bonadonna and Houghton (2005) methods, have produced volume estimates of > 5 km$^3$ for some fallout deposits (e.g. Granadilla Member fallout, Bryan et al. 2000; Fasnia Member fallout deposits, Edgar 2003; Edgar et al., in review), indicating extremely intense and large volume eruptions. Estimates of plinian column heights are up to 25 kms for the Granadilla Pumice fall deposit (Edgar 2003) and up to 30 kms for the Fasnia Member fallout deposits (Edgar 2003).

**Pyroclastic flows** were largely pumice and ash flows, although near vent spatter flows may also be represented in the near vent record. All distal pyroclastic flow (ignimbrite) deposits with the exception of the Arico Ignimbrite, are non-welded, which is surprising given the phonolithic magma composition, and the abundance of welded rocks near vent. The non-welded nature of these distal non-welded ignimbrites is presumably related to high original eruption columns before the collapse of the eruption columns to produce pyroclastic flows, inflow turbulence related to high edifice slopes and to interaction between flows and irregular topography, relatively small volumes, and high lithic clast content. Ignimbrite deposits are not well preserved on the upper slopes of the edifice, probably due to a combination of post-depositional erosion on the upper steep slopes, and/or “by-passing” (i.e. no deposition) of the upper steep slopes by pyroclastic flows. Ignimbrites are commonly very thick near the coast, indicating that depocentres lie either close to the shore or off-shore. This in turn indicates that calculated preserved volumes of ignimbrites are several times smaller than actual original volumes. In addition, there is no indication of what volumes of the erupted ignimbrites are buried in the calderas under the young Teide and Pico Viejo lavas. It is therefore possible for many ignimbrites, that original volumes were at least 5 to 10 times greater than the preserved on-land volumes, which in some cases are also > 5kms$^3$. The Abrigo Ignimbrite has a preserved on-land volume of 1.8 kms$^3$, and an estimated original deposit volume of between 15 and 20 kms$^3$ because it is thickest close to the coast (Pittari 2004; Pittari et al. in review).

Many ignimbrites are very lithic clast rich. Many of the lithic clasts are hydrothermally altered phonolite and basalt clasts, as well as a significant volume of syenite clasts, which were clearly sub-surface derived. That is, there are very large volumes of accessory lithic clasts in the ignimbrites. This suggests that very large volumes of upper crustal rock material were excavated during major eruptions, and perhaps several kms3 in some cases (e.g. 3 – 4 km$^3$ in the Abrigo Member ignimbrites, Pittari 2004). Together with the inferred DRE erupted magma volume calculations, it is difficult to escape the conclusion that the erupted volumes of magma and crustal rock debris were high enough to have caused calderas to form or further incrementally subside during many of the explosive phonolitic eruptions on Tenerife.

The current Las Canadas Caldera Complex has a WSW-ENE length of 17 kms. It is thought to consist of at least three nested, overlapping calderas, the Ucanca (SW and oldest), Guajara (middle) and the Diego Hernandez (NE and youngest), each of which was probably 6 to 8 or more kms in diameter at the end of collapse. It is intriguing however, that the oldest caldera (Ucanca), has not been filled with the products of the Guajara and Diego Hernandez caldera forming eruptions, and that the Guajara caldera was not filled with the products of the Diego Hernandez caldera forming eruption. This suggests that magma chambers associated with the younger calderas, may have extended under the older calderas, and that during the younger caldera forming eruptions, lateral magma withdrawal from under the older calderas caused renewed secondary subsidence in the older calderas, which even the young lavas from Teide and Pico Viejo have not been able to fill in. Although the erupted volumes were substantial, and in many cases big enough to cause discrete caldera collapse, it is likely that caldera subsidence was incremental and multi-stage during many eruptions, given that no erupted volume appears to have been huge (many tens or more kms$^3$). Given that the Diego Hernandez Caldera is still an unfilled deep hole, it appears that much of its final subsidence
occurred after the Abrigo eruption. It is also possible that some of the large fallout forming eruptions or phases of eruptions were big enough to cause some caldera subsidence.

**DISTRIBUTION OF UPPER GROUP ERUPTION PRODUCTS**

The distribution of the eruption products of Upper Group volcanism was controlled by dispersal processes, wind directions, topographic controls, post-depositional erosion, and in terms of exposure by the density of the vegetation cover. The arid southern slopes of Tenerife preserve by far the most complete and continuously exposed succession. As discussed above, the caldera wall also has good exposure of the products of all three explosive eruption cycles of the Upper Group, and parts of the Lower Group as well. Lack of completeness of the stratigraphy of the caldera wall succession, requires care in terms of identifying possible correlatives in the Bandas del Sur succession.
FIELD TRIP GUIDE

PICO DE GUAJARA TREK

The trek up Pico de Guajara is a steep climb of nearly 800 m, that including geology and recovery stops, will take about 4 hours. Take plenty of water, go slowly, and stop to use sunscreen frequently. From the Parador Hotel drive to the main road, turn left, drive past the turn off into another Parador car park on the left, and then turn left again about 600 m along the main road, onto the National Park road that runs to the east along the foot of the caldera wall. Drive about 1 km to a locked gate. Park at the locked gate and walk back to take the track that begins about 100 m before the gate, heading left (south) up the slope towards the caldera wall. Take care to stick to the correct track, which is often marked by cairns of rock and painted arrows on boulders. In places there is a maze of criss-crossing paths.

