Spatial and temporal constraints on the pattern of crustal rotation
in the Central Andean forearc of Northern Chile (27-30°S)

By

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SPATIAL AND TEMPORAL CONSTRAINTS ON THE PATTERN OF CRUSTAL ROTATION IN THE CENTRAL ANDEAN FOREARC OF NORTHERN CHILE (27-30°S)

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ABSTRACT

A total of 125 sites have been collected for palaeomagnetic analysis from two localities within the Coastal Cordillera and Precordillera regions of Northern Chile, between 27-30°S, in addition to ~200 samples collected from a magnetostratigraphic profile through the Pabellón Formation.

In the southernmost (Tres Cruces) sampling locality, spanning the Coastal Cordillera-Precordillera boundary, the early, mid and latest Cretaceous-earliest Paleocene magmatic arcs intrude the earliest Cretaceous country rocks to the west and late Cretaceous rocks to the east. An ~50km transect sampled along Quebrada de Los Choros (~29°45'S), indicates that a consistent clockwise rotation of ~10° is recorded by primary magnetisations isolated from all of the units sampled, regardless of age, lithology or location. This is consistent with existing palaeomagnetic data immediately to the south, which has been recalculated, and suggests that localised rotations did not play a significant role in accommodating deformation at this latitude.

Palaeomagnetic sampling of two plutons from the latest Cretaceous-earliest Paleocene magmatic arc situated in Quebrada de Los Choros and ~50km to the north, indicates that a sharp discontinuity exists in the regional rotation pattern, with the northern most pluton recording 30° of clockwise rotation. This discontinuity is observed to coincide with an area of diffuse deformation along which sinistral displacement is accommodated along predominantly NW orientated faults. This zone of deformation is interpreted as reflecting a pre-existing fundamental fault zone, similar in nature to a number of such NW striking crustal anisotropies that are observed to pre-determine much of the modern architecture of the modern forearc. These reactivated fault zones are interpreted to form the boundaries of large domains that display homogenous patterns of large magnitude clockwise rotation and are defined through a large number of palaeomagnetic studies.

The age of rotation is not well defined in the Tres Cruces area, with a maximum age of 70-60Ma suggested for the youngest rotated plutons sampled, and the observation of a remagnetisation to the south. A second sampling area, situated in the Chilean Precordillera (c.27°45'S), was chosen to try and investigate the temporal accumulation of rotation in the northern Chilean forearc. A wide range of rocks of Triassic to Eocene age were sampled, albeit within the La Temerá Fault System, associated with the Incaic orogeny, which marked the initial stages of the most recent phase of mountain building in the Andes. The oldest rocks record primary magnetisations that indicate ~40° of crustal rotation, very similar in magnitude to that recorded in the Coastal-Cordillera-Precordillera boundary zone to the west, suggesting that the La Temerá Fault System does not control the regional rotation pattern. In addition, the youngest material sampled, a 40Ma pluton, suggests that rotation was completed prior to the Incaic orogeny, suggesting that crustal rotation in the present day forearc is not a consequence of plateau uplift or crustal thickening.
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At no time during the registration for the degree of Doctor of Philosophy has the author been registered for any other University award without prior agreement of the Graduate Committee.

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Signed

Date
Chapter One

Introduction

1.1 Background

The central Andean margin of South America (Figure 1.1-1) is characterised by its response to the subduction of the oceanic Nazca (and previously Farallón) plate [e.g. Isacks, 1988; Allmendinger et al., 1997]. Studies from convergent margins around the world indicate that a significant component of the observed deformation of continental crust is often accommodated through the rotation of crustal blocks [e.g. western North America-Magill et al., 1982; Sonder et al., 1994; Jackson & Molnar, 1990; New Zealand-Vickery & Lamb, 1995; Roberts, 1995; Southern Spain-Platzman & Lowrie, 1992; Allerton et al., 1993]. The central Andes is no exception to this and palaeomagnetically determined crustal rotations have been widely reported throughout, Peru [e.g. Rousse et al., 2002, 2003; Gilder et al., 2003], Bolivia [e.g. MacFadden et al, 1990, 1995; Butler et al., 1995], northern Chile [e.g. Hartley et al., 1988, 1992; Somoza & Singer, 1999; Amagada et al., 2003a, 2006] and north-western Argentina [e.g. Coutand et al., 1999].

One of the major features of the Central Andes is referred to as the 'Bolivian Orocline', a ~55° change in overall strike of topography and trench centred at ~19°S (Arica Deflection-Figure 1.1-1). The overall pattern of crustal rotations in the Central Andes appears to be intimately associated with the morphology of the orocline, with anticlockwise (ACW) and clockwise (CW) rotations observed to the north and south of Arica respectively, reflecting the relative displacement of the northern and southern limbs [e.g. Oviedo et al, 1991; Beck, 1999, 2004; Lamb, 2001a, 2001b; Amiagada et al., 2003b, 2003c].
Outline map of South America showing the extent of the Andean Cordillera. Dark (light) grey shaded area corresponds to elevations >1000km (>3000km). The extent of the 'Bolivian Orocline' and Arica Deflection are indicated in red, with the sense of crustal rotation indicated for both the Northern and Southern limbs. Currently active volcanic centres are indicated, as are the northern, central and southern Andean sectors following the division of Gansser (1977).
A persistent assumption is that the observed pattern of crustal rotations throughout the Central Andes, is causally related to the most pronounced period of uplift to have affected the Andean margin, which occurred during Neogene times. The overall pattern of anticlockwise (clockwise) rotations to the north (south) of the Arica Deflection has consequently been interpreted by many authors to result from a single period of rigid (oroclinal) bending of the western margin of South America, as proposed by Carey (1958) [e.g. Kono et al., 1985].

As the quantity of palaeomagnetic data throughout the Central Andes has increased in recent years, so the level of detail concerning the spatial distribution and temporal accumulation of crustal rotations has increased accordingly. The increased spatial resolution exposes regional-scale intricacies in the overall rotation pattern that cannot be rationalised using a single, orogen-wide rotation mechanism, whilst the increasingly more reliable age constraints placed on more recent studies contradict the proposed Neogene age for the rotations [e.g. Arriagada et al., 2006, Roperch et al., 2006].

Much of the discussion within this thesis concerns the distribution of crustal rotations throughout the Central Andes, with data drawn from a substantial number of palaeomagnetic studies during the past three decades. For the ease of discussion in the remainder of this thesis, the overall pattern of crustal rotations in the Central Andes (which encompasses all of the palaeomagnetic data published from the region in the form of peer-reviewed journal articles or conference contributions), is referred to as the Central Andean Rotation Pattern (CARP), and this simply refers to the distribution of the palaeomagnetically determined crustal rotations within the Central Andes.

The overall database is summarised in Appendix B, with palaeomagnetic data from individual studies quoted directly from the original source, or combined with
that from other studies where appropriate. Data from this study is also included for completeness, but where the pre-existing dataset alone is discussed in isolation, this will be indicated in the text. Crustal rotations are calculated in an identical manner for all of the palaeomagnetic data presented and this will be discussed further with respect to the interpretation of palaeomagnetic data in Section 2.7.

In this thesis, new palaeomagnetic data is interpreted from two areas in northern Chile, specifically chosen to investigate both the spatial extent and timing of crustal rotations in the southern limb of the Bolivian Orocline. Previous studies and experience of sampling within the Andean margin suggest that the majority of lithologies exposed in the field areas would be suitable for palaeomagnetic sampling and are likely to record crustal rotations.

1.2 Plate Tectonic Setting of the Present-day Andes

The western continental margin of South America is dominated by the 8,000 km long Andean Cordillera with maximum elevations of 7,000 m a.s.l. (Figure 1.1-1). The leading (western) edge of South America is, today, in contact with three oceanic plates, the Cocos, Antarctic and Nazca plates (Figure 1.2-1). The current process of subduction (related to the Andean orogenic cycle) is believed to have been ongoing for ~200Ma, with its initiation related to the rifting of Gondwana [e.g. Coira et al., 1982]. Consequently subduction related features dominate the geological development of the Andean margin since this time.

The Nazca plate is being consumed along the Peru-Chile Trench at a rate of 70-80 mm.a⁻¹, which represents one of the highest convergence rates observed in the world [e.g. DeMets et al., 1990; Bird, 2003-Figure 1.2.1]. The extreme relief of the present-day Andes developed primarily during the Neogene (25Ma-Present), in response to accelerated convergence between the oceanic Nazca plate and
Figure 1.2-1  Bathymetric map of the Nazca plate based on ETopo2 data. The main spreading ridges are indicated, as are the Easter and Juan Fernandez microplates, major fracture zones and aseismic ridges [after Bird, 2003]. White arrows represent the relative motion of the Nazca plate with respect to stable South America [from Bird, 2003].

Figure 1.2-2  Schematic cross section showing the tectonic activity of the central Andes (15°-27°S). VE is vertical exaggeration. Diagram illustrates 'normal' Andean-type subduction [Figure 1 from Wdowinski & O'Connell, 1991].
continental South America [e.g. Pardo-Casas & Molnar, 1987; Isacks, 1988; Allmendinger & Gubbels, 1996].

The present-day tectonic setting of the Andes has long been considered to represent the definitive example of an active orogenic margin resulting from the continued subduction of normal oceanic lithosphere beneath continental crust [e.g. James, 1971]. An "Andean-type" convergent margin is therefore modelled as an active volcanic arc, flanked by a deep trench to one side and an active foreland fold and thrust belt to the other (Figure 1.2-2) [Wdowinski & O'Connell, 1991]. While this representation may satisfactorily describe certain segments of the Andean subduction margin, it is a gross oversimplification of the current tectonic setting of the present day Andean Cordillera, which is characterised by numerous morphological, tectonic and magmatic along strike variations [e.g. Whitman et al., 1996], including the subduction of aseismic ridges (or otherwise thickened oceanic crust) and the observation of a number of areas of 'flat' subduction, where the dip of the subducting slab shallows significantly (Figure 1.2-3).

1.3 The Andean Orogeny

In contrast to earlier orogenic periods (a discussion of which is beyond the scope of this thesis), the Andean cycle is/was controlled by the subduction of generally normal oceanic (Pacific) crust. Starting from the final break-up of Gondwana, the Andean cycle or Andean Orogeny is linked to several compressive phases since the early Cretaceous, but is most commonly used to refer to Tertiary to present day deformation [Coira et al., 1982].

Although Andean-type subduction was initiated in the late Triassic the main period of 'Andean' deformation began with separation of the South American and African-Indian plates, as rifting and seafloor spreading formed the South Atlantic Ocean
Figure 1.2-3  Flat subduction in the Central Andes. A-Contour map illustrating the depth to the upper surface of the subducted Nazca plate throughout the southern Central Andes (in km), indicating that the dip of subducted plate shallows substantially between 28-33°S [redrawn from Cahill & Isacks, 1992]. Elevations >1000m (>3000m) indicated in light (dark) grey, with active/recent volcanoes identified as triangles. Note the coincidence between flat subduction and volcanic gap. B & C-idealised cross-sections illustrating steep slab and flat slab subduction [Figure 8 from Gutscher, 2002].
basin during the earliest Cretaceous [c.130Ma-Renne et al., 1992; Tumer et al., 1994; c.150Ma-Eagles, 2006]. As the rate of convergence between South America and the Farallón plates increased, so the generation of plutonic and volcanic rocks (typical of the modern margin) became predominant. Another typically 'Andean' feature is the development of ensialic extensional basins, as well as a passive Atlantic (eastern) margin [Pankhurst & Rapela, 1998].

The first major contractional episode in the Andean cycle occurred in response to the increased westward movement of South America, possibly due to the early Aptian separation of North Africa and South America. Generally referred to as the Peruvian Orogeny [e.g. Megard, 1984], this event affected both northern Chile and coastal areas of southern Peru in an area that represents the modern-day forearc region. The actual amount of shortening accommodated during this event is difficult to decipher from the much more extensive deformation (plateau uplift) during the Tertiary and this event is generally recognised from unconformities within the rock record.

A marked increase in the rate of convergence between the South American and Farallón plates initiated a second period of pronounced uplift and an intensification of convergent arc tectonics in the present-day Precordillera region between 43-33 Ma [e.g. Pardo-Casas & Molnar, 1987; Somoza, 1998]. Referred to as the Incaic orogeny, deformation during this period is characterised by both easterly and westerly directed thrusting at the flanks of the northern and central Andes. Whilst this phase of deformation is viewed as the beginning of truly 'Andean' orogenesis, many authors now agree that the majority of Andean plateau uplift occurred in the Neogene (i.e. post-25Ma) [e.g. Isacks, 1988, Whitman et al., 1996], during what is generally referred to as the Quechua Orogeny. The accelerated convergence along the length of the Andes c.25Ma is related to the separation of the Farallon
oceanic plate into the Cocos and Nazca plates to the north and south of the Galapagos Rift respectively (Figure 1.2-1), producing a pulse of magmatism and tectonic thickening.

1.4 The Central Andes

The Andean margin has been divided into various segments by a number of authors, but generally a three-fold division is employed [e.g. Gansser, 1973; Ramos, 1999]. This separates the margin into Northern, Central and Southern geological provinces, based on their differing tectonic histories (Figure 1.1-1). The geological history of the Northern Andes is dominated by the accretion of Mesozoic and younger oceanic terranes, while the Southern Andes record a predominantly sedimentary history with little or no volcanic-arc activity [Ramos, 1999]. This study primarily concerns the Central Andes and spans southern Peru, Bolivia, northern Chile and northwestern Argentina, with the main study area situated in the southern Central Andes (Figure 1.4-1).

1.4.1 The Central Andean Plateau

The Central Andes represent the world's largest example of a non-collisional,\(^1\) continental plateau, with an average elevation approaching 4km across an area of approximately 300 by 2000 km [Isacks, 1988] (Figure 1.4-1). The plateau is underlain by a substantially thickened crust (Figure 1.4-1), which is estimated to reach a maximum thickness of 80km near to where Argentina, Bolivia and Chile come together [e.g. Zandt et al., 1994; Allmendinger et al., 1997].

The area of maximum crustal thickness coincides with the current location of active arc volcanism as shown in Figure 1.1-1. This coincidence led many early

\(^1\) In contrast to the continent-continent 'Tibetan-type' collisional orogen
Figure 1.4-1  Topography of the Central Andes (as indicated in Figure 1.2.1), based on ETopo2 data, with the study area indicated. Contours Redrawn from Figure 4 of Allmendinger et al., (1997), based on published interpretations of refraction data (Wigger et al 1994) and broadband data [Beck et al., 1996; Zandt et al., 1994]. Dashed contours in northernmost Bolivia and Peru are from James (1971), and the dashed 60-km contour in the southern Puna is from Götze et al., (1994). *Area shown in Figure 1.4-2.
studies to suggest that the extreme elevation of the central plateau resulted from the process of magmatic addition in vast volumes [e.g. James, 1971]. Topographic and structural studies of the eastern plateau however, indicate that shortening could account for a large part of the observed crustal thickness, thereby negating the need for the intrusion of massive volumes of melt into the crust [e.g. Isacks, 1988; Allmendinger et al., 1997].

**Structural evolution of the central Andean plateau**

The development of the central Andean Altiplano/Puna Plateau predominantly occurred during the late Cenozoic [e.g. Isacks, 1988] and then most likely during the past 10 Ma, generally since the late Oligocene to present [e.g. Jordon et al., 1983; Jordon & Alonso, 1987; Sempere et al., 1990]. The plateau is observed to be at its widest (and therefore shortening is assumed to be greatest) through central Bolivia, with the total crustal shortening between the trench and limit of the fold and thrust belt of the Eastern Cordillera estimated to be ≥320 km (Schmitz, 1994). The elevated area decreases in width along strike of the margin both to the north and south, with the overall continental plateau of the central Andes limited to between 10-30°S (Figure 1.4-1).

Isacks (1988), related the topographic expression of the elevated central Andean plateau to the amount of crustal shortening accommodated by the margin, and in particular proposing that along strike differences in the amount of shortening could explain the pronounced decrease in width of the elevated plateau region towards the northern and southern limits of the central Andes. Isacks (1988) inferred that a consequence of differential shortening along the Andean margin would be the enhancement of the seaward concavity of the leading edge of western South America.
1.4.2 Geomorphology of the northern Chilean Andes

The southern Central Andes of northern Chile can be divided into a number of predominantly longitudinal morphological zones that may be traced along a considerable length of the margin (Figure 1.4-2). Various divisions are used but generally most classifications identify almost identical features. The names that refer to these specific areas or zones of the Central Andes (with abbreviations in brackets as used in this thesis) will be introduced with a brief description of their general characteristics.

The most westerly morphological zone, the Coastal Cordillera (CC), represents a low-lying coastal mountain range, which is bordered to the east by a notable depression, widely referred to as the Central Depression and which opens up into the Central Valley (CV-otherwise referred to as the Longitudinal Valley), to the north of 26°S but absent south of there. The CV passes eastwards into the Chilean Precordillera/High Cordillera (PC), which represents a belt of reasonably highly incised topography that demarks the western foothills of the Andes, where elevations approach 4000m. As shown in Figure 1.4-2, both the Tres Cruces and La Guardia field areas are situated in the CC-PC morphological zones (the CV being absent).

Between 21-26°S a series of distinct depressions are observed to separate the PC from the high Andean Cordillera and are related to the inversion of the Mesozoic back-arc basin during the late Cretaceous [Comejo et al., 1993; Mpodozis & Clavero, 2002]. These basins are referred to collectively as the Pre-Andean Depression (PAD). The high Andean cordillera is separated into several distinct morphological zones, with the internally drained plateau area flanked by two bands of extreme topography, referred to as the Western and Eastern Cordilleras (WC & EC).
Figure 1.4.2  Morphological zones of the northern Chilean Andes [modified from Scheuber & Reutter, 1992].
The plateau area itself is the locus of the recently active volcanic arc and is divided into two distinct areas, which reflect topographic differences arising from the diachronous structural history of the plateau. To the north lies the Altiplano, an area characterised by extremely low relief, situated in southern Peru and Bolivia (15-23°S-Figures 1.4-1 & 1.4-3a). To the south, situated in northwestern Argentina, the Puna plateau segment is characterised by substantially more topography than observed in the Altiplano (Figure 1.4-3a).