The track winds its way up the face of Pico de Guajara through five distinct lithofacies (Fig. G1):

- hydrothermally altered, stratified pumice cone deposits and associated dykes of the Lower Group (base), overlain by
- welded rocks of the Ucanca Formation, overlain by
- laterally discontinuous basalt lava (Series II), overlain by
- a succession of non-welded fallout, pyroclastic flow and associated surge deposits at the base of the Guajara Formation, overlain gradationally by
- welded rocks of the Guajara Formation,

Stop 1: Ascent of the north face of Pico de Guajara

The lower part of this trek begins in hydrothermally altered, stratified pumice cone successions thought to belong to the Older Group (> 1.8 Ma). These have not been well-studied. At these lower levels there are also dykes of various ages.
Above the Older Group rocks, the path crosses and climbs up through welded phonolite rocks of the Ucanca Formation (1.57 – 1.07 Ma). In outcrop the units are defined by rough breaks, and vary from massive and sparsely porphyritic, to clastogenic with welding textures. Again the details of these rocks have not been well documented, but their origins, as well as of other welded rocks in the Guajara Formation have been considered by Soriano et al. (1993) and Soriano et al. (in prep.). They are considered to be welded fallout deposits, and rheomorphic lava-like welded fall deposits, although there could also be some true lavas in the succession (see below for discussion of characteristics). The first cliff exposure in the Ucanca Formation shows a weathered, non-welded, pumice-rich ignimbrite (>5 m thick) overlain and cross cut by a sequence of highly altered, stratified units with undulating bedforms. This in turn is overlain by a thin (35 cm) lithic-rich ignimbrite with a stratified base that becomes intensely welded towards the top. It contains lithic clasts up to 53 cm in diameter, some of which have created impact sags in the stratified units below. The units above include a series of thick non-welded to welded fall deposits (from 2 to 20 m thick).

The Series II basalt lava outcrops discontinuously above the welded Ucanca Formation rocks in the caldera wall, and we will clamber down to an exposure if time permits.

Above the basalt lava, the base of the Guajara Formation is marked by a succession of coarse near-vent fall deposits and thin non-welded ignimbrite and associated surge deposits exposed in a steep gully. At Pico de Guajara, the Guajara Formation is ~250 m thick (Martí et al., 1994) and is composed of non-welded to welded fall deposits and small volume ignimbrites with interbedded pumice-rich surge deposits. The majority of the stratigraphic succession dips gently south away from the caldera although some units do appear to be draped inwards towards the caldera. Previous stratigraphic interpretations of the Guajara products at Pico de Guajara by Martí et al. (1994) and Bryan (1998) suggested that both thick sequences of welded ignimbrite and fall deposits were exposed in the wall. However, studies by Soriano et al. (2003), Soriano et al. (in preparation) and Middleton (in preparation) have found that the majority of welded units in the caldera wall are in fact fall deposits. Non-welded to welded ignimbrites do appear in the stratigraphy, but they are a minor component.

The basal non-welded deposits include finely stratified ash fall deposits, lithic-rich and pumice-rich fall deposits, massive pumice-rich ignimbrites with abundant large lithic clasts and fine grained, pumice-rich, low angle cross-stratified surge deposits. This lower succession could correlate with the basal succession of the Guajara Formation on the Bandas del Sur (Units A to E of Booth and Walker, unpublished data).

The track then climbs to the base of the very thick welded unit that forms the upper face of Guajara Peak. The base of this unit is non-welded, but includes spatter agglutinate beds, non-welded pumice clast rich and lithic clast rich beds, and alternations of non-welded and welded fallout deposits, passing upwards into a very thick, up to 30 m thick densely welded, lava-like unit (Fig. G2). In the transitional zone, from non-welded to welded, there are highly vesiculated pumice clasts, highly flattened obsidian fiamme, obsidian spatter, and lithic clasts. The fiamme clasts and spatter are interpreted to have been deposited from a low in a fountain-like eruption column, whereas non-collapsed, vesiculated clasts are thought to have been transported high into the column where they cooled before deposition. The welded fall units are characterised by a well-developed eutaxitic texture defined by large (2 to 24 cm in length), flattened obsidian or pale green devitrified glass clasts. They also lack a fine-grained matrix, and often grade up from coarse, pumice-rich, non-welded fall deposits. The welded fall deposits also commonly display rheomorphic textures, which gives them a lava-like appearance. Confusion between phonolitic lavas and rheomorphic welded fall deposits is common, however there are certain characteristics of a welded fall deposits that separate them from phonolitic lavas. Welded fall deposits grade laterally and vertically into non-welded deposits, mantle topography, contain lithic clasts, do not have autobreccias at their margins, unless they have become rheomorphic, in which case autobrecciation can occur, and can form deposits meters to tens of meters thick. Phonolitic lavas range from 2 to 10 m thick, have autobreccias at their margins, form valley ponded units and do not contain lithic clasts (Martí et al., 1995).
Figure G2: Generalised stratigraphy and facies of the upper welded unit, Pico de Guajara (From Bryan, 1998).
The track follows a bed of lithic rich planar stratified to low angle cross-stratified tuff, interpreted as a base surge deposit and also an interval of obsidian, which is locally altering and devitrifying. Bryan (1998) considered that this upper welded sequence possibly correlated with the 0.570 Ma Granadilla Member of the Bandas del Sur succession, but subtle variations in geochemistry, difficulties in establishing age correlations make this uncertain. It is possible that the Guajara Peak succession represents, like many of the units in the caldera wall, locally erupted and deposited units erupted from fire fountain eruptions, perhaps from fissure vents along or close to the caldera wall, whereas the widespread non-welded deposits of the Bandas del Sur represent the widely dispersed products of high plinian column forming, central or point source vent eruptions.