A comparison of the elevation distribution within the internally drained areas of the Altiplano and Puna segments indicates that elevation within the Altiplano is asymmetrically distributed across a narrow range, which attests to its relative stability and the effectiveness of cut-and-fill processes [e.g. Whitman et al., 1996; Allmendinger et al., 1997-Figure 1.4-3b]. The more symmetrical and broad distribution of the Puna, reflects a tectonically active topography due to its more recent compressional history and, as a consequence, average elevations are ~1km greater than observed in the Altiplano (Figure 1.4-3b). Many authors propose a two or three stage evolution of the high Andes and explain the physiographical differences noted between the Altiplano and Puna regions in terms of diachronous deformation in the two areas. The overall plateau is thought to have developed in a reasonably uniform manner between 30-10Ma, after which time however, deformation north of 23-24°S shifted eastwards into the sub-Andean belt, whilst deformation to the south continued in the Puna until 4-2Ma [e.g. Whitman et al., 1996]

Neotectonic deformation in the present-day Andes focuses on the EC where an active fold and thrust front is propagating eastwards into the foreland area termed the sub-Andean zone (SA). The EC passes southwards into the Sierras Pampeanas (SP) and Pampean Depressions (PD), which, supposedly, represent
Figure 1.4-3  A-Along-strike variation, in the Central Andes, of lithospheric thickness and corresponding changes in topography, highlighting the differences between the Altiplano and Puna, with hatching indicating an area of high seismic attenuation [modified from Whitman et al., 1992, 1996]. In the cross section at the top, the white area above the "Subducted Nazca Plate" is the asthenosphere beneath South America; the white area beneath it is the asthenosphere and deeper mantle beneath the Nazca Plate. In Lower diagram dark grey shading indicates areas with average elevations over 3km, light grey (stippled) shading indicating areas of thin-skinned (thick-skinned) deformation in the Sub-Andean (Pampean) ranges located to the east of the high plateau area [Figure 3-Allmendinger et al., 1997]. B-Comparison of the elevation distribution of the Altiplano and Puna segments of the central Andean plateau. Elevation distribution was calculated from topography contained within internally drained regions of the plateau only [Figure 6, Whitman et al., 1996].
the transition from thin- to thick-skinned (basin and range) type deformation. Further east again, the flat Chaco Plains represent the undeformed foreland.

1.4.3 Geology of the Central Andes and northern Chile

The geology of northern Chile (and indeed the Andean Cordillera as a whole) reflects the long history of convergence and continuous active subduction at the western margin of continental South America [e.g. Dalziel & Forsythe, 1985]. The following section provides a brief overview of the surface geology as observed in northern Chile.

The most recent phase of subduction related to the Andean orogenic cycle during the past 200Ma, began with the fragmentation of Gondwana, which was associated by the emplacement of a broad suite of Permo-Triassic granitic plutons [Brown, 1991], into a Palaeozoic basement comprised of metamorphosed marine sediments [Bell, 1987]. These plutons are observed to crop out in the Coastal Cordillera, but are also exposed in the Precordillera as uplifted blocks (as observed c.28oS-Figure 1.4-4).

Subsequent to the initiation of subduction, the development of the Andean margin is characterised by the intrusion of a number of discrete magmatic arcs parallel to the coastline, starting in the Jurassic. These magmatic arcs (as well as the country rocks into which they are extruded) are observed to young progressively eastwards, with successive arcs only ever observed to overlap with the immediately preceding incarnation (Figure 1.4-4). Each successive arc is therefore observed to occupy a unique situation within the margin, which remains consistent along strike (Figure 1.4-4).

Scheuber & Reutter (1992) identified four such generations of the active magmatic
arc, detailing separate Jurassic-early Cretaceous (also referred to as the La Negra Arc), mid Cretaceous, latest Cretaceous-earliest Paleocene/Eocene and Miocene-recent incarnations (situated in the present day high Andean Cordillera-Figure 1.4-5). Plutons belonging to the mid Cretaceous arc are often difficult to separate from early Cretaceous intrusions, being of similar composition and often in reasonably close proximity. In such cases, only geochronological dating of the plutons can separate the different generations. Many of these magmatic arcs were sampled during this study and a fuller description of these arcs will be given in Chapters Four and Five.

The emplacement of many of the magmatic arcs appears to be intimately associated with and often controlled by, one of a number of (trench-linked) margin parallel fault systems such as the Atacama, Coastal Cordillera-Precordillera and West Fissure-Domeyko Fault systems that were active during the Jurassic-early Cretaceous, late Cretaceous and Eocene-Oligocene respectively [e.g. Diaz, 2000; Grocott & Taylor, 2002; Truelove et al., 2003] (Figure 1.4-4). Displacement along these faults was likely to have been initiated by the thermal weakening of the crust associated with the eastward migration of the active magmatic/volcanic arc, corresponding to periods of deformation where a significant component of the relative motion between the Nazca (Farrallon) and South American plates was accommodated along a small number of discrete structures.

The history of the older, Atacama Fault System (AFS), is largely represented by sinistral displacement within a transtensional-transpressional deformation regime, during Jurassic and early Cretaceous times.

The effects of the Atacama and Coastal Cordillera-Precordillera fault systems concerning the crustal rotation pattern in Northern Chile has been previously
Figure 1.4-5  Map showing the four, easterly youning, magmatic arcs identified in the Antofagasta region of northern Chile [modified from Scheuber & Reutter, 1992].
investigated, whilst the precise effects of the Domeyko Fault System remain to be fully established and will be discussed with regard to palaeomagnetic data from this study in Chapter Seven.

The Mesozoic country rocks of Northern Chile are dominated by the marine and continental sediments and volcanics that accumulated in a series of elongated (margin parallel) ensialic basins. These sediments/volcanics also tend to young progressively towards the east (Figure 1.4-4), but marked division is noted between the fill of intra-arc basins to the west in the Coastal Cordillera (filled by strata belonging to the Punte del Cobre Formation and Bandurrias and Chañarcillo Groups) and retro-arc basins in the Precordillera to the east (filled by the La Ternera, Lautaro, Lagunillas and Quebrada Monardes Formations-Figure 1.4-4).

During the late Cretaceous-Paleocene, the active volcanic arc migrated to a position in the present day Precordillera, during which time a series of volcanic centres were developed (Figure 1.4-4), before the active arc once again migrated during the Eocene-Oligocene, towards its present day location in the High Andes. Subsequent to this the Coastal Cordillera-Precordillera have shared a common history of erosion, resulting in the deposition of the Atacama gravels during Miocene-recent uplift (Figure 1.4-4).

1.5 The Arica Deflection

One of the major morphological features of the South American margin is the ~55° change in strike (from NW (NNE) to north (south)), reflected in the orientation of the subduction trench, coastline and topography (Figure 1.1-1), as well as of the gross geological strike. The point of inflection is located near to the Peru-Chile border at ~19°S and is variously known as the Arica Deflection or Arica Bend, with the gross structure, incorporating deflection and limbs to the north and south often
referred to as the Bolivian Orocline. It is unclear whether or not the deflection represents a primary (inherited) feature that predates Andean orogeny, or if the on-going process of subduction has deformed an initially (near-) straight continental margin, to create a secondary bend. Palaeomagnetism offers a way of testing the origin of the Arica Deflection and many studies have now been undertaken to the north and south of the Arica deflection, in an attempt to address this question.

The distribution of crustal rotations determined from palaeomagnetic studies from the Central Andes will be introduced in Chapter Three, as will many of the models proposed to explain the observed pattern of rotations prior to the completion of this study. The relative merits of these models will be discussed further with regard to constraints placed by palaeomagnetic data from this study in Chapter Seven, before a preferred rotation model is proposed with regard to the spatial and temporal accumulation of crustal rotation.

1.6 Selection of Field Areas

This thesis comprises a study of palaeomagnetically determined crustal rotations in northern Chile. Palaeomagnetic data has been collected from three localities situated in the present day forearc of northern Chile between 27-30°S, with data from two other studies reinterpreted with respect to recently published maps [Palmer et al., 1980a; Riley et al., 1993] (Figure 1.6-1). The magnitude of crustal rotations from these areas is interpreted with respect to the overall rotation pattern observed throughout the Andean margin. In order to test the rotation models proposed for the Andean Margin convincingly, the full areal extent of crustal rotations needs first to be established. Although there is now palaeomagnetic data spanning much of the forearc region of northern Chile, there are still a number of latitudinal "gaps", in the overall Andean palaeomagnetic dataset, some of which
Figure 1.6-1 Sketch maps indicating the location of the sampling areas from this thesis. A-Study area of Palmer et al., (1980a), B-Tres Cruces field area, C-La Guardia field area, D-Pabellón field area.
span areas of noticeable change in the magnitude of observed crustal rotation. It was primarily to address one such a “gap”, that this study was initiated.

1.6.1 Tres Cruces Study Area

The initial premise of this study was to address deficiencies within the overall Andean palaeomagnetic dataset through targeted sampling, with the intention of establishing the southernmost limit of the Central Andean Rotation Pattern (CARP). The Tres Cruces study area (c.29°30’S) was primarily chosen on the basis of existing palaeomagnetic data, located to the east of the city of La Serena (c.30°00’S) [Palmer et al., 1980a-Figure 1.6-1], which indicates that CW crustal rotations at this latitude are of markedly lower magnitude than observed further to the north at ~28°30’S [Gipsen, unpublished data]. The implication of this is that the magnitude of crustal rotation decreases southwards, but no data exists between these two areas to test this hypothesis. The Tres Cruces study area spans the Coastal Cordillera-Precordillera boundary and was chosen to address this gap in the overall Andean palaeomagnetic dataset, as it potentially afforded access to strata of a similar age to that sampled to the north and south.

1.6.2 La Guardia Study Area

After analysis, palaeomagnetic data from the Los Choros area suggested that although the maximum age of rotation was well constrained, no reasonable estimate could be inferred concerning the minimum age of CW rotation in the Northern Chilean Forearc. As various processes have operated at different times during the evolution of the central Andean margin, any or all of these processes could potentially have driven, or otherwise have affected the magnitude of crustal rotation. In order to try to reconcile the observed rotation pattern with a single process or event, the timing of rotation within the Andean margin, as a whole
needs to be more precisely determined. It was felt that this could be addressed through the sampling of wide age-range of material in as small an areas as possible and the La Guardia area (c.27°30'S-Figure 1.6-1) was chosen for such a purpose as it provided excellent access to Triassic to Eocene aged rocks, as well as to the uplifted Palaeozoic basement (although this was not sampled).

1.6.3 Study Aims and Objectives

The basic aims of this study can be summarised as follows:

1. To sample Mesozoic and Tertiary aged units in the forearc region of northern Chile and identify stable characteristic palaeomagnetic remanence directions for each sampling unit.

2. To determine the magnitude of crustal rotation recorded by each sampling unit, in relation to cratonic South America.

3. To attempt to rationalise observed crustal rotations with respect to the structural setting of each field area.

4. To confirm the spatial limits of the overall rotation pattern suggested by the existing palaeomagnetic database and determine the timing of crustal rotation.

5. To consider the implications of crustal rotations determined from each field area on the overall "Central Andean Rotation Pattern" and hence evaluate previously suggested rotation models.

6. To suggest a geologically plausible model to account for both the extent and timing of observed crustal rotations throughout the Andean margin.

1.7 Thesis Structure

The palaeomagnetic methods used during this study and the concepts and assumptions concerning the understanding of the geomagnetic field upon which
they are based, are outlined in Chapter Two, with a discussion of the relationship between palaeomagnetically observed crustal rotations and crustal deformation, with regard to studies around the world forming Chapter Three. In addition the existing pattern of crustal rotations within the Andes and the reference poles used in their calculation, are reviewed in Chapter Three and many of the models proposed to explain the overall central Andean rotation pattern are introduced. The palaeomagnetic data collected during this study is presented and interpreted in Chapters Four (Tres Cruces), Five (La Guardia) and Six (magnetostratigraphic profile through the Pabellón Formation), with the rotation pattern from each field area and implications concerning the overall Andean dataset discussed in Chapter Seven.
Chapter Two

Palaeomagnetic Sampling of the Ancient Magnetic Field

This chapter introduces some of the fundamental concepts concerning the study of the ancient geomagnetic field and the practical basis of palaeomagnetic measurements.

2.1 The Geomagnetic Field

The geomagnetic field may be specified at any point on the Earth’s surface by measurement of the total intensity (F), declination (D) and inclination (I) (Figure 2.1-1) [McElhinny & McFadden, 1999]. The declination (D) represents the deviation of a compass needle from true (geographic) north (+ve eastwards). A compass needle aligns itself to magnetic north, which lies within the vertical plane containing the total magnetic field (F) known as the magnetic meridian. A perfectly balanced compass needle free to swing within the magnetic meridian (shaded in Figure 2.1-1), would take up a position at an inclined angle to the horizontal known as the inclination (I). The inclination is positive when the north-seeking end of a compass needle points downwards (i.e. present day northern hemisphere) and negative upwards (southern hemisphere).

The total magnetic field (F) can be split into horizontal and vertical components, (H & Z), with the vertical component (Z) considered positive downwards. The intensity of the Earth’s magnetic field is measured in Tesla (T), although the maximum value of the Earth’s magnetic field is currently only 70 µT, with variations to the field often only of the scale of nanotesla (nT).

Through the experimental work of Gilbert in the 17th century and later Gauss (1839), it was recognised that the Earths’ magnetic field (at the surface), can be
Figure 2.1-1 Description of the geomagnetic field direction at any point on the Earth's surface [redrawn after McFadden, 1996].

Geographic North pole

North magnetic pole ($\theta = 90^\circ$)

Geographic equator

Magnetic equator ($\theta = 0^\circ$)

South magnetic pole ($\theta = -90^\circ$)

Geographic South pole

Geomorphic South pole

Figure 2.1-2 The best-fitting geocentric axial dipole (GAD), geographic and magnetic poles [from McElhinny & McFadden, 1983].

Figure 2.1-3 Variation in the direction of the magnetic field over the past 400 yrs, measured at Greenwich Observatory, London, U.K. [from Butler, 1992].
well approximated by a magnetic dipole positioned at the centre of the Earth's core. The best fitting dipole is inclined 10.5° to the axis of rotation (Figure 2.1-2) and accounts for approximately 90% of the observed overall magnetic field at the surface, with the remaining 10% being the non-dipole field component [Merrill & McElhinny, 1983]. The non-dipole field itself can be modelled as the result of a combination of higher order quadrupole and octupole field contributions.

Palaeo-secular variation (PSV) and the GAD hypothesis

Although the morphology of the Earth's current magnetic field is best defined as resulting from a geocentric dipole inclined at an angle to the Earth's axis of rotation. The actual direction and magnitude of the surface geomagnetic field at any fixed position has been observed to vary with time. Both dipole and non-dipole components of the field vary on a range of timescales. Long-term variation in the geomagnetic field is referred to as secular variation and represents the deviation from the best fitting magnetic dipole, believed to be a function of fluid motion in the outer core. Over even longer periods, the polarity of the geomagnetic field is observed to reverse when field directions are observed to change through 180°.

Secular variations of the geomagnetic field of a few degrees per century were recorded at Greenwich Observatory (London), where declination varied from 11.5°E in 1576 to 24°W in 1823 before turning eastward again (Figure 2.1-3) [Butler, 1992]. Studies of ancient secular variation are carried out mainly in high fidelity lake sediments [e.g. Turner & Thompson, 1981, 1982] (Figure 2.1-4). Such studies are referred to as studies of palaeo-secular variation (PSV) and demonstrate long period oscillations in the recorded geomagnetic field direction.

In one of the first reviews of Palaeomagnetism, Cox & Doell (1960) concluded that;

"...the Earth's time-averaged magnetic field was closely approximated by a Geocentric Axial Dipole (GAD) field from the Oligocene to the present...although
Figure 2.1-4  Secular variation of the magnetic field recorded by lake sediments in the UK [from Tarling, 1982; after Turner & Thompson, 1981, 1982].

Figure 2.1-5  A-The variation of the inclination of a dipolar magnetic field with latitude, observed for a magnetised sphere by William Gilbert [from Stern, 2002]. B-The GAD model [after McElhinny & McFadden, 1999].
The GAD (Geocentric Axial Dipole) theory extends the idea of uniformitarianism to the Earth's magnetic field. Although the observed magnetic north pole is observed to wander around geographic north (Figures 2.1-3 & 4), when sampled over a suitably long period of time, the ancient geomagnetic field can be modelled as resulting from a magnetic dipole placed at the centre of the Earth (Geocentric), and aligned with the rotation axis (Axial).

William Gilbert was the first person to show how latitude (λ) could be derived from the dip of the magnetic field. Through his experiments with lodestone 'Terrellae' or 'Little Earths', he observed that the orientation of magnetised spikes within the 'Orbis Virtualis' or sphere of influence was similar to that observed for the Earth itself (Figure 2.1-5a). The extension of Gilbert's model to the Earth (Figure 2.1-5b, where a = radius of the Earth and all other parameters indicated are as displayed in Figure 2.1-1) is, in essence, the Geocentric Axial Dipole model.

The assumption that the time-average geomagnetic field can be produced by a single geocentric axial dipole (GAD) field, is crucial to the interpretation of palaeomagnetic data, as it provides a reference frame by which palaeolatitude and therefore palaeogeographic reconstructions may be constrained [Butler, 1992]. One of the primary implications of the GAD hypothesis is that the inclination of the magnetic field is intrinsically related to latitude via the relationship

\[ \tan I = 2 \tan \lambda \]

This is often referred to as the "Dipole Equation", where λ is the geographic latitude. It should also be noted for a GAD field that declination (D) would equal zero, everywhere, as magnetic north coincides with geographic north [Butler, 1992].
The GAD model takes no account of PSV, as the effects should be averaged out as a consequence of sampling a wide age (temporal) range of material. PSV has the effect of creating statistical scatter in palaeomagnetic data where each measurement should represent a separate and instantaneous record of the ancient geomagnetic field.

How appropriate is the GAD hypothesis?

There is much ambiguity concerning the amount of time over which the geomagnetic field will satisfactorily approximate that of a GAD field. In one of the most recent discussions of the GAD field assumption, Merrill & McFadden (2003) conclude that in the absence of excursions in the ancient geomagnetic field, the length of time required to obtain a good GAD field approximation could be as little as $10^4$ years, although a preferred interval of $10^5$ years is suggested. Tauxe (2005) points out that at any particular instant in time over the past 5 Ma, the geomagnetic field may have departed significantly from that of a GAD. This is mainly due persistent long-term non axial-dipole field components [Merrill & McFadden, 2003].