**Stop 2: The summit of Pico de Guajara**

The summit of Guajara Peak offers breath-taking views of the Las Cañadas Caldera Complex, with the Ucanca Caldera to the west (left), separated from the Guajara Caldera in the centre by the Roques de Garcia ridge (Older Group rocks), and the Diego Hernandez Caldera to the east. In addition the line of Pico Viejo stratovolcano (left), El Teide stratovolcano, with the dark bands of the 1492 a’a lavas down its sides, and the Montana Blanca lava dome complex to the right produce a spectacular volcanic landscape.

**Stop 3: Descent of caldera wall via Barranco del Rio**

After lunch at Pico de Guajara, we will walk along the caldera rim to the east, initially on top of the preserved top of the Guajara welded unit. In places this unit is well layered (flow banded) and the layering shows well-defined ramping of the layering near the margins of the unit. Bryan (1998) interpreted this unit as a rheomorphic, lava-like, welded fall deposit. Elsewhere along the crater rim other welded units, some draping the palaeo-rim, are visible. These also appear to be localised, although other welded (rheomorphic) units extend several kilometres down the outer slope of the edifice towards Vilaflor.

Once we climb down through the Guajara welded unit, the path crosses an area covered with coarse, non-welded pumice deposits of presumed Guajara Formation age. The path arrives at Barranco del Rio, which is a major, long lived canyon that almost certainly acted as an incised bedrock channel for pyroclastic flows from the caldera to the Bandas del Sur. In the walls of Barranco del Rio both the Ucanca and Guajara Formations are again exposed (Fig. G3), with the lower part of the Guajara Formation again being largely non-welded above Ucanca Formation welded rocks, and capped by another welded unit. If time permits we will stop briefly at outcrops of bedded non-welded to welded fallout, flow and limited reworked pumiceous sediments exposed along a track that leads south towards Vilaflor along the western rim of Barranco del Rio.

![Figure G3: General view and stratigraphy in the eastern face of Barranco del Rio near its head.](image)
From here we will walk back to the caldera rim and take a track down the caldera wall, and then walk back along the base of the caldera wall, along the edge of the young Teide a’a and block lava flow field that is filling the caldera floor.

**BANDAS DEL SUR**

**DRIVE FROM THE LAS CAÑADAS CALDERA AND OVERVIEW OF THE BANDAS DEL SUR**

The bus will exit the Las Cañadas caldera via the Boca de Tauce pass and descend the southern slopes of the Las Cañadas edifice, passing by exposure of welded pyroclastic deposits and lavas of the Upper and Lower Groups. Basaltic lavas and scoria cones contemporaneous with phonolitic products of the Upper Group become extensive below Vilaflor. The extensive arid coastal plain, the Bandas del Sur, which can viewed during the descent, is comprised of a sequence of mostly non-welded (with the excetion of the Arico ignimbrite) phonolitic pyroclastic deposits of the Ucanca, Guajara and Diego Hernández formations interbedded with flank basaltic lavas and pyroclastic deposits.

Over the course of the day at two localities, Tajao and Poris de Abona, we will examine many of the major phonolitic pyroclastic eruption packages (members) from both the Guajara and Diego Hernández eruptive cycles (formations) of the Upper Group. The major units we will observe are summarised below.

**Diego Hernández Formation (0.37 – 0.188 Ma)**

The Diego Hernández Formation consists eight major complex eruption packages/members (Fortaleza, Roque, Aldea, Fasnia, Poris, Araf o, Caleta and Abrigo Members) consisting of interbedded plinian fall, surge and flow deposits, a sequence of plinian fall deposits beneath the Abrigo Member (Cruz Sequence), and several minor members. Many of the larger members are exposed radially around the Las Cañadas edifice. The major eruption packages we will see are described below.

**Abrigo Member:** The Abrigo Member (Ar/Ar: 188 ka; Edgar, 2003; Pittari, 2004; Edgar et al., in review; Prittari et al., in review) consists solely of a widespread sheet-like ignimbrite deposit up to 25 m thick exposed radially on the coastal plains all around the Las Cañadas edifice, although a possible 30 cm thick plinian fall deposit is locally exposed near Poris de Abona. An estimated minimum onshore deposit volume of 1.8 km$^3$ has been proposed although, given constraints on volume calculations, this is likely to be as great as 15 to 20 km$^3$. On the Bandas del Sur there are two lithic-rich ignimbrite depositional units (Sur-A and Sur-C units) separated by a 1-4 cm thick co-ignimbrite ash layer (Sur-B unit). The ignimbrite units display distinct valley ponded (thick, massive, coarse-grained) to ignimbrite veneer (thin, stratified, finer-grained) transitions over the radial drainage pattern, and feature several local substrate-derived and vent-derived lithic concentration zones, pumice concentration zones and massive pumice-rich ignimbrite lobes. The Abrigo eruption was the climactic phase of the Diego Hernández magmatic cycle and was dominated by pyroclastic flows fed by a collapsing gas-pyroclast fountain contemporaneous with caldera collapse.

**Cruz Sequence:** Up to at least 14 pumice lapilli plinian fall deposits occur below the Abrigo Member on the Bandas del Sur, Guimar Valley and Orotava Valley. The oldest, the Batista Member (Ar/Ar: 232 ka; Edgar, 2003; Edgar et al., in review), also consists of a basal zone of interbedded lithic-rich surge horizons and normally graded plinian fallout horizons. This sequence represents a period of successive plinian eruptions with sustained high eruption columns, which were erupted from the same magma chamber as the Abrigo Member over the preceding ~44,000 years.

**Caleta Member:** The Caleta Member (Ar/Ar: 223 ka; Brown et al., 2003; Edgar, 2003; Edgar, in review) is exposed mainly on the northeastern side of the Bandas del Sur, but also locally exposed on the south, west and north coasts. A volume of 3.5 km$^3$ DRE has been estimated. It consists of a basal unit comprising interstratified ash and surge deposits (Unit A), a coarse plinian fall deposit (Unit B), a complex sequence of ash surge deposits and plinian fall beds (Unit C), an internally complex intraplinian ignimbrite (Unit D) and an upper plinian fall deposit (Unit E). The fall units have a southeasterly dispersal centred on Poris de Abona (2.6 m max. thickness). Unit A (formerly known as the Wavy Deposit; Bryan, 1998; Bryan et al., 1998a), frequently displays...
irregular and often isolated undulations and local thickening that can be attributed to a range of factors at different localities, such as localized surge bedforms, draping of an irregular substrate (rocks, vegetation), impact sags caused by Unit B pumice clasts, dewatering effects, wind effects and post-depositional slumping (on steep slopes). The Caleta eruption is associated with a partial caldera collapse and is characterised by a plinian eruption column interrupted by phreatomagmatic episodes and a phase of column collapse leading to the formation of the Caleta ignimbrite (Unit D).

Poris Member: The Poris Member (Ar/Ar: 268 ka; Edgar et al., 2002; Edgar, 2003; Edgar et al., in review) is widespread across the Bandas del Sur and across the northern, northeastern and western coastal regions, with a total estimated deposit volume of 13-14 km$^3$ (3-4 km$^3$ magma DRE). It consists of a complex sequence of a basal stratified surge deposit, Unit A; a coarse plinian fall deposit, Unit B; two accretionary lapilli ash surge deposits, Unit C; a white phreatomagmatic ignimbrite, the Manteca ignimbrite; a thin plinian fall bed, Unit D; another accretionary lapilli ash surge deposit, Unit E; a complex progressively aggrad ed ignimbrite sequence, subdivided into the Manteca, Abona, Quinta and Mareta ignimbrites; a further interval of plinian fallout, Unit F; an interval of more mafic, dark grey plinian fallout, Unit G; and a complex upper ignimbrite-surge sequence, the Confital ignimbrite. As in the Aldea and Fasnia Members (see below), a complex stratigraphy developed through repeated (partial) column collapse, recurring phreatomagmatic episodes, chemical zonation and magma mingling, and ongoing plinian fallout. Caldera collapse is also associated with this eruption.

Fasnia Member: The Fasnia Member (Ar/Ar: 309 ka; Edgar, 2003; Edgar et al., in review) represents perhaps the most complex eruption sequence preserved on Tenerife. It has an estimated deposit volume of 62 km$^3$ (13 km$^3$ magma DRE) and is widespread across the Bandas del Sur, the Guimar Valley to the northeast, and along the northern and western slopes of the Las Cañadas edifice. 21 units have been recognised, which can be best summarized as a very complex plinian fall sequence (>10 m thick) with seven distinct intercalated ignimbrite units (Maracay, Fuento, Ravelo, Santo, Cueva, Bueno, and Atogo ignimbrites), eight lithic-rich surge deposits and a basal accretionary lapilli ash fall/surge deposit. The Fasnia eruption had a complex history beginning with a phreatomagmatic phase, followed by the development of sustained plinian eruption columns, which collapsed at several times to produce pyroclastic flows and was also interrupted by at least a further six phreatomagmatic phases. A disconformity within the sequence represents a temporary eruption hiatus. Several stages of gradual caldera collapse are thought to have occurred during the eruption.

Aldea Member: The Aldea Member (Ar/Ar: 319 ka; Edgar, 2003; Edgar et al., in review) is widely exposed across the Bandas del Sur and consists of a complex interbedded sequence of four ignimbrites (Tajao, Antagas, Guama and Infantes ignimbrites) and 11 plinian fall/minor surge deposits. A deposit volume of 13.5 km$^3$ (3 km$^3$ magma DRE) has been estimated. The Aldea eruption was characterised by an intense plinian phase, with variations in column height and vent stability and intermittent phreatomagmatic activity, followed by repeated partial column collapse events, which generated pyroclastic flows.

Guajara Formation
The Guajara Formation consists of more than 20 individual plinian fall deposits overlain by seven named major members (Río, Eras, Helecho, Arico, Incendio, Abades and Granadilla; Table 2). We will see four major members, listed below from youngest to oldest.