If non-dipole contributions to the ancient geomagnetic field are considered to have been significant, then the main tenet of the GAD assumption is seriously undermined. Tauxe (2005) suggests that deviations from the GAD, previously argued to require the greater influence of a non-dipole contribution to the ancient geomagnetic field, can also be explained by distortions in the actual palaeomagnetic recording mechanisms (such as inclination flattening due to sedimentary processes). This would create a shallow bias in the time averaged geomagnetic field.
True Polar wander

An important aspect of all palaeomagnetic studies that is that all measurements are made using the geographic poles (the Earth's spin axis) as the main frame of reference. The concept of polar wander in plate tectonic studies tracks the (apparent) motion of the geographic reference frame for a stationary continent. In contrast, true polar wander or TPW, on the other hand, describes the displacement of the entire earth (or of an outer shell) with respect to the Earth's axis of rotation [Goldreich & Toomre, 1969; Gordon, 1987]. It is believed that TPW is controlled by those variations in the surface gravity equipotential, unrelated to the equatorial bulge of the Earth. Known as the nonhydrostatic geoid, these variations are related to internal mass anomalies and have been interpreted to control long-period TPW [McElhinny & McFadden, 1999]. Evans (2001), describes that [after Ricard et al., 1993]:

"True polar wander arises from centrifugal forces acting on mass anomalies either on the surface or within the body of a quasi-rigid planet. In simplified terms, excess masses are driven to the equator, mass deficiencies are driven to the pole, and the instantaneous pole location is determined by the integrative effects of all the planet's mass anomalies, taking into account complexities such as mantle viscosity variations."

A thorough discussion of TPW is beyond the scope of this thesis, although it is noted that work by McElhinny (1973) and Jurdy & Van der Voo (1974, 1975) demonstrate that little significant TPW is recorded for the past 50 Ma. In addition Evans (2001), suggests that although TPW is a significant geodynamic process which, in terms of continental motions, may even dominate plate tectonics for certain intervals of Earth history, very little TPW has been recorded during the past 200 Ma. As a consequence of this, the effects of TPW are not considered during this study.
2.2 Properties of Magnetic Grains

2.2.1 Diamagnetism & Paramagnetism

All substances exhibit some form of magnetism as any moving charge (such as an orbital electron for example), experiences a force when placed within a magnetic field producing a negative magnetic moment, known as diamagnetism [Butler, 1992]. Should an atom possess an overall resultant magnetic moment the application of a magnetic field will align these atomic dipole moments in a direction parallel to the applied field. Known as paramagnetism, this effect tends to swamp the diamagnetic effect [McElhinny & McFadden, 1999], inducing a magnetisation in the direction of the applied field.

In metallic elements another mechanism also induces paramagnetism. Where the outer valence electrons are free to move throughout the solid metal, the application of a magnetic field will align these free electrons parallel to the magnetic field, of which equal numbers will have opposite spins, inducing an overall dipole moment.

2.2.2 Ferromagnetism, Antiferromagnetism & Ferrimagnetism

Pure diamagnetic and paramagnetic substances are only weakly magnetic because the dipole moments involved are small. Certain (transition) metallic elements such as iron, nickel and cobalt however, display very strong magnetic effects through the phenomenon of ferromagnetism [Butler, 1992]. The ferromagnetic effect results from the fact that the individual atoms and their inner electron orbital electrons are much closer together than the virtual radii of the valence electrons when compared to other paramagnetic metallic elements. There are also a far greater number of free outer valence electrons, which therefore are more closely spaced and as a consequence react strongly with each other. The
energy exchanged between these electrons forces their spins into alignment, even in the absence of a magnetic field, and so ferromagnetic substances exhibit a spontaneous magnetisation and hence may retain a permanent magnetic dipole [McElhinny & McFadden, 1999].

With increasing temperature the greater amount of thermal agitation disrupts the electron spin alignment process. At a critical temperature, known as the Curie temperature, ferromagnetism is completely destroyed and the spontaneous magnetisation is reduced to zero [Butler, 1992]. At temperatures above the Curie point a substance behaves paramagnetically.

Ferromagnetism in the broadest sense refers to the existence of a spontaneous magnetisation in the absence of an applied magnetic field and this can be derived in several ways. Ferromagnetism in the strictest sense refers to a substance where the electron spin moments in each lattice are aligned in the same direction (Figure 2.2-1a). Where two sub-lattices (usually termed the A & B lattices) are present, the electron spin moments are aligned but antiparallel. If the moments of each sub-lattice are equal, then the ferromagnetic effects cancel each other (Figure 2.2-1b), and there is no net magnetic moment. This is referred to as antiferromagnetism and substances displaying this behaviour do not possess a Curie temperature, but the atomic moment ordering is destroyed at a critical temperature known as the Néel temperature [McElhinny & McFadden, 1999].

In some cases, the atomic moments of the A & B sub-lattices may be unequal (ferrimagnetic-Figure 2.2-1c), or the equal magnetic moments of the two sub-lattices may not be exactly antiparallel (canted antiferromagnetic-Figure 2.2-1d), with both cases resulting in a net spontaneous magnetisation. Both ferrimagnetic and canted antiferromagnetic substances display all the characteristics of a
Ferromagnetic

Antiferromagnetic

Ferrimagnetic

Canted antiferromagnetic

Resultant spontaneous magnetisation

None

Figure 2.2-1 Exchange-coupled spin structures and resulting spontaneous magnetisation (from McElhinny & McFadden, 1999). Arrows indicate the direction of the total magnetic moment resulting from the alignment of electron spins within each lattice or sub-lattice of a "ferromagnetic" mineral.

Figure 2.2-2 Uniformly magnetised sphere with saturation magnetisation, $j_s$ (A), and the associated internal demagnetising field $H_D$ (B). From Butler (1992).

Figure 2.2-3 Shape anisotropy (from Butler, 1992). $H_D$ is minimised when an elongate grain is magnetized along the long axis or "easy" direction (A) and reaches a maximum along the short axis or "hard" direction (B).
ferromagnetic mineral (such as a Curie temperature) and henceforth the term ferromagnetism will refer to this 'gross' ferromagnetic behaviour.

2.2.3 Properties of Ferromagnetic Grains

The properties of a ferromagnetic grain are dependent upon many factors including composition, shape, size and ambient temperature.

*Magnetostatic energy and shape anisotropy*

For the uniformly magnetised (spherical) grain shown in Figure 2.2-2a, one hemisphere has positive and the other negative charge. The magnetic charges of adjacent atoms internally cancel but produce a magnetic charge distribution at the surface of the particle. Due to repulsion between adjacent charges, magnetostatic energy ($e_m$), is stored within this distribution [Butler, 1992]. The magnitude of $e_m$ is proportional to the square of the magnetisation of a grain ($j^2$) and is therefore extreme for grains with particularly high saturation magnetisations ($j_s$). Ferromagnetic grains therefore attempt to acquire magnetisation in such a way to minimise $e_m$.

The magnetic charge distribution also produces a magnetic field internal to a grain, referred to as the demagnetising field ($H_D$) as it generally opposes the direction of magnetisation (Figure 2.2-2b). The magnitude of the internal demagnetising field is related to shape anisotropy, such that a particle may be more easily magnetised along certain directions [McElhinny & McFadden, 1999]. Elongate grains are more easily magnetised along their long (easy) axis because this minimises the magnetostatic energy of a grain (Figure 2.2-3a). When an elongate grain is magnetised along the easy axis, the magnetostatic energy of the grain is minimised [McElhinny & McFadden, 1999].
If the same elongate grain is magnetised at right angles to its long axis the magnetostatic energy will be large (Figure 2.2-3b) [McElhinny & McFadden, 1999]. The short axis of elongate grains is often referred to as the “hard” magnetic direction, as it is more difficult to magnetise the grain in this direction because of the large magnetostatic energy that results.

Magnetic domains and grain size

The magnetostatic energy of larger grains can be reduced through the formation of magnetic domains. These are regions of opposite magnetisation that are separated by walls within which magnetic energy is stored, and grains with many domains are referred to as multi-domain (MD) grains. The number of domains will increase until the energy required to create a new wall surpasses the subsequent reduction in magnetostatic energy [McElhinny & McFadden, 1999]. Below a critical grain size, the formation of magnetic domains is energetically unfavourable and these grains are referred to as single domain (SD) grains (Figure 2.2-4).

As the magnetostatic energy of a ferromagnetic grain is related to the saturation magnetisation of that grain, minerals with low js such as haematite will have little cause to develop numerous magnetic domains for grains of <15μm in diameter. For this reason, much of the naturally encountered haematite is SD. For minerals with high js, such as magnetite, only fine-grained particles are SD, although as shown in Figure 2.2-4, very elongate grains (with low width/length ratios), also behave as SD grains [McElhinny & McFadden, 1999; after Newell & Merrill, 1999].

There isn’t a sharp boundary that defines the grain dimensions at which the formation of domain walls becomes favourable and Stacey (1962), proposed that grains with only a very small number of domains, termed pseudo-single-domain (PSD), behave in a similar manner to SD grains, with similar coercivity and time
Figure 2.2-4  Superparamagnetic and SD fields, defined for rectangular parallelepiped magnetite grains as compiled by Newell & Merrill (1999) [from McElhinny & McFadden, 1999].

Figure 2.2-5  Hypothetical change in relaxation time $\tau$, as a function of grain volume (V) [from McFadden & McElhinny, 1983].
stability spectrum with respect to carrying remanence. The maximum remanence observed for SD magnetite particles ($L_{SD}^*$), occurs at the SD-MD transition (Figure 2.2-4).

Relaxation time

In weak magnetic fields, the magnetic remanence carried by a group of ferromagnetic grains, will decay over time as thermal vibrations cause magnetic domains to align along an easy direction that has a component in the direction of the applied field [Tarling, 1982]. At a constant temperature, this will occur at a rate known as the characteristic relaxation time, $\tau$, which is related to the volume, coercivity and saturation magnetisation of a ferromagnetic grain. The theoretical change in the relaxation time of a grain with increasing volume is illustrated in Figure 2.2-5, with SD, PSD and MD grains indicated. Particles with large relaxation times will retain an acquired magnetisation for very long periods of time and are considered as effective palaeomagnetic recorders if the relaxation time is $>>$ than the age of the magnetisation.

Small (and more equant), SD and large MD grains possess very short relaxation time, meaning although they may acquire magnetisation, it will decay extremely quickly, often on a scale of seconds. Such behaviour is referred to as superparamagnetism and SD particles displaying such behaviour can be highly magnetically susceptible and retain very large magnetisations. A commonly stated threshold for defining superparamagnetic grains is $\tau = 100$secs [e.g. Tarling, 1982; Butler, 1992; McElhinny & McFadden, 1999] (see Figure 2.2-4).

The longer relaxation time and higher coercive forces associated with equant or elongate SD and PSD magnetite grains (Figure 2.2-4), mean that stable magnetic remanence is generally most effectively carried by particles of these domain
states. The much shorter relaxation times of superparamagnetic and MD grains mean that particles of this size/domain structure, will often only retain secondary viscous magnetisations associated with the most recently applied field.

2.3 Magnetic Minerals

The fact that many rocks are found to preserve a fossilised record of the ancient geomagnetic field is due to a small number of naturally occurring minerals that display ferromagnetic properties (Table 2.3-1). The two most important mineral groups form part of the iron-titanium oxide (FeO-TiO₂-Fe₂O₃) ternary system (Figure 2.3-1) [McElhinny & McFadden, 1999]. These are the cubic titanomagnetites, a solid solution series between magnetite (Fe₃O₄) and ulvöspinel (Fe₂TiO₄) and the more weakly magnetic solid solution series between haematite (αFe₂O₃) and ilmenite (FeTiO₃), known as the titanohaematites. In addition to the titanomagnetites and titanohaematites, several other minerals are often found to carry magnetisation within rocks and include iron sulphides (e.g. greigite (Fe₃S₄) and pyrrhotite (Fe₁₋ₓS, where 0 < x ≤ ½)), and the iron oxyhydroxides (e.g. goethite (αFeOOH) and lepidocrocite (γFeOOH), collectively referred to as limonite).

Titanomagnetites

The titanomagnetites are cubic members of the spinel group with the metal cations located in two (unequal) interacting sub-lattices giving rise to the observed ferrimagnetism. The Curie point of magnetite (Fe₃O₄) is 578°C, and this decreases as the proportion of ulvöspinel (Fe₂TiO₄) increases [Hunt et al., 1995]. Ulvöspinel is paramagnetic at room temperature, but antiferromagnetic at very low temperatures with a Neél temperature of -153°C (Figure 2.3-1) [McElhinny & McFadden, 1999].
<table>
<thead>
<tr>
<th>Mineral</th>
<th>Composition</th>
<th>Magnetic state</th>
<th>Ms (10^3 \text{ Am}^{-1})</th>
<th>Tc (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magnetite</td>
<td>Fe(_3)O(_4)</td>
<td>Ferrimagnetic</td>
<td>480</td>
<td>580</td>
</tr>
<tr>
<td>Titanomagnetite (TM60)</td>
<td>Fe(<em>{2.4})Ti(</em>{0.6})O(_4)</td>
<td>Ferrimagnetic</td>
<td>125</td>
<td>150</td>
</tr>
<tr>
<td>Ulvospinel</td>
<td>Fe(_2)TiO(_4)</td>
<td>Antiferromagnetic</td>
<td>-153</td>
<td></td>
</tr>
<tr>
<td>Haematite</td>
<td>Fe(_2)O(_3)</td>
<td>Canted Antiferromagnetic</td>
<td>2.5</td>
<td>675</td>
</tr>
<tr>
<td>Ilmenite</td>
<td>Fe(_{2})TiO(_3)</td>
<td>Antiferromagnetic</td>
<td>-233</td>
<td></td>
</tr>
<tr>
<td>Maghemite</td>
<td>Fe(_2)O(_3)</td>
<td>Ferrimagnetic</td>
<td>380</td>
<td>590-675</td>
</tr>
<tr>
<td>Pyrrhotite</td>
<td>Fe(_{1-x})S((0&lt;x&lt;1/8))</td>
<td>Ferrimagnetic</td>
<td>80</td>
<td>320</td>
</tr>
<tr>
<td>Greigite</td>
<td>Fe(_3)S(_4)</td>
<td>Ferrimagnetic</td>
<td>125</td>
<td>330</td>
</tr>
<tr>
<td>Goethite</td>
<td>FeOOH</td>
<td>Antiferromagnetic with</td>
<td>2</td>
<td>120</td>
</tr>
<tr>
<td></td>
<td></td>
<td>defect ferromagnetism</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Iron</td>
<td>Fe</td>
<td>Ferrimagnetic</td>
<td>1715</td>
<td>765</td>
</tr>
<tr>
<td>Cobalt</td>
<td>Co</td>
<td>Ferromagnetic</td>
<td>1422</td>
<td>1131</td>
</tr>
<tr>
<td>Nickel</td>
<td>Ni</td>
<td>Ferromagnetic</td>
<td>484</td>
<td>358</td>
</tr>
</tbody>
</table>

Table 2.3-1 Magnetic properties of some commonly occurring magnetic minerals [after McElhinny & McFadden, 1999].

Figure 2.3-1 Iron oxide ternary diagram displaying the three primary solid solution series found in igneous rocks. As all members of the pseudobrookite series behave paramagnetically above room temperature, they are considered to be insignificant in carrying remanence. Approximate Curie (or Néel) temperatures are indicated for the titanomagnetite and titanohaematite series based on the mole fraction of end-members \(x\) & \(y\) at 0.2 intervals [from McElhinny & McFadden, 1999].
Generally speaking, unless an igneous rock is cooled extremely rapidly (e.g. an oceanic basalt quenched by seawater), single-phase titanomagnetites of an intermediate composition are not preserved. This is because TM grains undergo deuterio oxidation through solid solution at high temperatures (600-1000°C), resulting in compositions tending towards the magnetite (TM₀) and ulvöspinel (TM₁₀₀) (or ilmenite) under normal cooling conditions [Tarling, 1983]. Continued deuterio oxidation (due to very slow cooling) can result in the total oxidation of TM to pseudobrookite and haematite and/or rutile [McElhinny & McFadden, 1999]. The ultimate product of the low temperature alteration of magnetite, through the weathering process for example, is maghemite (γFe₂O₃), which retains the structure of magnetite but has the chemical composition of haematite [Tarling, 1983]. As such the magnetic properties of maghemite are very similar to magnetite, but the mineral is destroyed on heating to approximately 350°C, when it reverts to haematite, accompanied by a very large decrease in spontaneous magnetisation (~99.6%) [Tarling, 1983].

Titanohaematites

The titanohaematites are represented by the complete solid solution (above 1050°C) between haematite (αFe₂O₄) and ilmenite (FeTiO₃) (Figure 2.3-1). Haematite displays canted antiferromagnetism (Table 2.3-1), with the oppositely magnetised Fe³⁺ ions in the A & B sub-lattices canted at a slight angle (~0.2°). This gives rise to a weak spontaneous magnetisation, ~0.5% of that of magnetite [McElhinny & McFadden, 1999]. Haematite has a Curie point of 675°C [Hunt et al., 1995], which coincides with the Néel temperature and decreases as the proportion of ilmenite increases, with the Néel temperature of pure ilmenite being ~223°C (Figure 2.3-1).
Apart from resulting from the high temperature deuterio oxidation of titanomagnetite, or from the inversion of maghemite, haematite is also an oxidation product of magnetite at room temperature and formed during a number of secondary processes, such as the dehydration of weathering products like goethite. Importantly for palaeomagnetic studies, haematite occurs as a fine pigment or cement in clastic sediments, producing the distinctive colouration of red beds, which have provided a wealth of palaeomagnetic sampling material around the globe.

Iron Sulphides

The dissolution of iron oxides in the iron reduction zone in deep-sea sediments leads to the production of iron sulphides, predominantly of non-magnetic pyrite. Under sulphate reducing conditions (such as in muds and rapidly deposited deep-sea sediments), the magnetic sulphides greigite and pyrrhotite may form and be preserved [McElhinny & McFadden, 1999]. Greigite has the same inverse spinel structure as magnetite and is ferromagnetic with a spontaneous magnetisation ~25% of that of magnetite and a Curie temperature of ~330°C (Table 2.3-1). Pyrrhotite is a common accessory mineral in many rocks, especially in igneous rocks derived from sulphur rich magmas or a hydrothermal alteration product, and has a Curie point of 320°C (Table 2.3-1). Above 500°C pyrrhotite transforms irreversibly to magnetite, and at higher temperatures to haematite [Butler, 1992].

Iron Oxyhydroxides

The most important iron oxyhydroxides is goethite, a weathering product common in soils and sediments. Although antiferromagnetic (Néel temperature of 120°C- [Hunt et al., 1995] (Table 2.3-1), it has a weak superimposed parasitic ferromagnetism with a Curie temperature coinciding with the Néel temperature
Goethite dehydrates to form haematite between 250-400°C.

2.4 Acquisition of Magnetisation

Natural Remanent Magnetisation (NRM)

As discussed in the previous section, a number of minerals are capable of retaining a permanent magnetisation. In the field, the in-situ magnetisation of a rock unit is the sum of the induced (transient and parallel to the local geomagnetic field), and the remanent (permanent) magnetisations [Butler, 1992]. In palaeomagnetic studies only the remanent magnetisation is of interest. Depending on magnetic mineralogy, the majority of rocks can record the direction of the ancient geomagnetic field at their time of formation, and this is referred to as the primary magnetisation. Subsequent to the acquisition of a Primary magnetisation, a rock may be wholly or partially remagnetised at a later date. This often results in the overall remanence of a sample being composed of two or more components of magnetisation and the net Natural Remanent Magnetisation (NRM) is the vector sum of primary and all secondary magnetisations.