Granadilla Member: The Granadilla Member (K/Ar: 570 ka; Bryan et al., 1998a; 2000) is the uppermost member of the Guajara Formation. It is widespread across the Bandas del Sur and consists of a lower plinian fall sequence and an upper ignimbrite. The plinian fall sequence (5.2 km$^3$) is comprised of (a) a 1.2 m thick multiple-bedded plinian fall deposit, (b) a 2 cm thick phreatoplinian ash fall deposit, (c) a ≤1.8 m thick plinian fall deposit and (d) an up to 6.3 m thick reverse-graded pumice-bomb rich plinian fall deposit. The fall succession had deposit had a >5 km$^3$ magma DRE volume. The ignimbrite deposit (>5 km$^3$ magma DRE) is up to 30 m thick and consists of multiple massive beds and an upper vent-derived lithic-rich zone. This member represents the final climactic eruption of the Guajara magmatic cycle beginning with an moderately high plinian column (~17 to ≥25 km), which was initially unsteady and interrupted by
phreatoplinian activity, but later become more intense, followed by column collapse and the generation of pyroclastic flows. The eruption culminated in widespread caldera collapse.

**Abades ignimbrite:** The Abades Member (K/Ar: 596 ka, Ancochea et al., 1990) is one of the most widespread members of the Guajara Formation. It is exposed across the Bandas del Sur and has an estimated preserved volume greater than 1.6 km$^3$ magma DRE. The Abades Member is composed of a pumice-rich plinian fall deposit (up to 2 m thick) and a lithified, grey to light brown, diffusely stratified, poorly sorted, pumice depleted phonolitic ignimbrite which has a thickness of >12 m. This unit formed initially by fallout from a plinian eruption column, which later collapsed to produce progressively aggrading pyroclastic flows. The ignimbrite also features interbedded local accretionary lapilli-bearing ash-rich pyroclastic surge units suggesting phreatric explosions at the coast due to the interaction of the pyroclastic flow with the ocean.

**Arico Member:** The Arico Member (Ar/Ar: 668-610 ka, Huertas et al., 2002; Brown et al., 2003) consists of a thin (up to 119 cm), poorly preserved, initial phreatoplinian and plinian fall deposit overlain by a thick non-welded to welded ignimbrite (~18 m). The Arico ignimbrite is one of the most distinctive units in the Bandas del Sur stratigraphy because it features an orange to grey colouration and is the only extensively welded ignimbrite on the southeastern coastal plains of Tenerife. Volume estimates based on the remaining exposures suggest a volume in excess of 2 km$^3$ magma DRE. The ignimbrite is composed of several individual depositional units, which were emplaced by pyroclastic surges and flows channelled down large barrancos on the upper slopes of the volcano. Welding occurred due to a sudden vent widening event and rapid column collapse as evidenced by an influx of lithic clasts before the onset of welding.

**Eras Member:** The Eras Member (not dated) outcrops sporadically across the Bandas del Sur and consists of a 2.3 m thick plinian fall deposit overlain by a cream to white, generally massive ignimbrite (up to 4 m), with banded pumice. The Eras ignimbrite is distinct from other Guajara ignimbrites by its relatively high free crystal content. A volume of >0.4 km$^3$ magma DRE has been estimated. Like many other members of the Guajara Formation, the Eras eruption began with a high sustained plinian eruption column which later collapsed to produce pyroclastic flow(s).

**Tajao Locality**

The area surrounding the coastal town of Tajao provides one of the best records of the large-scale explosive activity of the Las Cañadas Caldera. Not only were many of the major plinian fall deposits dispersed to the southeast, but many pyroclastic density currents were funnelled down major drainage lines (the Barranco del Río, in particular) and reached the coast at or near Tajao. Between the emplacement of the Arico (665 ka) and Abrigo (188 ka) ignimbrites, major drainage lines (barrancos) were repeatedly filled with ignimbrite or mantled by fallout deposits and then rejuvenated by erosion. Lateral migration of barrancos between eruptions led to the development of a highly complex stratigraphy and a continually evolving palaeotopography. Today we will observe just a small part of this complex pyroclastic record, beginning with some of the major deposits of the older Guajara Formation and then moving on to some of the most significant members of the more recent Diego Hernández Formation.

**Stop 1: Playa de Cayado Hondo**

From the Autopista del Sur, take the San Miguel de Tajao exit and drive southeast towards the town (Fig. B1, Locality Map). Turn right on to the sealed road that heads south along the coast towards La Caleta. Stop on the far side of the Barranco de Guasiegre and walk along the gravel track on the coastal side of the road to the large cliff exposures near the town of Tajao. This is an excellent exposure of two of the major members of the Guajara Formation (Fig. B1, Stop 1). As shown in the field sketch, the Abades ignimbrite forms the ~15 m high cliffs at the northern end. It can also be seen extending out along the headland on which the town of Tajao is situated. A steep erosional contact on the southern margin of the Abades ignimbrite is buttressed by poorly sorted and stratified reworked material, apparently derived from the Abades by surface erosion during the ensuing repose period. This unconformity is then mantled by the plinian fall deposit of the Granadilla Member (600 ka), which elsewhere reaches a maximum thickness of more than 9 metres. It is conformably overlain by the Granadilla ignimbrite, the final deposit to be emplaced in the Guajara cycle.
Stop 2: Barranco de Guasiegre

Walk back along the track to the Barranco de Guasiegre and follow the southern wall of the barranco up along its first tributary (Fig. B1, Locality Map). The almost continuous exposure and deep dissection of the stratigraphy in this area provides an excellent opportunity for detailed investigation of facies transitions and ignimbrite emplacement processes (Fig. B1, Stop 2). This section of the barranco wall is formed almost entirely of deposits belonging to the younger Diego Hernández Formation (DHF), although remnants of the Guajara Formation (including the Arico, Abades and Granadilla Members) are preserved locally at the base.