The three basic forms of primary NRM are (a) thermoremanent magnetisation (TRM), acquired as rocks cool from high temperatures; (b) detrital remanent magnetisation (DRM), acquired during the formation of sedimentary rocks; and (c) chemical remanent magnetisation (CRM), acquired as ferromagnetic minerals are chemically precipitated within a magnetic field [Butler, 1992].

Thermo-remanent Magnetisation (TRM)

Igneous rocks are derived from a cooled molten magma source either at some depth or after extrusion at the surface. Such rocks must eventually cool through
the Curie point of any magnetic minerals formed from the magma, at which point a spontaneous magnetisation is acquired parallel to the ambient geomagnetic field and is referred to as a thermo-remanent magnetisation (TRM). TRM is not acquired only at the Curie point, but across a broad range of blocking temperatures down to the ambient temperature, associated with the range of grain volumes. As each fraction of grains of similar volume passes through the associated blocking temperature, the relaxation time of the grains (i.e. the amount of time it takes for a grain to realign itself with a different magnetic field) rapidly increases so that the magnetisation becomes ‘frozen’. Any subsequent changes in the field direction at lower temperatures will have no effect in changing the direction of magnetisation. The very long relaxation times associated even with grains possessing relatively low blocking temperatures means TRM is considered to be stable over geological time [McElhinny & McFadden, 1999].

If a rock unit is heated at some point during it's history after formation, the blocking temperatures of some or all of the grains may be reached or exceeded, reducing the relaxation time for these grains. This means that a component of the primary magnetisation may be reset to record a secondary component of magnetisation parallel to the then ambient geomagnetic field direction. The degree to which primary magnetisation is reset is a function of temperature reached and duration of the heating event, as well as being dependent on the distribution of grain volumes.

Detrital Remanent Magnetisation (DRM) & post-Depositional Remanent Magnetisation (pDRM)

The process by which detrital magnetic particles are aligned to the geomagnetic field during sedimentary deposition is termed detrital (or depositional) remanent magnetisation (DRM), although DRM is not ‘locked-in’ until the sediment has been compacted by overlying material and all water is driven out. Although most mass
transported detrital (allochthonous) sediments are deposited in sub-aqueous environments, in the following discussion, the processes involved in the sub-aerial deposition of sediments can be considered equivalent, with air taking the place of water. A typical sediment load will contain detrital magnetic grains of various size, composition and shape, associated with the source rock and weathering and transport processes that have taken place. The stability of magnetisation is dependant on the nature of the detrital grains and not the actual depositional processes [Tarling, 1983].

The magnetic grains are magnetised prior to deposition and will be acted on by the aligning force of the Earth's magnetic field as they fall through the water column, as well as by gravitational and hydrodynamic forces. As a particle reaches the water-sediment interface, it will come to rest in the most stable gravitational position, with the long axis of elongate particles parallel to the sediment surface and usually aligned parallel or perpendicular to the water flow, imparting a fabric to many sediments [Butler, 1992].

Laboratory experiments indicate that the alignment of larger and elongate particles (which are usually magnetised along their longer axes), parallel to the sediment-water interface, has the effect of causing a shallowing of the net magnetisation towards the horizontal (Figure 2.4-1-Tarling 1983). Inclination shallowing of up to 20° has been observed compared to the ambient field inclination [King, 1955], although this amount of error does not appear to be as pronounced in natural sediments.

The lack of inclination flattening in natural sediments may be due to the effects of the post-depositional rotation of magnetic particles towards the geomagnetic field within pore spaces (Figure 2.4-1). Rapidly deposited and or low porosity sediments are likely to record more significant amounts of inclination shallowing.
Figure 2.4-1  The acquisition of depositional remanent magnetisation and post-depositional magnetisation [from Tarling, 1982].
This mode of acquisition of magnetisation is referred to as Post-Depositional Remanent Magnetisation (pDRM) and occurs most efficiently when the magnetic grains are significantly finer than the silicate grains [McElhinny & McFadden, 1999].

Crystallisation (Chemical) Remanent Magnetisation (CRM)

A second way in which sediments can acquire a primary magnetisation involves the chemical precipitation of magnetic minerals at low temperatures in the geomagnetic field. As sediments accumulate, the lower layers become increasingly compacted, pore space reduced and water expelled. This leads to changes in the physical/chemical environment and many detrital minerals will undergo diagenesis as a result.

Ferromagnesian minerals like olivine, pyroxene and amphibole generally break down to produce iron-rich chlorite clays whilst feldspars decompose to illite and kaolinite. Mobile iron ions released during this process form a series of compounds that often react to form iron hydroxides, or if there is significant microbial activity for example, will combine with sulphur compounds to form iron sulphides [Tarting, 1982]. Detrital titanomagnetites can also transform to maghemite.

One of the most widespread processes leading to the acquisition of a stable CRM is through the development of pigmentory haematite that gives continental red-beds their distinctive colouration. The actual process by which fine-grained haematite forms is speculative at best and the actual voracity of palaeomagnetic data collected from red-beds has been the source of much controversy in the past.
Viscous & Thermoviscous Remanent Magnetisation (VRM & TVRM)

Assemblages of superparamagnetic grains can maintain strong magnetisations, although they are very unstable and will quickly decay on removal from the magnetising field [Butler, 1992]. If a significant proportion of the ferromagnetic grains present within a sample are in the superparamagnetic grain size range, then a significant proportion of sample NRM may be carried by this fraction. This proportion of NRM however will not be geologically stable and is likely to represent a record of the last magnetic field the sample was placed within.

2.5 Collection of Palaeomagnetic Sample Material

Collection of orientated sample material

Palaeomagnetic laboratory work is usually carried out using cylindrical specimens prepared from independently orientated core or block samples, collected from a number of sites within a rock unit [Collinson, 1983]. During the course of this study, only core samples were collected in the field. A portable drill, equipped with Ø 2.54cm diamond tipped water-cooled drill-bits, was used to drill individual core samples. Standard practise concerning the actual collection of sample material, generally suggests that 6-8 core samples are collected from a number of geographically or geologically distinct sites within the chosen rock unit.

A palaeomagnetic sampling site usually comprises an individual lava flow, a short sedimentary sequence, or small area of an igneous intrusion or individual dyke, with the overall mean site magnetisation considered to represent an effectively instantaneous record of the local magnetic field, assuming that all samples within a site share a common magnetisation and deformation history [Butler, 1992]. The actual number of sites collected is generally limited by the amount of time available in the field area and the quality of the outcrop, with the collection of at
least six sites considered adequate, but ten or more sites preferred for most purposes.

The orientation of individual core samples was measured in-situ, using both standard magnetic and solar compasses using standard operating procedures (see Figures 2.5-1a and b). Each individual core sample was cut into 2.2cm long specimens, with a resulting volume of ~11cm³ (Figure 2.5-2). These dimensions are chosen to closely approximate those of a unit sphere. Often the length of the recovered core is only sufficient to allow the preparation of a single sample, while longer cores may provide up to three specimens denoted A, B & C, with the A sample taken from the base of the core sample (Figure 2.5-2).

The hierarchical manner in which an overall rock unit is sampled serves two purposes. Firstly, the geological noise, inherent within the palaeomagnetic recording mechanisms discussed in Section 2.4, can be reduced through averaging the observed components of magnetisation at the specimen, sample, site then overall rock unit level. Secondly, by measuring instantaneous records of the ancient geomagnetic field at a number of sites within a rock unit, the effects of PSV are averaged and the overall direction of magnetisation recorded by a rock unit is assumed to have been acquired within a GAD field (Section 2.1).

Field Tests for Stability

The primary objective of this palaeomagnetic study was to determine the magnitude of crustal rotation recorded within a field area and to establish the timing of any observed rotation events. In order to compare the magnitude of rotation recorded by several rock units within a field area, it is necessary to reasonably accurately constrain the age of magnetisation and to assess the provenance of an isolated remanence with respect to its origin.
Figure 2.5-1  Core coordinate system established for samples collected using a portable drill, with positive Cartesian directions labelled (A). Core azimuth (α) and hade (β) measured within the core coordinate system (B) [from Butler, 1992].

Figure 2.5-2  Preparation of palaeomagnetic samples from a drill-core sample. Specimens are marked so that when samples are the correct way up, tick marks are always to the left of the fiducial mark. Core coordinates are as indicated in 2.5-1a.

Figure 2.5-3  Field consistency tests of stability of magnetisation. Unheated country rock records both normal and reverse polarity magnetisations, acquired prior to the observed deformation. Intrusion occurred during a period of reverse polarity. Solid (open) circles represent directions in the upper (lower) hemisphere [redrawn from Morris, 2004].
Palaeomagnetic samples are often collected in such a way that the stability of the preserved magnetisation may be geologically tested. Known as field tests of stability, these sampling strategies exploit naturally occurring geological situations in order to indicate whether an isolated remanence was recorded as a primary or a secondary magnetisation. Several of these techniques are illustrated with respect to hypothetical geological situations in Figure 2.5-3 [Morris, 2004], with idealised palaeomagnetic data indicated for each situation.

The Fold (or Tilt) Test

As palaeomagnetic samples are often collected from folded or tilted strata, standard sampling procedure dictates that the bedding attitude, or that of another surface considered to represent the palaeohorizontal at the time of formation, is usually recorded at each sampling site [e.g. Butler, 1992]. Graham (1949) recognised that if rocks were magnetised prior to being folded or tilted, restoration to the palaeohorizontal (through the application of a bedding correction), should bring individual magnetisation directions into closer agreement (i.e. Figure 2.5-3f (in-situ) to 2.5-3c (restored to palaeohorizontal). Where a rock unit is magnetised after the strata has been deformed, through the proximity to an igneous intrusion for example (Figure 2.5-3b), application of the tilt correction will cause individual magnetisation directions to disperse. The Fold Test therefore represents a powerful test of the stability of the isolated remanence and is more generally referred to as the Tilt Test, thereby avoiding the implication that a series of contiguous folds have been sampled.

Although several statistical evaluations of the Fold Test have been suggested [e.g. McFadden & Jones, 1982, extended by McFadden, 1998; Tauxe & Watson, 1994], this study uses the Fold Test suggested by Watson & Enkin (1993) that employs Fisher’s precision parameter and assumes that the total population of
magnetisation directions is most tightly grouped in the orientation in which magnetisation was acquired. Unlike the more inflexible Fold Test of McFadden & Jones (1982), that requires strata to have been deformed as a series of rigid limbs (Figure 2.5-4b), meaning that groups of sites with a particular tilt correction must be used, the Fold Test of Watson & Enkin (1993) allows the use of strata sampled with a 'continuum' folding geometry (Figure 2.5-4c), as is commonly observed in the field. Another reason for choosing the Fold Test of Watson & Enkin (1993) is because an underlying Fisher distribution is assumed, as used in interpreting demagnetisation data during this study.

The Fold Test of Watson & Enkin (1993) is classified in three categories. Where application of the tilt correction clearly improves the grouping of individual site mean directions, a 'positive' fold test is inferred, indicating that a pre-deformation magnetisation has been isolated. A 'failed' or 'negative' fold test infers that the isolated magnetisation post-dates the observed deformation, with application of the tilt correction causing a marked dispersion of the individual site mean directions. In both cases, the change in the Fisher precision parameter, $\kappa$, between in-situ and tilt-corrected coordinates must be statistically significant to produce either a positive or negative classification otherwise the Fold Test is declared 'indeterminate'.

**Caveats concerning the Fold Test**

The Fold Test is most useful where a series of palaeomagnetic sites have undergone differential tilting during identical periods of deformation [Morris, 2004]. Tilt-corrections of small magnitude will have very little effect on the individual site-mean directions, and small improvements in $\kappa$ can be observed on application of small tilt-corrections, even where the magnetisation is known to post-date deformation. As a consequence it can be difficult to determine the true age of
Figure 2.5-4  The Fold Test. A-Original geometry of rock stratum with arrows drawn perpendicular to surface. B-Folding as rigid limbs as required by the test of McFadden & Jones (1981). C-More realistic 'continuum' folding geometry as encountered naturally [from McFadden, 1998].

Figure 2.5-5  Tectonic corrections of A-simply tilted strata and B-strata folded about a non-horizontal axes (plunging fold axis) [from Tarling, 1982].

Figure 2.5-6  Declination anomaly associated with plunge of fold axis. Anomaly is observed to be insignificant for strata dipping <30° [from Tarling, 1982].
deformation where only gently inclined, or homoclinal strata are available for sampling.

The conventional tilt correction as applied in the majority of palaeomagnetic studies, is intended to correct for simple bedding tilt, assuming that it took place about the line of strike of the bedding, with no strain involved (Figure 2.5-5a). This however may not be valid, as suggested by McDonald (1980), who highlighted how rotations about a non-horizontal axis can produce declination anomalies in palaeomagnetic data.

The incomplete application of the traditional tilt correction is particularly evident in the case of plunging folds, where strata is deformed about both the vertical and horizontal axes (Figure 2.5-5b), and large declination anomalies of up to 25° can be incurred through ignoring the plunge of the fold axis for steeply dipping limbs (Figure 2.5-6).

Consistency of magnetisation and the Reversal Test

Often one of the most instructive appraisals concerning the stability of the overall magnetisation of a rock unit is to observe the consistency of the observed magnetisation both over a wide geographical area and through a considerable stratigraphic thickness. If a similar direction of magnetisation, significantly different from that of the present Earth’s field and nearby younger material, is observed throughout the rock unit or is carried by a number of different rock types with differing mineralogy, there is good reason to believe that the overall magnetisation of that rock unit is stable and likely to represent a primary magnetisation. Agreement with known directions of younger age can be a strong indication that a remagnetisation has taken place [McElhinny & McFadden, 1999].
The most definitive consistency test results from the presence of reversals in the polarity of observed magnetisation. If a substantial thickness of strata belonging to a single rock unit is sampled, it is likely that a number of polarity reversals will be encountered. If all of the site-mean directions were acquired within a GAD field with no appreciable continental drift or deformation occurring between polarity reversals to alter the localised magnetisation direction, two groups of directions separated by 180° will be observed (Figure 2.5-3g).

The presence of antiparallel site-mean directions of magnetisation within a single rock unit, indicates that the isolated magnetisation is stable and likely to be primary in origin. However, as shown in Figure 2.5-7, should a secondary component of magnetisation be acquired after the formation of the rock unit, both the normal and reverse polarity total NRM directions will be changed towards the secondary magnetic field direction, with the resultant directions much less than 180° apart.

The acquisition of a secondary component of magnetisation in this situation is of little concern if the secondary component may be completely removed. However this is often not the case and the secondary component may not be satisfactorily separated from the primary component of magnetisation. It should therefore be tested whether or not two antiparallel groups of site-mean magnetisation directions determined from a single rock unit, differ discernibly from being 180° apart [McElhinny & McFadden, 1999].

This study uses the classification proposed by McFadden & McElhinny, (1990), where site-mean directions of both normal and reverse polarity are observed. The test classifies the data, based on the amount of data available by calculating the critical angle (γc) between the mean directions of the two sets of observations at which the null hypothesis of a common mean direction would be rejected at the
Figure 2.5-7  The acquisition of a common secondary (horizontal) component of magnetisation by antiparallel (horizontal) normal and reverse primary components of magnetisation, cause them to deviate by \(<180^\circ\) (after McElhinny & McFadden, 1999).

Figure 2.5-8  The baked contact test. Variation in the direction of magnetisation (arrows), magnetisation (M) and the scatter of direction (\(\sigma\)) with distance from an igneous intrusion in five geological situations [from McElhinny & McFadden, 1999; after Irving, 1964].
95% confidence level, given the observed dispersion in the two sample populations. A positive reversal test is classified as 'A' if $\gamma_c \leq 5^\circ$, 'B' if $5^\circ < \gamma_c \leq 10^\circ$, 'C' if $10^\circ < \gamma_c \leq 20^\circ$ and as 'INDETERMINATE' if $\gamma_c \geq 20^\circ$.

One of the advantages of using the reversal test proposed by McFadden & McElhinny (1990) is that isolated observations of one polarity may still be tested, although tests using fewer observations are less rigorous by definition.

**Conglomerate Test**

Graham (1949) proposed that strata containing conglomerate layers can be used to determine the age of magnetisation carried by a *rock unit*, as long as clasts are of a sufficient size to allow for the collection of orientated samples. Should clasts retain a stable magnetisation relating to their initial formation (i.e. prior to being reworked), the erosion and sedimentary processes responsible for producing the conglomerate will result in the orientation of the magnetic directions carried by all clasts being randomised (Figure 2.5-3d). This suggests that the conglomerate has not undergone remagnetisation since deposition and therefore the magnetisation of the parent formation is likely to have been unaltered since the time of deposition. Should the clast magnetisation directions record the same direction (Figure 2.5-3e), the conglomerate records a secondary remagnetisation that is likely to have also affected the parent formation.

**Baked Contact Test**

The elevated temperature associated with the intrusion of igneous bodies will often serve to remagnetise large volumes of the surrounding country rock, resulting in in a secondary magnetisation parallel to the primary magnetisation of the intrusion. In order to prove that a stable magnetisation is carried by the intrusion, the direction carried by the unbaked country must also be established. Depending on the
stability of the magnetisation recorded by the country rock, those areas unaffected by the intrusion should retain a magnetisation direction different to that of the intrusion (Figure 2.5-8). Depending on the size of the intrusive body and the nature of the outcrop, it may be impractical to successfully sample such a situation in the field. In such cases, a positive baked contact test may be shown if it can be demonstrated that there are changes in properties with distance from the baked contact, coincident with the diminishing effect of the igneous body [Everitt & Clegg, 1962].

2.6 Sample Measurement, Demagnetisation and Rock Magnetic Techniques

The objective of many palaeomagnetic studies is to discern the characteristic remanent magnetisation (ChRM) direction of a sampling unit, from the total NRM recorded by individual specimens at a site level. NRM is a vector sum of all of the components recorded by a sample, and secondary components can be progressively removed through step-wise demagnetisation. The most commonly used techniques are those of thermal and alternating field (AF) demagnetisation. Principle Component Analysis (PCA) of demagnetisation data is then employed [Kirschvink, 1980], in order to determine the ChRM direction of the individual specimen, as well as any other secondary components that may also be recorded.

The bulk magnetic mineralogy of samples is investigated using various rock magnetic techniques. Comparison of bulk rock magnetic data with demagnetisation data can then be made to try to establish which minerals actually contribute towards carrying the observed NRM.

Measurement of Remanence

Most specimens collected proved to be reasonably strongly magnetised, with even the more weakly magnetised sedimentary samples often characterised by initial
NRM intensities of \(-10\ \text{mA/m}\) or greater. The majority of samples were measured using Molspin MS2 fluxgate magnetometers at the University of Plymouth Palaeomagnetic Laboratory, which have a quoted noise level of \(0.025\ \text{mA/ms}\) [www.molspin.com] and operate by rotating an orientated sample within a shielded fluxgate coil detector.

An orientated reference sample of known magnetisation is used to calibrate the amplitude and phase of the reference waveform and the magnetometer is routinely recalibrated approximately every 60 minutes during use, in order to correct for any instrumental drift. The intensity and direction of the magnetisation is measured within the \(yz\), \(xz\), \(-y-z\) and \(-x-z\) planes for each sample, from which the overall sample magnetisation is then calculated in both polar and Cartesian coordinates by an attached computer.

Unlike the majority of the other samples collected, many of the limestone samples collected from the Pabellón Formation, proved to be too weakly magnetised to be measured reliably using the MS2 magnetometers. These samples were measured using the 2-G Enterprises DC SQUID Cryogenic magnetometer, housed at the Oxford University Palaeomagnetic Lab. (capable of measuring samples with very small magnetisations) and has a quoted noise level of \(0.0001\ \text{mA/m}\) for a \(10\ \text{cm}^3\) volume rock sample [www.2genterprises.com].

Demagnetisation Techniques

All rocks generally contain dispersed ferromagnetic grains that contribute to the overall NRM. The coercivity/temperature spectrum, over which the NRM magnetisation is carried, depends on the size distribution and composition of these grains. In order to investigate the characteristic palaeo-direction retained by a rock unit, it is desirable to remove all secondary magnetisations and isolate the primary
(or ChRM) component of remanent magnetisation. The progressive destruction of a sample's NRM can be achieved using a number of techniques that exploit differences in the magnetic properties of the various ferromagnetic grains contributing to the overall magnetisation. In this study Alternating field (AF) and Thermal demagnetisation were routinely used and the only ones further discussed. Where possible, at least one pair of sister specimens per site were demagnetised using both stepwise AF and thermal techniques. This was carried out not only to observe the consistency of magnetisation within individual core samples, but also to ensure that the demagnetisation equipment did not influence the measured component of magnetisation.

A (ferro-) magnetic grain will only realign to an applied magnetic field when the microscopic coercive force applied by the external field exceeds the coercivity of that grain. AF demagnetisation exposes rock specimens to an alternating magnetic field with pre-determined peak amplitude, (B), so that any ferromagnetic grain with a coercivity \( \leq B \) will be aligned parallel to the applied field. The specimen is tumbled about three axes so the alternating field is applied to all possible grain orientations [McElhinny & McFadden, 1999]. The amplitude of the field is then slowly reduced to zero so that ferromagnetic grains with successively lower coercivities become randomly orientated (Figure 2.6-1). The overall magnetisation of all grains with coercivities less than the peak field will therefore cancel, effectively removing the contribution of that grain fraction from the observed magnetisation.

An Agico LDA-3 unit, capable of producing peak fields of 100mT with a three-axis tumbler arrangement, was used for AF demagnetisation. The unit is shielded by several layers of mu-metal, as well as being housed within Helmholtz coils, to
Figure 2.6-1  Schematic ramped AF demagnetising field. Tumbled specimens are exposed to an alternating magnetic field with peak amplitude $B_{MAX}$, which is then progressively ramped to zero (10% decrease in B per ½ cycle is shown as an example only).

Figure 2.6-2  The acquisition of a GRM during AF demagnetisation, identified by either fluctuations of measured magnetic remanence about the true vector of magnetisation (declination indicated-A), or increasing amplitude of GRM component coinciding with increasing $B_{MAX}$-field (B).
prevent the acquisition of an anhysteretic remanent magnetisation (ARM) due to the presence of a direct extraneous field.

AF demagnetisation using tumbling apparatus has been shown in some cases to induce a component of rotational remanent magnetisation (RRM) \cite{Smith & Merrill, 1980; Stephenson, 1980} to be a form a gyroremanent magnetisation (GRM) \cite{Roperch & Taylor, 1986}. To identify GRM acquisition during AF magnetisation, the orientation of the sample was routinely reversed for successive demagnetisation steps. Should a component of GRM be acquired, reversing the orientation of a sample for every step will result in the reversal of the polarity of the acquired GRM and results in the measured direction of magnetisation fluctuating around the 'geological' component of magnetisation (Figure 2.6-2a). The amplitude of GRM should also increase as the peak field increases (Figure 2.6-2b).

The 3-G SQUID magnetometer used at the Oxford University Palaeomagnetic laboratory, features an inline static AF demagnetiser utilising three mutually perpendicular demagnetising coils and batches of eight samples could be AF demagnetised and measured automatically. Although maximum fields of 150mT could be achieved, a maximum field of 100mT was used to avoid the acquisition of a static GRM.

**Thermal Demagnetisation**

A magnetic grain undergoes a transition from stable magnetic behaviour to superparamagnetic behaviour and is therefore free to realign to an external field, at a particular (unblocking) temperature ($T_b$). Depending on the range of magnetic grain sizes and compositions present, the overall magnetisation of a rock sample can be unblocked across a range of temperatures. $T_b$ however, does not
necessarily equal the Curie point \( T_c \) of a carrier mineral [Butler, 1992]. A magnetic remanence may be progressively unblocked through stepwise thermal demagnetisation. If a sample is heated to a set temperature and held at this temperature such that it is evenly heated, any magnetic grains with unblocking temperatures \( T_b \) equal to or less than the maximum oven temperature \( T_{\text{demag}} \) will be unblocked. By progressively increasing the temperature in steps, a larger proportion of the magnetic grains, and hence magnetisation, will be unblocked until completely destroyed.

Two ovens were used for the thermal demagnetisation of samples, a Magnetic Measurements MMTD01 fully programmable oven, capable of automatically demagnetising 12-15 samples in one batch, and a manually operated Pyrox oven, with separate heating and cooling chambers. A maximum number of 30 samples were demagnetised in each batch using the Pyrox oven, although one batch of samples could be cooled while a second was heated. Both ovens are shielded by several mu-metal layers and housed within Helmholtz coils, reducing the ambient magnetic field within both of the ovens to <40 nT, approximating a field-free environment.

Excluding the NRM measurement, thermally demagnetised samples were routinely subjected to 17 heating steps between 100-670°C, with specimens held at the higher temperatures for longer periods to equilibrate. Smaller temperature intervals were used at higher temperatures to provide more demagnetisation steps above 400°C, where the majority of the most commonly encountered magnetic minerals are unblocked. The bulk magnetic susceptibility of samples subjected to thermal demagnetisation was monitored between measurements, in order to observe any chemical changes that may occur during the repeated heating cycles.
In order for thermal demagnetisation data to be compared and analysed with respect to the observed unblocking temperatures, each oven was thermally calibrated to ensure that the correct temperature was actually reached. Using the method of Tiano (Unpublished PhD), five drilled palaeomagnetic specimens with thermocouples cemented in place were used to observe the variation of temperature along the length of the Pyrox oven as shown in Figure 2.6-3, with the results of Tiano (Unpublished PhD) used for the MMTD01 oven for comparison.

The voltage from each thermocouple was for every heating step (Figure 2.6-3) was converted to temperature using standard conversion values for a k-type thermocouple. Drilled samples remained in the same position for each measurement and the voltage was measured every two minutes after the set temperature was reached until the reading stabilised.

Barring the thermocouples placed next to the insulating bungs at either end of the sample boat (Figure 2.6-3), the temperature within the central 40cm of the Pyrox oven oversteps the set temperature by <10°C (Table 2.6-1), with the most noticeable overstep in temperature associated with the oven being initially heated to high temperatures from cold. Tiano (Unpublished PhD) found that the MMTD01 failed to reach or overstepped the set temperature by ~10°C at temperatures below 500°C (depending on position within the oven), and exceeded the set temperature by up to 25°C above 500°C. The temperature at the centre of the oven most closely approximated the set temperature.

A series of fifteen sister (granite) samples were demagnetised using the MMTD01 and Pyrox oven (B and A samples respectively). The variation of normalised specimen intensity against demagnetisation temperature of two sister samples as well as the unblocking spectrum are shown in Figure 2.6-4. Both ovens clearly unblock equivalent components of magnetisation, across similar temperature
Figure 2.6-3  Arrangement of drilled samples containing thermocouples and granite 'spacer' samples used to determine oven temperature during oven calibration. Exploded diagram indicates the ceramic sample boat used and insulating bungs (5cm thick) fixed in place at either end. Not to scale.

<table>
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<th>Temp (°C)</th>
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<th>TC3</th>
<th>TC4</th>
<th>Av.</th>
</tr>
</thead>
<tbody>
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<td>-7.00</td>
<td>-7.00</td>
<td>-7.67</td>
</tr>
<tr>
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<td>5.60</td>
<td>3.87</td>
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<tr>
<td>400-560</td>
<td>7.57</td>
<td>6.00</td>
<td>10.86</td>
<td>8.14</td>
</tr>
<tr>
<td>580-670</td>
<td>8.00</td>
<td>6.00</td>
<td>11.20</td>
<td>8.40</td>
</tr>
</tbody>
</table>

Table 2.6-1  Difference between required and actual oven temperature during oven calibration. Readings from only the central three thermocouples are summarised (see text).

Figure 2.6-4  Comparison of the demagnetisation of sister samples using the Pyrox (A-samples) and MMTD01 (B-samples) ovens.
ranges and are therefore considered to produce directly comparable thermal demagnetisation results. Of the fifteen sister samples tested using both ovens, all produced similar unblocking spectra within acceptable instrument error.

Determination of Sample Magnetic Mineralogy

Ferromagnetic minerals have characteristic coercivities and thermomagnetic properties (Table 2.3-1). The magnetic mineralogy of specimens was investigated for at least one representative sample from each site, using thermomagnetic and IRM experiments. Comparing the bulk magnetic mineralogy of a sample to those magnetic minerals observed to contribute towards sample NRM, can give information on processes that have affected a rock sample, which can in turn help to interpret whether or not an isolated remanence is of primary or secondary origin.

Variation of Low Field Susceptibility vs. Temperature

The measurement of the low field susceptibility of a material identifies the ease with which it may be magnetised. Increasing temperature reduces the relaxation time of a ferromagnetic grain, until the grain begins to behave paramagnetically at its Curie Point ($T_C$) (or Néel Point ($T_N$)), at which point the bulk susceptibility reduces to zero. The Agico Kappabridge KLY3-CS offers a fully automated way of determining the Curie point(s) of any magnetic minerals in a powdered sample. A sample is gradually heated in steps up to a maximum temperature (usually set at 700°C), and then cooled to room temperature, with the low field susceptibility measured continuously.

As a mineral is heated beyond its Curie point, the observed susceptibility decreases, with the Curie temperature corresponding to the inflexion in the heating
curve (Figure 2.6-5). Where more than one magnetic mineral is present, several drops in susceptibility may be observed.

IRM coercivity experiments

The step-wise acquisition of isothermal remanent magnetisation (IRM) is commonly used to identify ferromagnetic minerals based on their characteristic coercivity spectra [Dunlop, 1972]. A magnetic mineral will become more strongly magnetised as the applied field is increased in magnitude, until the mineral reaches saturation, from which point no further increase in magnetisation will be observed.

After AF demagnetisation to 100mT, samples were magnetised along the +ve z-axis in fields up to 800mT, the largest available, and the total magnetic moment was measured after each acquisition step. IRM data (Figure 2.6-6a) is displayed with the measured intensity of magnetisation normalised by the intensity of saturation magnetisation (or maximum intensity should a sample not become saturated).

Where more than one magnetic mineral is present within a sample, the resultant magnetisation will be the sum of the contributions of each. Should one mineral become saturated, the remaining minerals will continue to acquire magnetisation in successively greater fields until they reach saturation, but a change in the slope of the acquisition curve will be observed (Figure 2.6-6b).

Once the maximum IRM was acquired, it was progressively destroyed by applying the same step-wise magnetic fields along the –ve z-axis and the result measured. At some particular field strength (Figures 2.6-6a & b) the residual magnetisation along the +ve z-axis, will be completely countered by the magnetisation acquired along the –ve z-axis, resulting in a net sample magnetisation of zero. This field-
Figure 2.6-5 Variation of (normalised) bulk susceptibility with temperature. Inflexions in the heating curve (black) indicate the presence of at least three magnetic minerals with the Curie temperatures indicated. Thermochemical changes to the sample during heating are obvious in the cooling curve (white) with the lowest temperature mineral obviously absent.

Figure 2.6-6 IRM acquisition (white) and 'Back-field' Coercivity (black) experiments. A-IRM acquisition spectra of a low coercivity mineral assemblage (titanomagnetite) shows steep acquisition curve up to 80 mT & fully saturated by ~300mT. Back-field coercivity of 27.3 mT. B-IRM acquisition spectra of a multi-mineral assemblage, steep acquisition curve up to 80 mT (titanomagnetite), with sample not saturating in max fields (haematite). Back-field coercivity of 258.6 mT indicates dominance of haematite in carrying overall IRM.
strength is referred to as the 'back-field coercivity', and is one characteristic of the overall magnetic mineral assemblage within a sample.

Lower coercivity magnetic carrier minerals such as titanomagnetite, are observed to possess much higher saturation magnetisations than 'higher' coercivity minerals such as haematite (Table 2.3.1). For this reason, contributions from low coercivity minerals may dominate the observed IRM.

Three-axes IRM (Lowrie-type) experiments

A more diagnostic extension of the analysis of IRM acquisition curves was proposed by Lowrie (1990) and involves the magnetisation of different coercivity fractions along three orthogonal directions, using successively smaller fields. The composite magnetisation is then thermally demagnetised, with the contribution from each component of magnetisation recorded separately. First a "hard" component of 800mT was applied along the z-axis, then a second "medium" component of 300mT was applied along the y-axis, and finally a "soft" component applied along the x-axis using a 50mT field.

The "medium" 300mT component is chosen to discriminate between magnetite and pyrrhotite, as pyrrhotite has a maximum coercivity of 500-1000mT and is therefore less likely to become fully saturated in a 300mT field. The "soft" IRM component of 50 mT will identify the presence of any MD magnetite that may be present, which is characterised by a much lower coercivity than PSD and SD magnetite.

The composite magnetisation is thermally demagnetised, with the intensities of the three orthogonal magnetisations recorded at each step and plotted separately (Figure 2.6-7). By comparing the coercivity of the magnetic mineral assemblage, with their unblocking temperatures, it is possible to more effectively identify the
Figure 2.6-7 Thermal demagnetisation of a three component IRM [after Lowrie, 1990]. Squares represent the 'hard' component (800 mT along the z-axis), triangles the 'intermediate' component (300 mT along the y-axis) and circles the 'soft' component (50 mT along the x-axis).
magnetic mineral present within a sample, than through IRM acquisition
experiments alone [Lowrie, 1990].

2.7 Interpretation of Palaeomagnetic Data

Demagnetisation Data

All of the magnetometers used during this study measure three mutually
perpendicular components of the overall magnetic moment of a specimen \((M_x, M_y\)
and \(M_z\) (Am\(^2\))), in reference to the core-coordinate system (Figure 2.5-1). In
palaeomagnetic studies the overall magnetisation of a sample is most simply
described by declination \((D)\), inclination \((I)\) and overall intensity of magnetisation
\((A/m)\).

In core coordinates, the total magnetic moment \((M)\) of a specimen is defined as

\[
M = \sqrt{M_x^2 + M_y^2 + M_z^2}
\]

With intensity of \(NRM\) (or magnetisation) associated with a specimen defined by

\[
NRM = \frac{M}{V}
\]

Where \(V\) is the sample volume (11 cm\(^3\) for all standard core specimens). In core
coordinates declination \((D_c)\) and inclination \((I_c)\) are defined by the simple
trigonometric relationships

\[
D_c = \tan^{-1}\left(\frac{M_y}{M_x}\right)
\]

And
As each core is uniquely orientated in the field, individual specimen demagnetisation data must be transformed from core-coordinates, into a common geographic coordinate system. This is achieved by rotating the measured vector of magnetisation using the azimuth (α-in the direction of drilling) and dip from vertical (hade-β) measured during the recovery of a core sample (Figure 2.5-1).

Display of Demagnetisation Data

The direction of a magnetic vector can be displayed using an equal area stereonet projection (with upper and lower hemispheres identified). The convention used in this thesis indicates points in the upper (lower) hemisphere as open (closed) symbols (Figure 2.7-1a). One shortcoming of the display of specimen demagnetisation data using stereonets is that the relative intensity of each measurement cannot be displayed. It is therefore common practise to accompany a stereonet displaying demagnetisation data with a plot of sample intensity against demagnetisation step the so-called Zijderveld (1967) diagram (Figure 2.7-1b). The declination and inclination components of successive demagnetisation steps can be imagined separately as orthogonal vectors in three-dimensional space starting from a common origin, with the intensity represented by the length of each vector [McElhinny & McFadden, 1999]. By projecting the vector end-points of successive demagnetisation steps onto two orthogonal planes, declination data is plotted in the horizontal plane containing the N-S and E-W axes as solid points, whilst the inclination is plotted in the vertical plane containing the N-S (or E-W) and Up-Down axes as open points. The N-S axis is therefore common to both planes (Figure 2.7-2).
Figure 2.7-1  Example of demagnetisation data displayed on a equal-area stereonet. A- Stereonet plot displaying well clustered demagnetisation data plotted in the upper hemisphere. B-Normalised intensity plot to accompany stereonet.

Figure 2.7-2  2D representation of stepwise demagnetisation data using a Zijderveld diagram. Theoretical NRM comprised of two components. Lower stability component $D = 45^\circ, I = -60^\circ$ (sub-vertical), removed by demagnetisation step 3. Higher stability component $D = 45^\circ, I = -15^\circ$ (sub-horizontal).
Principal Component Analysis (PCA)

Demagnetisation data from individual samples in field-corrected coordinates were interpreted using the PM (DOS) and PMGSC (Windows) software packages of Randy Enkin (www.pgc.nrcan.gc.ca/tectonic/enkin.htm). The software employs the multivariate technique of principal component analysis [PCA-Kirschvink, 1980] to estimate the lines and planes of least-squares fit along the demagnetisation path of a specimen, displayed on a Zijderveld plot. All components of ChRM or other commonly observed components of remanent magnetisation interpreted using PCA are based on at least three consecutive points. Those components of magnetisation common to many or all of the individual specimens were averaged to determine site mean ChRM directions using Fisher (1953) statistics (after combining data from sister specimens where appropriate). Overall mean directions were then calculated for the overall rock unit from the site mean directions, again using Fisher statistics.