Aldea Member: Only the ignimbrites of the Aldea Member are exposed here in the Tajao region, and in the eastern wall of the Las Cañadas caldera, as the plinian fall sequence has a southerly dispersal (2.6 m, max. thickness) and we are on its eastern margin here. In the upper part of the tributary, the lowest exposed unit is the Tajao ignimbrite, which was the first ignimbrite in the Aldea eruption sequence. It is a highly variable unit but is generally rich in coarse banded pumice and reddish brown hydrothermally altered lithic fragments, and is often fines-depleted. It is overlain here by the 75 cm-thick Unit C deposit, which consists of a basal plinian fall bed (5 cm), a lithic-rich surge deposit (9 cm) with an upper 2 mm-thick ash layer, and a reverse-graded, stratified, lithic-rich plinian fall deposit (up to 60 cm). Banded pumice content decreases up through Unit C but reappears at higher levels in the Aldea Member. This is conformably overlain by the coarse, pumiceous Antagas ignimbrite, which has an unusual coarse pumice facies (lacking an ash matrix) on the upslope side of the ridge of Granadilla ignimbrite (see Fig. B1, Stop 2) which is suggestive of a lateral or terminal levee deposit.

Fasnia Member. The plinian fall sequence of the Fasnia Member has an easterly dispersal and we are on its southern margin here, with much reduced unit thicknesses. The ignimbrite units have highly varied dispersal patterns and internal facies variations but the two most widespread deposits are the Ravelo and Santo ignimbrites, which are well exposed at this locality. In the upper part of the tributary, the Ravelo deposit consists of three distinct facies: a lower grey lithic concentration zone (LCZ), a middle stratified, fine-grained facies (SFG) and an upper pinkish pumice concentration zone (PCZ) (Fig. B1, Stop 2, inset). Together, these form a regionally extensive ignimbrite veneer deposit laid down progressively by a high velocity pyroclastic density current of changing componentry. At the top of the ridge of Granadilla ignimbrite, the Ravelo veneer deposit is reduced to a thickness of only 2 cm, although it is still divided into a lower grey lithic-rich facies and an upper pink pumiceous layer. Note the lack of a palaeosoil on the Granadilla ignimbrite, despite an intervening time interval of about 290 ka. After bypassing much of this area, the Ravelo grades downstream into thick valley-ponded deposits near the coastline, best exposed at the nearby town of La Caleta (>12 m thick). In this section, the downstream facies is a fines-depleted coarse lithic breccia, which may have been deposited en masse by frictional freezing. The Ravelo deposit is mantled by Unit H, consisting of a lower plinian fall bed, an erosive, stratified lithic-rich surge deposit and an upper reverse-graded plinian fall horizon. Unit H is the thickest fall deposit in the Fasnia Member, reaching a maximum thickness of 3.8 m near the town of Fasnia. It is overlain by the Santo ignimbrite which forms the thickest unit in this section (up to >30 m near the coast). The Santo ignimbrite is generally cream or pink in colour and pumiceous, but it has three widespread lithic concentration zones (LCZ1-3). LCZ1 was responsible for erosion of a basal stratified, fine-grained (SFG) facies in the head of the tributary (Fig. B1, Stop 2, inset) and LCZ3 is very well-developed, with basal load and flame structures.

Walk back down to the bend in the tributary formed by a ridge of the Ravelo ignimbrite (coarse lithic breccia facies) and walk up along the top of this ridge to the top of the Santo ignimbrite. It is directly overlain here by the uppermost unit in the Fasnia Member, the Atogo ignimbrite. This unit actually consists of a complex succession of more than 15 beds, comprising thin ignimbrite deposits, stratified surge deposits, ash layers and thin intervals of plinian fallout. Here, it consists of two thin ignimbrite sheets with lower LCZs and upper PCZs, a third ignimbrite sheet with reverse grading of lithic clasts, a very complex surge deposit (75 cm max), another thin (30 cm) fines-depleted ignimbrite with reverse-graded lithic clasts, and a coarse lithic ground layer (180 cm thick) containing a range of volcanic, hydrothermally altered and syenitic lithic fragments up to 25 cm in diameter. Note the outsized lithic fragments scattered through the lower ignimbrite beds. The Atogo unit is interpreted to represent the climactic final stages of the Fasnia eruption,
in which caldera collapse (or large-scale vent collapse) loaded the collapsing eruption column with lithic debris.

Figure B1 (continued next page): Locality map of the Tajao Locality and field sketches at Stops 1 and 3. A stratigraphic log of the Fasnia Member is also shown at Stop 3.
**Upper Members.** Four upper members of the DHF form sheet-like deposits on top of the Fasnia Member in this area: the Poris, Caleta, Batista and Abrigo Members. These can be observed with greater ease at later stops.

**Stop 3: Tajao road section**

Return to the coach and drive back to the main access road to Tajao (Fig B1, Locality Map). Turn right and follow the road down to the town. We will walk back up and examine the cuttings on both sides of the road.

This section provides good exposures of the Fasnia, Poris and Caleta Members of the Diego Hernández Formation (Fig B2, Stop 3).