Lines & planes of best fit

In the simplest cases, the various components of magnetisation that comprise the NRM of a sample, are carried by magnetic factions with discrete coercivity or thermal unblocking spectra and are easily distinguished as linear segments of the demagnetisation path [Kirschvink, 1980] (Figure 2.7-3a). This is because only one component is destroyed at any time, as the sample is progressively demagnetised. Where successive points on a demagnetisation path form linear segments on a Zijderveld plot, PCA can be used to determine the best-fitting line, with the precision of the fit estimated by the maximum angular deviation (MAD). Only those PCA line-fits with MAD ≤ 10 were considered well defined and used for subsequent analysis. In determining the direction of ChRM, the origin was often considered as a separate point data point and as such the best-fitting line was
Figure 2.7-3 Demagnetisation of three-component NRM with various degrees of overlap of coercivity/blocking temperature spectra. First column indicates amount of overlap between components A, B & C, with numbers indicating the number of components being simultaneously removed. Second column indicates demagnetisation of [A, B] plane (light grey) and [B, C] plane (dark grey), with linear segments indicated in bold. Stereonets in the third column display the difference vector paths as described by Hoffman & Day, (1978) [after Kirschvink, 1980].

Figure 2.7-4 Use of converging remagnetisation samples to determine an unresolved direction.
forced through the origin. This was only undertaken however, if the deviation of the best-fit line from the origin was at worst equal to the RMS deviation from the data points.

Many rocks sampled for palaeomagnetic analysis do not retain a single component of magnetic remanence, with the overall observed NRM often comprising of two or more components. As the amount of overlap between the components of magnetisation increases, the length of the observed linear segments in the demagnetisation path decreases, until at some point, certain components of magnetisation may become hidden. *Hoffman & Day* (1978) constructed difference vector paths, calculating the vector removed sequentially by each demagnetisation step, to show that adjacent component pairs form the planes \([A, B]\) and \([B, C]\), resulting in two great-circle difference vector paths which intersect at the direction of component B (Figures 2.7-3b, c & d).

*Kirschvink* (1980) found that rather than relying on the vector subtraction process of *Hoffman & Day* (1978), which tends to increase the amount of noise, especially for more weakly magnetised samples, neighbouring demagnetisation planes could be used to determine the hidden component, even where no linear segment is evident. The common line defining intersection of planes \([A, B]\) and \([B, C]\) in Figure 2.7-3 is parallel to component B, and can be calculated as the vector cross-product of the respective poles to the these demagnetisation planes. In this way, if a palaeomagnetic specimen has \(k\) components of magnetisation, \(k-2\) of these may be resolved [Kirschvink, 1980].

*Remagnetisation circles*

When two or more components of magnetisation are progressively destroyed through stepwise demagnetisation, it is commonly observed that one of these
components is preferentially removed, with the resulting demagnetisation path observed to follow a great-circle when plotted on a stereonet, but defined by linear segments on Zijderveld plot [McElhinney & McFadden, 1999]. Where the coercivity/unblocking spectra of these components overlap significantly, or where the intensity of magnetisation drops below the sensitivity of the instrument used, the ChRM is often not fully resolved, with the only information concerning the direction of the ChRM residing along the great-circle path. A single great circle will not provide enough information to define the ChRM direction, but where several specimens exhibit such behaviour, an estimate of the ChRM direction may be estimated as the point where these great circles converge (Figure 2.7-4).

In some cases it is possible that a number of samples may display well-defined ChRM vectors, whilst some only provide great circles. It is desirable to combine the two sets of data, and a method of doing this is suggested by McFadden & McElhinny (1988), who use sector constraints on the great circles, based on the direct observations of ChRM. In this way, the overall mean direction becomes biased towards the directly observed data and away from the intersection of the great circles, which provides a less exact estimate of the ChRM direction.

Fisher Statistics

Palaeomagnetic directions are routinely interpreted as vectors of magnetisation, described by three variables, declination and inclination and intensity of magnetisation. The intensity of magnetisation however, is not necessarily related to the reliability of the direction recorded by a specimen. Because of this, observations from individual specimens are given unit weighting, regardless of the intensity of magnetisation, with the ends of the unit vectors intersecting the surface of a unit sphere [McElhinny & McFadden, 1999]. Palaeomagnetic data can
therefore treated as essentially bivariate data described by declination and inclination.

Fisher (1953), proposed a distribution where concentration of points on the surface of a unit sphere about the true direction is proportional to \( \exp (\kappa \cos \psi) \), where \( \kappa \) is the precision (or concentration) parameter and \( \psi \) is the angle between the observed and true direction [McElhinny & McFadden, 1999]. This is referred to as the Fisher Distribution, and is analogous to a bivariate normal distribution.

The probability density of points in a Fisher distribution on the unit sphere is given by

\[
P = \frac{\kappa}{4\pi \sinh \kappa} \exp(\kappa \cos \psi)
\]

where for a Fisher distribution with \( \kappa = 0 \), points will be uniformly distributed over the unit sphere, and will become more tightly clustered about the true direction as \( \kappa \) increases, to infinity if they are all identical to the mean [Tarling, 1982].

The underlying assumption of the Fisher Distribution is that if all of the magnetisation directions recorded by individual specimens within a site, or site mean directions within an overall rock unit, were acquired within a similar palaeomagnetic field, the individual directions must all be drawn from the same parent distribution of magnetisations about the true direction. The true (mean) direction and the degree of scatter about this direction may then be estimated.

Fisher (1953) showed that for \( N \) palaeomagnetic observations drawn from a Fisher distribution with precision parameter \( \kappa \), and a resultant vector of length \( R \), the mean direction will also be Fisher distributed about the true direction, but with a precision of \( \kappa R \). The probability \( (1-P) \) that the true mean of number of
palaeomagnetic observations drawn from a population with $\kappa > 3$, lies within a circular cone of semi-angle $\alpha_{(1-P)}$ about the resultant vector $R$, is given by

$$\cos \alpha_{(1-P)} = 1 - \frac{N}{R} \left( \frac{1}{R} \right)^{N-1} - 1$$

where $P$ is usually 0.05 so that a circle of 95% confidence about the mean is defined. In this way, should a direction fall within the $\alpha_{95}$ error, there is no reason to suspect it was not acquired within the same palaeomagnetic field as recorded by the remaining samples.

Calculation of Rotation

Once magnetised any deformation affecting a rock unit (as well as any continental drift that may have occurred), will also affect the observed direction of magnetisation. As discussed in Chapters Three and Four, studies from the Andean margin and elsewhere document large crustal rotations about vertical axes, by observing the displacement of the direction of remanent magnetisation with respect to reference pole/direction.

Palaeomagnetic reference poles represent the apparent position of a palaeomagnetic pole for a particular continent and period, in present day coordinates (latitude & longitude). A reference pole can therefore be used to predict the expected direction of magnetisation at the site of sampled rock unit, for a discrete geological epoch or time window [e.g. Besse & Courtillot, 2002, 2003]. As the position of a reference pole accounts for the continent's motion, the difference between the observed and expected magnetisation directions at a sampling locality, will describe the gross local tectonic rotation experienced by a rock unit. Traditionally tectonic motion is described in direction space by the magnitude of rotation ($R$) and flattening ($F$) which are given by
\[ R = D_o - D_x \]

\[ F = I_x - I_o \]

Where \( D_o \) & \( I_o \) are the observed declination and inclination and \( D_x \) & \( I_x \) the expected declination and inclination. Consequently +ve rotations are considered to be clockwise with -ve rotations, anticlockwise, while +ve flattening represents a relative poleward transport of the rock unit [Beck, 1980].

Beck (1980) suggested that the discordance of a palaeomagnetic direction should be assessed more rigorously with quantified errors. This combines the statistical confidence limits of the measured palaeomagnetic direction and the expected direction determined from a reference pole.

\[ \Delta R = \sqrt{\left(\Delta D_o^2 + \Delta D_x^2\right)} \]

\[ \Delta F = \sqrt{\left(\Delta I_x^2 + \Delta I_o^2\right)} \]

Using these statistical error limits it is therefore possible to observe the actual degree of concordance between the observed and expected direction. If \( R \) (or \( F \)) exceeds \( \Delta R \) (or \( \Delta F \)), then the direction can be considered statistically discordant.

Demerest (1983) recognised that previous studies had exaggerated the errors in the magnitude of tectonic rotation (\( \Delta R \)) and inclination flattening (\( \Delta F \)) and linked the correction to the number of site mean measurements. As the directions from six or more sites were generally used in this study the standard correction factor of 0.8 is applicable, so that;

\[ \Delta R = 0.8 \times \sqrt{\left(\Delta D_o^2 + \Delta D_x^2\right)} \]
\[ \Delta F = 0.8 \times \sqrt{\Delta I_x^2 + \Delta I_y^2} \]

These approximations are considered to be valid where \( \alpha_{95} < 10^\circ \) and the observed inclination \((I_o)\), < 70\(^\circ\).
Chapter Three

Crustal Deformation and Rotation about a Vertical Axis

Palaeomagnetism provided some of the most compelling evidence in support of the concept of continental drift. The advent of the Geocentric Axial Dipole (GAD) hypothesis meant that the past motion of stable (cratonic), continental interiors could be determined through the construction of Apparent Polar Wander Paths [APWP's-e.g. Creer et al., 1957; Irving & Irving, 1982]. Although continental plate interiors remain relatively undeformed for significant periods of time\(^1\), plate boundaries suffer quite intense periods of deformation, as stresses associated with the relative motion between adjoining (mobile) plates, are accommodated.

The active deformation of a continental plate boundary at a convergent margin (such as western margin of South America), is usually accommodated within an actively deforming zone, whilst the continental interior remains largely unaffected. The crust of the deforming zone is typically uplifted and thickened through crustal shortening [e.g. Isacks, 1988], but this process is commonly accompanied by the horizontal displacement, or in-situ rotation of (upper) crustal 'blocks', as a response to motion along the margin, either within the crust or (upper) mantle. The rotation of a crustal block within a deforming margin can be described as a translation with respect to the stable continental interior. An APW path therefore provides a reference frame within which the deformation of a plate boundary (in the terms of crustal rotation) may be quantified.

\(^1\) In plate tectonic theory, continental cratons are assumed to be rigid and to undergo only very slow rates of deformation. Satellite geodetic/GPS measurements now indicate that the interiors of the great plates such as South America are stable to less than 2mm a\(^1\) [e.g. Leffler et al., 1997].
In many areas of active deformation, the magnitude and sense of observed crustal rotation has been related to the overall structural setting of the affected block. Rotations in these areas are interpreted to accommodate, at least a part of the relative motion between two converging plates. This strongly suggests that crustal rotations are a direct consequence of first-order tectonic processes, such as subduction or continental collision. Ancient crustal rotations could therefore provide valuable insight into past tectonic processes, assuming tectonic features responsible for accommodating rotation are preserved.

A variety of mechanisms have been proposed to explain how crustal rotation has been (or is) accommodated within deforming margins, such as western South America. This chapter explores the relationship between continental deformation and crustal rotation, as well as how crustal rotation may be driven and accommodated within deformational margins. The existing palaeomagnetic database from the Central Andes is introduced, as well as some of the models proposed to explain the observed pattern of crustal rotations, which include a number of mechanisms active at the largest (orogen-wide) scale, to smaller-scale, localised processes affecting small crustal blocks. Some of these rotation mechanisms will be discussed using examples of palaeomagnetic (and structural geology/geodetic) studies from several areas from around the world, where either active deformation is, or previous deformation was responsible for producing crustal rotation.
3.1 What are Palaeomagnetically Defined Rotations?

The magnitude of crustal rotation, \( R \), recorded by a crustal block\(^2\) in relation to the stable cratonic interior of a continent, is simply calculated as the difference between the declination of the observed mean palaeomagnetic direction, \( D_0 \) and the expected declination at the sampling latitude, \( D_x \) (Section 2.7). Where the calculated rotation is greater in magnitude than the associated error, \( \Delta R \), the observed palaeomagnetic direction is considered statistically significant (discordant) to the expected direction [Beck, 1980].

The choice of a suitable reference pole should account for the effects of plate motion, hence the amount of rotation is assumed to be a record of the gross tectonic motion experienced by a crustal block, with respect to the plate. The simplest interpretation of a palaeomagnetically determined rotation is that it occurred about a vertical axis, with the pole of rotation either situated within (or very near to) a crustal block, or about a Euler pole of rotation some distance from the crustal block (Figure 3.1-1).

Where tectonic rotation occurs about a localised, vertical axis (Figure 3.1-1a), the affected crustal block will undergo no appreciable latitudinal movement. Therefore only the observed declination of the remanent magnetisation vector will be discordant with the expected direction and the crustal block remains at the same palaeolatitude. In such a situation, the observed rotation is usually attributable to localised deformation.

\(^2\) For the sake of discussion during the remainder of this thesis, the term 'crustal block' or 'block' will be used to describe a body (of any scale) that has been rotated or otherwise translated with respect to the stable cratonic interior of a continent.
Figure 3.1-1  Tectonic rotations of a crustal block creating discordance in the preserved palaeomagnetic direction compared to the palaeomagnetic pole (PP). A-Rotation about a vertical axis situated within the crustal block. Block is rotated by angle R, with no associated poleward transport, therefore the observed inclination should be equal to the expected inclination. B-Rotation about a Euler pole situated some distance from the crustal block. Block is rotated by the angle Q, accompanied by a poleward translation by the angle p, therefore the observed inclination of the translated block will be shallower than the expected inclination at the palaeomeridian (inclination flattening). C-Cross-section of a dipolar field indicating expected inclination (grey arrows). A crustal block initially magnetised at a low palaeomeridian (black arrows) exhibits inclination shallowing, F, after (northerly) poleward translation [from Butler, 1992].
Where a crustal block is rotated about an external Euler pole, the block will undergo a translation. In the case shown in Figure 3.1-1b, the crustal block is translated from shallow equatorial latitudes, to a more northerly latitude (towards the palaeomagnetic pole, PP). The overall translation is comprised of a rotation, $R$, normally accompanied by a significant change in latitude, $p$, which will produce an inclination anomaly between the observed and expected direction of magnetisation.

Figure 3.1-1c shows a cross section of the dipolar field within which the crustal block illustrated in Figure 3.1-1b was magnetised and subsequently translated. As the block is transported northwards, the expected inclination at higher latitudes (grey arrows) becomes progressively steeper than the observed inclination (black arrows). At its final position, the observed inclination is shallower than the expected inclination by an amount, $F$, referred to as incline flattening. The magnitude of inclination flattening is assessed in the same manner as crustal rotation and is only considered significant if it is larger than the associated error \cite{Beck1980, Demerest1983}.

The actual accommodation of rotation within an active margin can be accomplished in a number of ways and at a range of scales. Often the manner in which rotation is accommodated is dependant on the deformational style of the active margin and the continuity of palaeomagnetic data can be used to identify rotational domains within a wider margin.

The following section discusses how continental deformation may be described most accurately and how this relates to the manner in which crustal rotation is accommodated within active deformational margins.
3.2 The Deformation of Continental Lithosphere

3.2.1 Discrete vs. continuum deformation

There has been much debate in recent literature as to how the deformation of continental lithosphere may be most accurately described. First order observations from convergent margins around the world indicate that the relative motion between two (rigid) colliding plates is accommodated within a deforming zone (Figure 3.2-1), the width of which will be dependant on the thickness of the crust involved, and/or the type of faulting which accommodates the deformation [e.g. McKenzie & Jackson, 1983; Gordon, 1995].

Plate tectonic theory successfully describes the movement of oceanic lithosphere, where thin and essentially rigid plates are deformed within belts, a few kilometres in width. England & McKenzie, (1982, corrected 1983) modelled the deformation of the lithosphere based on the properties of a thin viscous sheet, thereby approximating the deformation of oceanic crust. A map of global seismicity however (Figure 3.2-2), indicates that deformation at convergent margins involving continental lithosphere, is commonly distributed across much wider zones, characterised by domains of faults often hundreds to thousands of km in width [Scotti et al., 1991; Gordon, 1995]. Deformation of the continents therefore departs significantly from that predicted by traditional plate tectonic theory, and new descriptions of the deformation of continental lithosphere have been proposed.

Two kinematic models have been used to describe the nature of active faulting within a deforming zone, and represent the extreme end-members of an overall deformation spectrum affecting continental lithosphere at convergent margins. This spectrum ranges from the localised deformation along a single fault or small
Figure 3.2-1  A deforming zone of width $a$, between two plates. In this example Plate 2 moves with velocity $U_2$ ($V_2$) in the $x$ ($y$) direction (Plate 1 fixed) so the deforming zone is in extension [from McKenzie and Jackson, 1983].

Figure 3.2-2  Map of shallow global seismicity. Oceanic transform boundaries are often delineated by narrow bands of concentrated seismic activity, in contrast to areas of continental deformation that appear to generally form much wider zones of more diffuse seismicity [Figure 5.2 from Keaney & Vine, 1996].
number of faults, referred to as discrete or microplate deformation, to the near-continuous distribution of deformation across a large number of closely spaced faults, known as continuum deformation [e.g. Thatcher, 1995-Figure 3.2-3].

Discrete or microplate deformation

Discrete or microplate deformation describes how displacement and strain is concentrated along a small number of major faults, with very little deformation observed within the intervening crustal blocks (Figure 3.2-3a). This description/approach assumes that the size of the crustal blocks involved is equivalent to the width of the deforming zone and that the plates to either side are rigid and the deforming block itself behaves in a brittle manner. As such, discrete deformation is effectively analogous with traditional plate tectonics, which essentially describes block motion at the largest scale. This apparent corollary with plate tectonics was assumed to prove unequivocally that continental deformation is undertaken through the movement of a small number of large, fault-bounded blocks.

The discrete deformation model is very effective at describing the deformation of oceanic lithosphere, where oceanic plate boundaries are clearly defined by very narrow bands of seismicity associated with ridge transform boundaries for example (Figure 3.2-2), as well as for describing deformation associated with very narrow continental shear zones such as the San Andreas fault system in southern California (which forms the eastern boundary of the 'Sierran' microplate-Figure 3.2-4), or the Alpine Fault in New Zealand. The same generally cannot be said for regions of continental extension or shortening, where active plate boundaries are often characterised by broad zones of diffuse seismicity and consequently the discrete deformation approximation breaks down (Figure 3.2-2).
Figure 3.2-3 Discrete and continuum deformation of continental lithosphere. A-widely spaced faults delineating large-scale, rigid and brittle blocks across which displacement and strain is localised. B-as A but displacement and strain is accommodated across numerous smaller strands between major faults, resulting in a more diffuse or even distribution of observed deformation [from Thatcher, 1995].