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**Tajao coastal plain. Stratigraphic relationships and facies architecture (cont.)**

**Stop 3; field sketch and stratigraphic log of the Fasnia Member**

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**Fasnia Member.** At the bottom of the road on the left hand side is an excellent exposure of the Fasnia Member, which is very different from the one seen at the previous stop. The basal unit is a valley-ponded facies of the Ravelo ignimbrite with irregular lithic concentration zones, ash-rich zones and a coarse upper pumice concentration zone. An irregular gravelly sedimentary unit is exposed at the top of the Ravelo but appears to have been emplaced after the Fasnia eruption following erosion of the pyroclastic succession. The Ravelo ignimbrite is overlain by Unit H (as seen at Stop 2), followed by a discontinuous white accretionary lapilli ash bed, Unit I. This
distinctive unit is very widespread across the southeastern slopes of Tenerife and often mantles the Santo ignimbrite or appears more widely as its lateral equivalent. It is interpreted as the co-ignimbrite ash deposit associated with the Santo ignimbrite, which is curiously absent at this locality, given that it forms a 30 m-thick deposit only 400 m to the southeast at Stop 2. These two localities are separated by a palaeo-high (the thick cliffs of Abades and Granadilla ignimbrites seen at Stop 1) which must have separated two barrancos that were utilized by different pyroclastic density currents even during a single eruption. Unit I is overlain by the Cueva ignimbrite (the fifth in the Fasnia Member) which is laterally variable but has a lower LCZ and an upper PCZ here. It is overlain by its more widespread co-ignimbrite ash cloud surge deposit, Unit K, which consists of a lower crystal-rich bed and an upper white ash bed. This is followed by the Unit L plinian deposit, the complex Bueno ignimbrite, the Unit N plinian deposit and the Atogo ignimbrite. This alternation of plinian fallout and intraplinian ignimbrites is typical of the Fasnia Member and reflects the unstable eruption column dynamics caused by overwidening of the vent, periodic vent wall collapse and repeated episodes of phreatomagmatism (due to both a caldera lake and groundwater). Remember that the dispersal axis of the plinian fall sequence is much further north so that we only see the thin distal margins of the plinian units in this area (they are each 2-4 m thick around the town of Fasnia). Also note the different facies preserved in the Atogo unit here compared to Stop 2 and the well-developed asymmetric flame structures. This entire sequence was previously interpreted as marine sediments (Bryan et al., 1998a) but can now confidently be correlated with other exposures of the Fasnia Member.

Poris Member. The palaeosoil developed on top of the Atogo ignimbrite is overlain by the Poris Member. At this locality, the Poris Member is represented by Units A and C (separated by a single layer of pumice possibly representing the distal margin of the Unit B plinian fall deposit) and the Abona ignimbrite. The Abona is a white, pumiceous deposit with abundant etched-out tube pumice. Further up the road on the left hand side, the Abona grades up into the Quinta ignimbrite, which is a widespread coarse lithic-rich facies thought to represent an interval of caldera collapse. This grades up into a poorly preserved remnant of the Mareta ignimbrite, which is characterized by its high banded pumice content and relatively low lithic content. The Abona-Quinta-Mareta sequence is highly distinctive and well exposed at many other localities in southern Tenerife.

Caleta Member. As shown in the field sketch for this stop, there is a vertical contact between the Abona ignimbrite and the Caleta Member near the bottom of the road. This seems to represent the wall of a barranco formed between the Poris and Caleta eruptions. At this locality, Unit A consists of a single laminated white ash bed which displays some impact sags but is generally uniform in thickness. The Unit C ash bed thickens substantially adjacent to the palaeotopography and is intercalated with lower depositional units of the ignimbrite. The Caleta ignimbrite is thickest around Tajao (>12 m) and consists of several subunits of varying pumice and lithic contents. These are best seen on the right hand side about halfway up the road section. Note the prominent lithic concentration zone near the top of the unit, which may record a period of small-scale caldera collapse. A stratified facies is well developed on the lee side of a palaeotopographic high in this section. Such topographic effects on ignimbrite emplacement are very widespread in this and other ignimbrites of the DHF.

Stop 4: Lunch
From Stop 3, it is a short walk into Tajao to taste the latest catch in the fish restaurant.

PORIS DE ABONA LOCALITY, OLD ARICO ROAD
The area to the west of the Autopista near Poris de Abona displays excellent exposure of the mid to upper Guajara Fm and much of the Diego Hernández Fm, preserved in road cuttings and valley walls although individual units are discontinuous due to syn- to post-eruptive erosion. Take the Autopista exit to Poris de Abona.

Drive through Poris de Abona township taking the first main road to the right (ie the old road to Lomo de Arico) (Fig. B3, Locality Map). After passing beneath the Autopista through a small tunnel, continue for about 200 m to a suitable parking spot. At this locality we will walk along the old Arico road, first heading west along the northern wall of the Barranco de los Caballos and southern slopes of an old scoria cone, Montaña de Puerto, which is the highest point in the area.
After crossing the barranco we will continue south along the road to the uppermost hairpin bend. We will then return downhill along an ENE ridge and across the flats in the main barranco.

Poris de Abona, old Arico road. Stratigraphic relationships and facies

Figure B3 (continued next page): Map of the Poris de Abona Locality, field sketches at Stops 1 to 3, including stratigraphic logs at Stop 1, and detailed field sketch and stratigraphic correlation showing the lateral facies variation of the Abrigo Member at Stops 2 to 3.
Porís de Abona, old Arico road. Stratigraphic relationships and facies (cont.)

Stop 3 cont.; field sketches and logs showing facies architecture of the Abrigo Member

**ABBREVIATIONS**
- afl: massive flow-rich ignimbrite facies
- af: all-stratified flow-rich ignimbrite facies
- pc: pumice-concentrated facies
- qps: quarry pumiceible or soft facies
- ddd: dikes and dikes of dikes
- ADP: possible Abrigo fall deposit

**Figure B3 (continued)**
Stop 1: Cuttings near the Bus Parking Area

Stop 1 consists of multiple cuttings alongside the road displaying several major members of the Diego Hernández Fm (Fasnia, Poris, Caleta and Abrigo members) which have infilled a palaeovalley above the Arico ignimbrite (Fig. B3, Stop 1). The main widespread ignimbrite unit of the Arico Member is common throughout this region, although, unlike the deposits at Tajao, it is only poorly welded (orange-grey).

The Fasnia Member around Poris de Abona consists of a thin (10 cm) interval of Unit D plinian fallout, ~40 cm of Unit F plinian fallout (with a prominent layer rich in banded pumice), a stratified veneer facies of the Ravelo ignimbrite with several associated surge deposits and the Unit H plinian fall deposit (which is >2.5 m thick in this area). At this stop, on the south side of the road, the lithic-rich Ravelo ignimbrite and Unit H occur above the Arico ignimbrite, and this in turn is overlain by possibly reworked pumice.

The Poris Member overlies the Fasnia Member on the south side of the road and near the base of the cutting on the north side. Units A (interbedded ash and pumice lapilli), B (pumice lapilli plinian fall), C (fine ash surge deposit), Abona ignimbrite and Quinta ignimbrite of the Poris Member are preserved in this region. At this stop, all units, except the Quinta ignimbrite are exposed.

The Caleta Member overlies the Poris Member on both sides of the road. Both the basal wavy interbedded pumice lapilli and ash horizon (Unit A) and pumice lapilli plinian fall deposit (Unit B) occur at this stop, as well as stops 2 and 3. We will not see the Caleta ignimbrite at these stops, although it forms an extensive ignimbrite veneer deposit elsewhere in the region.

At least 2 or 3 older pumice lapilli plinian fall deposits occur between the Abrigo and Caleta Member in the area, including the lowermost, Batista Member, which can be viewed at this stop. The thin matrix-supported, lithic-rich unit above the Batista Member here is the Abrigo ignimbrite, which is best described at locality 3.

Stop 2: Uphill climb on the southern side of the Barranco de Caballos

Cross the Barranco de los Caballos and follow the road to the second hairpin bend below the top of the ridge (Fig. B3, Locality Map). Complex palaeotopographic relationships between various members of the Diego Hernández and Guajara formations are visible on a north-facing gully wall alongside the road (Fig. B3, Stop 2). Near the base, the Arico ignimbrite has been incised and infilled with a sedimentary deposit, which is subsequently overlain by a thick ignimbrite unit (possibly the Abades or Granadilla ignimbrites). The post-Guajara palaeotopography has been infilled by the Fasnia, Poris and Caleta Members, and finally mantled by the Batista Member and the sheet-like lithic-rich Abrigo ignimbrite. Note the lithic block concentration zone within the Abrigo ignimbrite here. This will be discussed further at Stop 3.

Stop 3: Uppermost hairpin bend along the road

Continue uphill around the spur from Stop 2 and stop in the gully on the other side, just below the uppermost hairpin bend at the top of the ridge (Fig. B3, Locality Map and Stop 3). The Abades Member is exposed at the base of the gully and consists of a relatively lithic-rich and mafic pumice-rich ignimbrite overlying a 35 cm pumice lapilli plinian fall deposit with a fine-grained basal zone. The Abades ignimbrite is widespread in this region.

The Abades Member is overlain successively by the Fasnia, Poris, Caleta, Batista and Abrigo members. The sheet-like Abades ignimbrite caps the stratigraphic sequence at stops 2 and 3 and both the Sur-A and –C depositional units are present. A 30 cm thick pumice lapilli deposit beneath the Abrigo ignimbrite at the end of the spur between Stops 2 and 3 could be an Abrigo plinian fall deposit, or alternatively could be an older fall deposit of the Cruz Sequence.

Walk to the uppermost hairpin bend and view the Abrigo ignimbrite across stops 2 and 3. Valley-ponded (thick, massive, lithic-rich ignimbrite) to ignimbrite veneer (stratified, pumice or finer-grained lithic-rich ignimbrite) transitions are seen to occur over a very low relief palaeotopographic high. A lithic block concentration zone in the upper part of the Sur-A unit also thins and becomes finer-grained towards this high. This zone contains a large variety of accessory ± accidental lithic blocks (fresh and altered varieties of syenite, basaltic/basalitic to phonolitic coherent to fragmental/welded volcanic rocks, pyroclastic/epiclastic breccias) and is interpreted to represent a syn-eruptive partial caldera collapse event. Also featured here are high
aspect ratio massive pumice-rich ignimbrite lobes near the base of the Sur-C unit, which are common at this stratigraphic level across the Bandas del Sur.

**Stop 4: Descent into the Barranco de los Caballos**

From Stop 3 walk northeast down the spur into the main valley (Barranco de los Caballos) (Fig. B3, Locality Map), where both the Abades and Arico ignimbrites are well exposed. The grey to orange colouration of the Arico ignimbrite is clearly evident here, although it is still poorly welded. The base of the ignimbrite is distinctly lithic-rich, with abundant vent-derived lithic clasts, and displays low angle lithic-rich bedforms. At one small outcrop older units of the Guajara Fm are exposed beneath the Arico ignimbrite: (a) several older plinian fall units, overlain by (b) the white Eras ignimbrite, then by (c) two additional pumice lapilli plinian fall deposits.


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