Figure 3.2-4 Oblique mercator projection of the California and Sierra Nevada region, illustrating the discrete nature of the San Andreas Fault, which accommodates ~75% of Pacific-North American convergence, with ~25% transferred to the continental interior by the Eastern California Shear Zone (ECSZ), a large part of which is accommodated by diffuse deformation within the Walker Lane Belt [Figure 1 from Unruh et al., 2003].
Continuum or distributed deformation

Where the deformation of continental lithosphere is accommodated within diffuse zones of seismicity, several hundreds of kilometres in width, rather than within narrow zones associated with discrete (single) fault systems, the strain associated with a deformational event, must somehow be distributed throughout the deformation zone. The continuum or distributed deformation model describes how such strains are more evenly distributed, with relative motion accommodated by displacement along many closely spaced and simultaneously active faults within the deforming zone (Figure 3.2-3b). Implicit to this kinematic model is the assumption that the width of the intervening crustal blocks is very small in relation to the width of the overall zone affected by deformation. Strain can therefore be considered as varying smoothly throughout the deforming region and continental deformation can be thought of as a quasi-continuous flow [e.g. McKenzie & Jackson, 1986, Thatcher, 1995].

3.2-2 Fault displacement and discrete/continuum deformation

In the seismogenic regions near active faults, crustal movements are often dominated by the cyclical build-up and release of seismic energy, where the interseismic strain accumulated prior to fault rupture, is balanced exactly by the coseismic strain release during fault rupture, as first proposed by Reid (1910) [Figure 3.2-5-Thatcher, 1995]. Although geodetic measurements of motion during the interseismic period of strain accumulation would show that motion is distributed quasi-continuously, the net effect of the entire seismic cycle would be represented by the displacement of a rigid block across a fault [Thatcher, 1995].

The overall distribution of motion across a deforming zone depends on a number of factors, most notably the number and spacing of faults, as well as their type, dip
Figure 3.2-5 Displacement across a fault during interseismic and coseismic periods. Strain is accumulated elastically during the interseismic period, approximating true continuum deformation. During fault rupturing, elastic strain is released through brittle deformation and total interseismic and coseismic motion is that of a rigid block displacement across the fault.

Figure 3.2-6 The relationship between fault angle and seismogenic zone width. Low-angle faults have wider seismogenic zones across which cyclic straining is accommodated. As faults become more closely spaced and seismogenic zones overlap, crustal deformation appears to be quasi-continuous.
and slip rate. Also of importance is the width of the seismogenic fault zone, which is greater for lower angle faults (Figure 3.2-6). If the spacing of the faults is much greater than the zone of cyclic straining, the net effect over a complete seismic cycle can be determined and the deformation can be described in terms of the motion of discrete blocks. With regard to the palaeomagnetic study of continental margins, the term ‘block rotation’ is often used to describe the motion of an aseismic block within a deforming zone. In the case of the converse situation, where faults are more closely spaced and the respective seismogenic zones overlap, the offset across individual faults will be indistinguishable, with the consequence that the displacement across an entire deformation zone will appear to be quasi-continuous [Thatcher, 1995] (Figure 3.2-6).

3.2-3 Application of the discrete & continuum deformation models

Both styles of deformation have been argued to apply to actively deforming margins and the prominent distinction between them appears to be the scale at which structural observations are made. Brittle deformation is clearly an important feature of the Earth’s crust and strain discontinuities in the form of discrete faults are common. The continuum model of deformation simply states that it is the net effect of motion of all faults (as well as rotation of the intervening crustal blocks), which accommodates relative plate motion [Lamb, 1994]. Truly continuum deformation would be therefore represented by a completely plastic flow, such as observed in the formation of mylonites.

The completely fluid-like deformation of over-thickened cold continental lithosphere however is likely to be unrealistic and many orogenic zones are characterised by the presence of large aseismic blocks (e.g. the Alpine-Himalayan belt or Sierra Nevada as shown in Figure 3.2.4). Discrete and continuum deformation models therefore represent idealised end-members and many areas
are probably most correctly characterised by a combination of the two models, as observed in southern California for example [as discussed by Scotti et al., 1991], where a wide deformation zone is divided into fault domains whose boundaries are major through going fault-systems.

Because of the lack of distinction between truly continuum and discrete-type deformation, the actual kinematics of the in situ rotation of small-scale crustal blocks has been the cause of much debate. Nelson & Jones (1987) illustrated the differences between the two end-members and the rotation pattern that might be expected for each type by considering the accommodation of shear deformation (Figure 3.2-7). In the undeformed state (Figure 3.2-7a), a block is subjected to right lateral shear from one side, with the resulting rotation illustrated by a number of passive makers (arrows).

In the first example shear is accommodated throughout the block without rotation, through the formation of strike-slip faults parallel to the 'master fault' (Figure 3.2-7b). This explanation has suggested to explain the lack of rotation along the North Anatolian Fault Zone, where deformation may have been distributed across a number of elongate fault blocks that lie parallel to the trace of the main fault, therefore preventing rotation [Platzman et al., 1994].

In the second example the block is deformed through pervasive continuum deformation (i.e. by crystal-plastic flow distributed throughout the block for example-Figure 3.2-7c). Although deformation is distributed throughout the block, the passive markers closest to the master fault are rotated by the greatest magnitude, with rotation observed to decline with distance away from the boundary fault. One of the most cited examples of continuum deformation concerns palaeomagnetically determined rotations from the Pacific Northwest continental margin of the USA where the magnitude of rotation recorded by a suite
Figure 3.2-7  General models for accommodating strike-slip deformation with arrows representing passive markers of rotation (i.e. palaeomagnetic directions). A- Undeformed block. B-Deformation accommodated on parallel faults with no resultant rotation. C-Pervasive continuum deformation where the amount of rotation increases towards the 'master fault'. D-In situ rotation of rigid blocks driven by displacement along block boundary faults. The magnitude of rotation is consistent within each block as well as within the whole rotated domain. E-Small block model. The magnitude of rotation generally increases towards the 'master fault'. [Figure 6-Nelson & Jones, 1987]
of Miocene (15-12 Ma) basaltic flows, decreases smoothly with increasing distance from the continental margin [England & Wells, 1991]. This assertion was founded on the assumption that the lithosphere contains such a large number of inhomogeneities and discontinuities that no individual feature has a significant effect on the overall rotation pattern at the scale of the overall margin and therefore the continental lithosphere behaves as a thin layer of fluid [e.g. Bird & Piper, 1980; England & McKenzie, 1982, corrected 1983; McKenzie & Jackson, 1983].

In contrast to 'fluid-like' continuum deformation, shear can also be accommodated within strike-slip zones through the rotation of rigid blocks, accommodated by displacement along block boundary faults within the rotation domain, through discrete deformation (Figure 3.2-7d). The length of the blocks in such a situation will be approximate to the width of the deforming zone itself. Not only is rotation consistent within each block, but it also remains constant within the overall domain of similarly orientated block boundary faults. The implication of this model is that reasonably large regional domains of homogenous rotation can be produced, such as the models have been widely produced to explain the rotation pattern in California [e.g. Jackson & Molnar, 1990; Luyendyk, 1991; Nicholson et al., 1986]. If block boundary faults should form two domains however, the sense of rotation will be opposite in each domain [e.g. Ron et al., 1984].

The idealised models of continuum and discrete deformation producing crustal rotations shown in Figure 3.2-7, are end-members of the deformation spectrum and true deformation is likely to be a mixture of both. Nelson & Jones (1987) refer to this as the 'small block model', in which the crust is regarded as being broken up into small blocks (much smaller than the deforming zone) and 'float' in a continuously deforming substratum. The overall rotation pattern from such an
arrangement is more complex than produced by idealised continuum or discrete deformation, with blocks of various shape observed to rotate at different rates and directions [e.g. Lamb, 1987]. Generally however the largest rotations are observed towards the boundary fault (Figure 3.2-7e).

Some models concerning the actual accommodation of crustal rotation are discussed in the following section with respect to examples proposed to explain the pattern of crustal rotations from both the Central Andes. Initially the gross pattern of rotation determined from studies throughout the Central Andes will be described with respect to the choice of reference poles used in this study.

3.3 Pattern of Crustal Rotation Identified in the Central Andes

One of the most conspicuous features of the central Andes, is the marked change in strike of the gross geological features, including mountain chain, trench and coastline that occurs at ~19°S (Figure 1.2.1). The bend in the western coastline of South America (convex to the east) is referred to as the Arica Deflection and represents an approximately 60° change in strike, with the overall trend of the margin deflected from approximately NNE-SSW in the south (northern Chile and Argentina), to NW-SE to the north (Peru & Bolivia).

An increasing number of palaeomagnetic studies have reported that much of the Andean margin has been affected by crustal rotations. The dominant pattern of rotation is well known, with ACW (CW) rotations to the north (south) of the Arica deflection and this appears to reflect the gross trend of the entire margin, leading to early models suggesting the entire orogen had undergone rigid bending [e.g. Carey, 1958; Kono et al., 1985]. As the number of palaeomagnetic studies increased, such models failed to describe the emerging complexities in the overall pattern of rotation and therefore many authors suggested models describing the
localised rotation of small-scale crustal blocks about a vertical axis, often associated with strike-slip displacement along marginal parallel fault systems such as the Atacama Fault [e.g. Forsythe & Chisholm, 1994]. Whilst such models are effective at describing localised perturbations in the overall rotation pattern, the observation of large areas of homogenous rotation that extend beyond the influence of these fault systems, suggests that purely localised rotation models cannot account for the gross regional deformation pattern. Several authors have therefore proposed that the pattern of rotations observed throughout the Central Andes is more effectively described by the rotation of very large crustal blocks (or domains) [e.g. Abels & Bischoff, 1999].

This chapter introduces the Central Andean palaeomagnetic database as a whole and defines the wider context of this study, with respect to the geographic distribution of palaeomagnetic data, as well as describing the various models that have been proposed to explain the overall rotation pattern. As described in the preceding chapters, palaeomagnetic data from the Central Andes has primarily been used to directly observe the (gross) magnitude of crustal rotation about a vertical axis and hence investigate the pattern of rotation, both along and across the Andean margin. The overall central Andean palaeomagnetic dataset has been compiled from published palaeomagnetic studies, with some of the data also drawn from conference abstracts. Where existing data has been re-worked or combined with later studies, the most recent source is used. Prior to a discussion of palaeomagnetic data from the Central Andes however, the choice of the reference poles used during this study to determine the magnitude of crustal rotation is discussed.
3.3-1 Palaeomagnetic Reference Poles for Stable South America

A persistent problem associated with palaeomagnetic studies in the Andean margin, concerns the lack of consensus over suitable reference poles for use in the analysis of South American tectonics. This is an unfortunate result of the lack of a wide age range of material within the stable South American craton with Tertiary aged poles particularly poorly constrained (Beck, 1999, 2004).

A number of authors have countered the lack of data by compiling reference poles using both South American palaeomagnetic data from the stable continental interior and African data transformed into South American coordinates [e.g. Roperch & Carlier, 1992; Randell, 1998 (Table 3.3-1); Somoza, 2001 (quoted in Somoza & Tomlinson, 2002-Table 3.3-2)]. Lamb & Randell (2001) also used inclination data from regions that have undergone rotation about a vertical axis to expand the South American palaeomagnetic dataset and therefore produce reference poles of higher quality (Table 3.3-3).

An alternative approach involves the reconstruction of global plate motion (i.e. a global rotation model) to create a global ‘Master’ APWP. By defining a plate circuit within the global rotation model, palaeopoles sampled from different plates, both continental and oceanic, are combined to determine an individual APWP for a particular continental craton such as South America [e.g. Besse & Courtillot, 2002; Schettino & Scotese, 2005]. In this way, Besse & Courtillot (2002, corrected 2003) compiled a Master APWP for South America, calculated using a 20Ma, sliding window every 10Ma (Table 3.3-4).

3.3-2 Comparison of South American palaeopoles

Although various authors suggest that the South American plate has undergone some north-south motion during the past 200Ma [e.g. Beck, 2004; Schettino &
<table>
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<th>Expected Direction (*)</th>
</tr>
</thead>
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Table 3.3-1 Palaeomagnetic reference poles calculated for South America calculated by Randall (1998). Expected direction calculated for a location at the centre of the La Guardia study area (27.78°S, 69.77°W).

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Table 3.3-2 Palaeomagnetic reference poles calculated for South America calculated by Somoza (2001). Expected direction calculated for a location at the centre of the La Guardia study area (27.78°S, 69.77°W).

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Table 3.3-3 Palaeomagnetic reference poles calculated for South America calculated by Lamb & Randall (2001). Expected direction calculated for a location at the centre of the La Guardia study area (27.78°S, 69.77°W).
Table 3.3-4 Master APWP for South America spanning the past 200Ma calculated by Besse & Courtillot (2002, corrected 2003). Expected direction calculated for a location at the centre of the La Guardia study area (27.78°S, 69.77°W).

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Figure 3.3-1 Expected direction of the palaeofield for a locality at the centre of the La Guardia study area, calculated using the reference poles from Tables 3.3-1 to 3.3-4.
Scotese, 2005], there is very little palaeomagnetic evidence for this, and overall it is unlikely to have exceeded ~200km [Beck, 2004]. Consequently, the motion of the stable South American plate has been predominantly east-west since the early Mesozoic, corresponding to the opening of the South Atlantic ocean. Purely latitudinal motion is undetectable through palaeomagnetic analysis and hence reference poles from all the published apparent polar wander paths (APWPs) for South America during this period, are coincident with the present-day geographic axis. This is illustrated by a comparison of the reference directions calculated for a location at the centre of the La Guardia field area, using the reference poles of Randall (1998), Somoza (2001), Lamb & Randall (2001) and Besse & Courtillot (2002, corrected 2003) shown in Figure 3.3-1.

In order to make a genuine comparison of the published South American APWPs, crustal rotations were recalculated for the entire Central Andean palaeomagnetic database using several sets of reference poles, the results of which are displayed in Figure 3.3-2. There are no major differences noted in the overall pattern of crustal rotations in the Central Andes, calculated using those reference poles compiled from South American (and African) palaeomagnetic data [Randall, 1998; Lamb & Randall, 2001; Somoza, 2001], or constructed using a global dataset [e.g. Besse & Courtillot, 2002, corrected 2003], at least for the past 100Ma. Differences do however become more pronounced in early Cretaceous-Jurassic reference directions, with the oldest rotations observed to differ slightly depending on the choice of palaeopole.

The most obvious disparity between the four sets of reference palaeopoles concerns the magnitude of calculated inclination flattening. Whilst little variation is noted along the margin for any of the individual Mesozoic (green), Paleocene (blue) and Neogene (red) palaeopoles, those of Randall (1998), Lamb & Randall
Figure 3.3-2  The overall Central Andean Rotation Pattern recalculated from published data using reference palaeopoles calculated by A-Randall (1998), B-Lamb & Randall (2001), C-Somoza (2001) and D-Besse & Courtillot (2002, corrected 2003). Both the magnitude of rotation and inclination flattening (& error) are plotted against latitude. Green line represents best fitting polynomial curve to Mesozoic data (unfilled circles), blue line-Paleogene data (unfilled triangles) and red line-Neogene data (filled diamonds).
(2001) and Somoza (2001), exhibit consistently pronounced flattening of up to 15-20° on average, which is most obvious for Paleocene aged data. The reference poles calculated by Besse & Courtillot (2002, corrected 2003) exhibit the least amount of inclination flattening, with similar amounts displayed by data of Mesozoic, Paleocene and Neogene age (Figure 3.3-2).

3.3-3 Sources of error in palaeopole construction

The discrepancy in the observed magnitude of rotation (as well as inclination flattening) calculated using the different reference poles probably results from a number of sources of potential error;

1. The quality of palaeomagnetic data used in pole construction can vary quite considerably (e.g. PSV may not be fully averaged, inclination flattening may have occurred through mechanical processes) and many authors have proposed strict selection criteria that they suggest should be used in order to minimise errors in the reference datasets used, such as the seven criteria proposed by Van der Voo (1990). Whilst fulfilling many of these criteria does not prove a palaeopole to be an accurate representation of the ancient magnetic field, it does increase the confidence in such an assumption. When applied in the strictest manner however, these selection criteria are often unreasonably restrictive when considering very small reference datasets.

2. When considering palaeomagnetic data that requires transformation into different (e.g. African to South American) coordinates, the transformation is undertaken using a plate reconstruction model which itself will inevitably suffer some error arising from overlaps/mismatches in the actual fitting process.
3. Differences between reference palaeopoles also result as a function of the time window each individual pole spans. For example, the reference poles determined by Lamb & Randall (2001) average time periods of 10-40Ma, whilst those constructed by Besse & Courtillot (2002, corrected 2003) average periods of 20Ma.

3.3-4 Choice of reference palaeopoles

The reference palaeopoles derived from the South American Master APWP of Besse & Courtillot (2002, corrected 2003) were selected to be used during this study, based mainly on the fact that they produce a much smaller spread in the magnitude of observed inclination flattening in comparison to the other reference palaeopoles used (Figure 3.3-2). The minimal inclination flattening is preferred due to the assumption that South America has undergone no detectible longitudinal motion, at least since the early Mesozoic.

3.4 Geographical extent of the Central Andean Rotation Pattern

The CARP is dominated by anticlockwise crustal rotations in central and southern Peru and Bolivia [e.g. Roperch & Carlier, 1992; Rousse et al., 2002], with clockwise rotations observed in northern Chile and north-western Argentina [e.g. Riley et al., 1993; Randall et al., 2001; Arriagada et al., 2003], as shown in Figures 3.3-2 & 3.4-1. The change in the sense of rotation occurs at ~19°S, corresponding to the overall change in trend of the entire margin referred to as the Arica Deflection (Figure 1.2-1). Although not shown in Figure 3.4-1, clockwise rotations are also observed along the Ecuadorian margin [e.g. Roperch et al., 1987], as well as along the northern Peruvian Andes to the north of the Huancabamba bend at ~5°S [e.g. Mourier et al., 1988; Mitouard et al., 1992], continuing the apparent
Figure 3.4-1 The existing Andean palaeomagnetic dataset prior to this study. Crustal rotations are separated into Mesozoic (green), Paleogene (blue) and Neogene (red) age groups, based on the age of the magnetisation stated by the original authors. The palaeomagnetic data is summarised in Appendix A, which includes the references to the original sources. Background DEM constructed from ETOPO 3 topographic data obtained from the USGS.
parallel between palaeomagnetically determined crustal rotations and the strike of the coastline of western South America.

The sheer amount of palaeomagnetic data now published from the Central Andes means that a thorough description of the overall pattern of crustal rotations is unfeasible. As this study specifically concerns palaeomagnetic data from the present day forearc of northern Chile, data from this region is discussed with regard extent and accommodation of crustal rotation.

The geographic distribution of crustal rotations within the Central Andes is displayed graphically in Figure 3.4-1 and was calculated by comparing the observed directions with the expected direction calculated from the Master South American APWP, constructed by Besse & Courtillot (2002, corrected 2003). As will be discussed in the following chapters, palaeomagnetic from the studies of Palmer et al., (1980a) and Riley et al., (1993), have been reworked during this study, taking into account recent changes to the sampled stratigraphy. The reworked data is also included in Figure 3.4-1, and in total, the complete dataset will be referred to as the Central Andean Rotation Pattern or CARP during the remainder of this thesis, as discussed in Chapter One.

3.5 Mechanisms for Accommodating Continental Crustal Rotation

A number of mechanisms have been proposed to explain patterns of palaeomagnetically determined crustal rotation in convergent margins around the world. Generally these models range from the whole-scale rotation of an entire orogen, to the in-situ rotation of small-scale crustal blocks, with the regional rotation of large-scale crustal blocks comprising a third general type of rotation model.
3.5.1 Models of large (orogen)-scale rotation

Many of the most prominent orogenic belts around the world, appear bent or otherwise curved in plan view [Figure 3.5-1-Marshak, 1988]. Several attempts have been made to explain the genesis of curved orogenic margins, through either the rotation of very large-scale (and rigid) limbs, or the 'bending' of an entire margin in response to convergence (often invoking indenter type models). Some of these models simply invoke mechanisms that explain the orientation of 'first-order' tectonic features (such as topographic expression), with little regard to the overall geological setting and are subsequently reworked in an attempt to rationalise these features with realistic geological processes.

Oroclinal bending

One of the earliest theories concerning the large-scale deformation of continental margins, that has survived the subsequent development of plate tectonic theory, was proposed by Carey (1958), who suggested that the arcuate shape of many continental deformation zones, could be explained through the whole-scale bending of the entire margin in response to plate convergence. Carey (1958) proposed that two categories of bent margin exist, based on their tectonic development. Primary bends were suggested to be orogenic margins that initiated with their present curved form-possibly through the fusion of younger orogens onto the irregular margin of a pre-existing craton. Secondary bends or 'oroclines' are suggested to result from the deformation of an initially straight margin.

Marshak (1988), expanded the classification of curved orogens by proposing the term 'non-rotational arc' to describe arcuate margins that developed without the rotation of rocks. Displacement along the length of the bend occurs in such a way that there are no changes to the curvature of the margin. The curved outline of the
Figure 3.5-1 Examples of orogenic bends defined by along-strike variation of structural grain. A-Alpine bends. B-Appalachian bends. C-Southern Asian bends. [Figure 1-Marshak, 1988].
margin, as a whole or in segments, translates into the foreland, where deformation is accommodated by vertical displacement, without altering the strike of the pervasive structural grain along the margin.

It is of great geological interest to determine whether the observed curvature of an orogen represents a primary (inherited) feature, or the result of 'oroclinal' bending. Palaeomagnetic data can provide a suitable tool to investigate the origin of curved margins. Assuming the cover rocks sampled are autochthonous in nature, the declination of remanent magnetisation may be compared to a reference direction calculated from a suitable pole for the stable continental interior. As no rotation is associated with the development of primary or non-rotational bends, the direction of remanent magnetisation should remain unrotated along the length of the margin, whilst rotations with magnitudes reflecting the strike of the margin should be observed along the length of a secondary orocline.

The orocline theory has been tested palaeomagnetically by authors such as Eldredge et al., (1985), with the assumption that the secondary bending of an orogenic margin (in plan view), will produce vertical axis rotation of rocks within the margin. This will produce a progressive variation in the observed declination of remanent magnetisation around a bend or bends, as long as the remanent magnetisation pre-dates the period of deformation responsible for the bending of the orogen. Eldredge et al., (1985), found that many apparently 'secondary oroclines' displayed much smaller magnitude rotations than would be expected from the apparent angular offset of the orocline limbs, suggesting that such bends are likely to have resulted from the enhancement of pre-existing bend.
Differentiai shortening

A major concern with the orocline concept as applied to the Andes by Carey (1958), is the necessary development of extensional voids at the northern and southern limits of an orocline in order to accommodate the rotation of large-scale rigid limbs [e.g. Beck, 1988; Randall et al., 1996] (Figure 3.5-2a). Isacks (1988) effectively reworked the orocline model of Carey (1958), to suggest that the Arica Deflection developed through the enhancement of a pre-existing bend in the Andean margin through differential shortening along the length of the margin (Figure 3.5-2b). Rather than creating voids at the northern and southern limits therefore, the orocline is proposed to have developed through the bending of a narrow belt within the forearc, with displacement behind the forearc accommodated by differential shortening, crustal thickening and uplift.

Estimates of the magnitude of crustal shortening accommodated along the margin indicate that ≥320km were accommodated at the bend (Schmitz, 1994), much greater than elsewhere to the north and south. As the greatest amount of shortening is accommodated at the bend itself, so the greatest amount of crustal thickening and uplift is observed in this area, which Isacks (1988) suggested was responsible for the creation of the Altiplano-Puna Plateau. The maximum width of the Andean mountain Chain is observed at the latitude of the Arica deflection and Isacks (1988) therefore inferred that crustal rotation was associated with the main Andean Orogenic phase (Quechuan Orogeny) during the Miocene-Oligocene (post-25Ma). Isacks (1988) however makes no attempt to describe just how shortening was accommodated either within or behind the forearc and does not discuss the effects relating to the geology along the Peru-Chile margin.

In his model, Isacks (1988) suggests that differential shortening could account for a ~10° clockwise rotation of the forearc in Northern Chile and this was supported
Figure 3.5-2 Cartoons of large-scale mechanisms proposed to explain the observed rotations about the Arica Deflection. A-Oroclinal bending of an initially straight margin [Carey, 1958], B-differential shortening in the interior of the margin [Isacks, 1988], C-oblique convergence at a pre-existing bend leading to localised but systematic rotations on either side of the deflection driven by differential shear [Beck, 1987, 1988] (Figure 7 from Randall et al, 1996).
by several reviews of the palaeomagnetic database from the Central Andes [e.g. MacFadden et al., 1995; Butler et al., 1995]. Many contemporary studies however indicate that crustal rotations recorded by pre-Tertiary aged rocks from the Coastal Cordillera-Precordillera region are much greater in magnitude than this predicted amount, suggesting that the differential shortening model is invalidated [e.g. Forsythe & Chisholm, 1994; Somoza, 1994]. Subsequent studies in the Northern Chilean forearc have also documented that much larger clockwise rotations are recorded and therefore this mechanism cannot explain the bulk rotation pattern, but do not rule out that differential shortening may have added a small component of rotation to an overall 'compound' rotation.

Kley, (1999) investigated the relationship between the magnitude of crustal rotation and estimates of crustal shortening and crustal thickening (approximated by crustal cross-sectional area) in an attempt to rationalise crustal rotations with Andean orogenesis. Crustal shortening and thickening were used to predict the expected amount of crustal rotation (see Kley, 1999 for details) but found that the three datasets did not agree closely, with often contradicting senses of rotations observed (Figure 3.5-3). Although the best fitting polynomial fitted to the Neogene rotations approximates the preferred rotation model, pre-Neogene rotations far exceed that expected, particularly in Northern Chile and north-west Argentina (i.e. south of 19°S), suggesting that the gross pattern of crustal rotation may not be directly associated with crustal shortening and thickening.

Several problems present themselves when estimating the amount of shortening from balanced cross-sections however, most notably the fact that out of plane displacement of material cannot be accounted for. Crustal rotation would inevitably lead to some transport of material along the margin and therefore balanced cross-sections would fail to recognise this.
The Central Andean Rotation Pattern (CARP)

Latitude (°S)

-75 -60 -45 -30 -15 0 15 30 45 60 75

Observed Rotation (°)

5 10 15 20 25 30 35

Figure 3.5-3 Predictions of the magnitude of crustal rotation affecting the Central Andean margin using balanced cross-sections and variations of crustal area. [modified from Kley, 1999]. Palaeomagnetically determined rotations are divided by age into Mesozoic, Paleogene and Neogene, with the best-fitting polynomial for the Neogene dataset indicated for comparison.
Distributed crustal shear

Beck (1987, 1988), proposed that crustal rotation in the Central Andes was driven by margin parallel shear driven by oblique subduction, a theory developed from the explanation of widespread clockwise rotations in the western Cordillera of North America [Beck, 1976] (Figure 3.5-2c). Also referred to as the 'ball bearing model', it was originally intended to illustrate the sense of rotation affecting small structurally controlled blocks, resulting from margin-parallel shear, but the process is suggested to have operated at the scale of the entire margin. This model when applied to the Andes assumes that the Arica Deflection is a Primary Feature of the Andean margin, leading right-lateral shear, leading to clockwise rotation in the forearc to the south of Arica and left-lateral shear, leading to anticlockwise rotation to the north of Arica (Figure 3.5-2c). Whilst this model explains clockwise crustal rotations in northern Chile, geological reconnaissance indicates that the major margin parallel fault systems in this region are left-lateral systems [e.g. Arévalo, 1994, 1995; Brown et al., 1993; Tomlinson et al., 1993], opposite to that required by the model proposed by Beck (1987, 1988) and therefore this model cannot be responsible for accommodating clockwise rotations.

3.5.2 Models of in situ vertical axis rotations of small-scale crustal blocks

Many palaeomagnetically determined crustal rotations determined from the forearc of Northern Chile, greatly exceed the ~10° of clockwise rotation predicted by the model proposed by Isacks, (1980). Several rotation mechanisms have therefore been proposed as alternatives to the large-scale bending/rotation models discussed above, often involving the in situ rotation of small crustal blocks within shear zones. Such models suggest that the crust deforms in a brittle manner (i.e. discrete deformation), defining blocks that rotate about a vertical axis in response
to faulting at the block boundaries. Hence such models are often referred to as vertical axis rotation models.

*Strike-slip tectonic domains*

The presence of several margin parallel fault systems within the northern Chilean forearc, has led to the proposal of a number of kinematically very similar vertical axis rotation models that involve the rotation of a number of blocks between two system-boundary faults. These 'discrete-type' models take the form of the rotating slat models proposed by *McKenzie & Jackson*, (1983, 1986-Figure 3.5-4) and similar 'bookshelf-type' models proposed by *Mandl*, (1987).

*Forsythe & Chisholm*, (1994) suggest that the ~30° clockwise rotations recorded by Jurassic to Cretaceous aged plutons in the Coastal Cordillera of Northern Chile, can be accounted for by displacement along the Atacama Fault System and a series of secondary faults, that trend obliquely to the AFS (Figure 3.5-5). These faults display similar (NW) orientations, characteristic spacing and displacement, (as displayed by the plutons they dissect-Figure 3.5-5). Restoration of the displacement along these faults is suggested to effectively remove the observed palaeomagnetic discordance (as well as restoring the displaced pluton Figure 3.5-5) and therefore suggests that clockwise rotations were accommodated by displacement along the AFS.

The AFS has a well-established history of predominantly sinistral displacement. A common assumption of many rotation models has been that net sinistral (dextral) shear produces anticlockwise (clockwise) rotations, but the sense of shear need not necessarily determine the sense of rotation [e.g. *Mandl*, 1987] (Figure 3.5-6). What appears to be more important in determining the sense of rotation is the sense of motion on the block boundary faults, the initial orientation of these faults
Figure 3.5-4  Pinned slat model proposed in Figure 1 of McKenzie & Jackson, (1986). The width of the deforming zone is a. The motion of plate 1 relative to plate 2 is given by the large white arrow, with the slip vector between the adjacent blocks shown by the large black arrow, involving both normal and left-lateral strike-slip motion. The blocks rotate in a clockwise direction.

Figure 3.5-5  Structural constraints on rotations in the Coastal Cordillera of northern Chile. A- Mapped displacements of granitoid intrusions, B-analogue model of present day situation, which is then restored in C. Thick black arrow indicates average palaeomagnetic direction recorded by intrusions and corresponds to expected direction on restoration of the faults, illustrating the consistency of these data with structural constraints [Figure 14 from Forsythe & Chisholm, 1994].
Figure 3.5-6 A-Anticlockwise rotation generated in a sinistral shear zone with dextral block boundary faults, with blocks pinned to the outside of the deformation zone. B- Clockwise rotations generated by sinistral shear on discrete system bounding faults with sinistral block bounding faults. Blocks therefore pinned to inside of system boundary faults. Opposite senses of rotation can therefore be generated by apparently identical faults (at least in orientation). Figure from Randall (Unpublished PhD).

Figure 3.5-7 Vertical axis crustal rotation models proposed to explain crustal rotations in the Coastal Cordillera of Northern Chile. Model A proposed by Forsythe & Chisholm (1994)-rotations accommodated within the Atacama Fault System (sinistral shear zone), through sinistral displacement along block boundary faults. Model B proposed by Randall et al., 1996-rotations accommodated between subduction trench and Central Valley Shear Zone (CC-PC FS) [Figure 8-Randall et al., 1996].

Model A

Model B

AFZ

Neocomian basins

Compression

Margin parallel slip

Fault bounded blocks within AFS

Central Valley Shear Zone

AFZ

Faults terminate away from main fault

Amount of rotation

σ1

σ3
with respect to the system boundary faults and how the edges of the blocks themselves are attached to the system boundary faults (3.5-6). In the example illustrated in Figure 3.5-6a, the block boundary faults attach to the outside of the deforming zone (i.e. the system boundary faults act as the margin of a wide shear zone) and therefore the individual blocks undergo anticlockwise rotation. If however the system boundary faults are discrete faults, the rotating blocks are pinned to the inside of the deforming zone producing sinistral displacement along the block boundaries, accommodating clockwise rotation (Figure 3.5-6b).

Further study of the Coastal Cordillera of Northern Chile by Randall et al., (1996) has shown that substantial clockwise rotations (of a similar magnitude to that recorded by Forsythe & Chisholm, 1994) are also recorded to the east (and therefore away from the effects of) the AFS. This suggests that displacement along the AFS cannot account for rotation in these areas and therefore is not likely to have been responsible for accommodating regional rotation in the northern Chilean forearc. Randall et al., (1996), proposed a similar model to that proposed by Forsythe & Chisholm (1994), except that the eastern boundary of the deforming zone was comprised by the Coastal Cordillera-Precordillera Fault System (Figure 3.5-7).

Other than the difference in system boundary faults, the two models are kinematically identical, with subsidiary block boundary faults sharing the same NW orientation and undergoing sinistral displacement. Palaeomagnetic sampling to the east of the CC-PC Fault System indicates that crustal rotations of a similar magnitude are recorded in the Precordillera region of Northern Chile, albeit with more pronounced variability, suggesting that like the AFS, the CC-PC FS does not control the regional rotation pattern [Randall et al., 2001].
This highlights a major flaw with models of *in situ* vertical axis rotations. Whilst they can very effectively describe localised rotation pattern, it is difficult to extend the effects of such mechanism over large areas, without the presence of suitable boundary structures. In addition, the rotation of a large number of small blocks might not be expected to produce a regionally consistent rotation pattern.

*In-situ rotation of fold and thrust sheets*

Crustal rotation has also been demonstrated to be associated with thrust (and detached fold) sheets as they propagate through a deforming zone [e.g. Allerton, 1998; Hartley et al., 1992; MacDonald, 1980; Onderdonk, 2005; Vickery & Lamb, 1995]. If the amount of slip varies along the thrust plane, or if the fault becomes pinned at one end for some reason; a thrust sheet will undergo a rotation. This mechanism has been suggested to operate in Northern Chile, where palaeomagnetically defined crustal rotations are associated with thrust sheets operating in a thin-skinned deformational setting [Hartley et al., 1992].

3.5.3 Models of *in-situ* vertical axis rotation of large-scale crustal blocks

Several rotation models have been proposed that explain the pattern of crustal rotations observed from the Central Andes as resulting from plate-scale processes inferred to affect the entire thickness of the continental lithosphere, but that do not operate at the scale of the overall orogen.

*Large-scale strike-slip domains*

The problems created by trying to explain the regionally consistent clockwise rotations observed in the southern central Andes, using only small-scale (localised) crustal rotation models mean that it is appealing to expand these processes, to a much larger (plate-) scale. Abels & Bischoff (1999) extend the the
idea of crustal rotation in the southern central Andes being accommodated by strike-slip large domains, suggesting that homogenous crustal rotations may be accommodated through displacement along a series of trans-continental (NW-oriented) structures (Figure 3.5-8a). This is effectively a scaled up version of the more localised domino-type mechanisms proposed by Forsythe & Chisholm, (1994) and Randall et al., (2001), which Abels & Bischoff (1999) suggest occurred during a period of transpression.

Noting that the AFS and CC-PC FS do not appear to control crustal rotation, Abels & Bischoff (1999) suggest that the Domeyko Fault System (DFS), situated to the east of the the CC-PC FS in the Precordillera, might represent the eastern boundary of the observed crustal rotations, or that the eastern limit of the rotating zone might be delimited by a system of thrust faults to the east of the DFS (Figure 3.5-8b). Crustal rotations are observed to exhibit far greater variability around the DFS, than observed to the west. If the observed rotations are associated with the DFS, the timing of rotation would necessarily be associated with the Incaic Orogeny, during which the DFS operated (mid. to upper Eocene ~43-33Ma). If rotation however pre-dates displacement along the DFS, the localised effects of the DFS would overprint the earlier regional rotation during the Incaic Orogeny.

Oblique convergence & subduction rollback

Hartley et al., (1988), suggest that crustal rotation in the forearc of Northern Chile (22-24°S) was related to an extensional tectonic regime associated with a combination of both subduction roll back and slab pull force. The direction of extension is proposed to have been oblique to the absolute motion of the South American plate, resulting in the rotation of discrete fault-bounded blocks (Figure 3.5-9). Whilst the Chilean forearc is believed to have been in extension for a long period of time during the late Triassic-early Cretaceous, with the margin
Figure 3.5-8  A-Large (plate) scale domino-type rotation mechanism proposed by Abels & Bischoff (1999) to explain the regionally consistent clockwise crustal rotations observed from the southern central Andes. Crustal rotation is accommodated by displacement along NW orientated transcrustal faults (as identified by Salfity, 1985). B-The eastern boundary is formed either by the DFS (in which case the rotations are Incaic in origin) or comprised of a series of thrust faults situated to the east of the DFS. To compare the relative scale of this rotation mechanism and proposed by Forsythe & Chisholm, (1994), whereas this mechanism operates at a plate scale, the effects of Forsythe & Chisholm's model, (1994) are limited to the Coastal Cordillera in the vicinity of the Atacama Fault zone.
Figure 3.5-9 Schematic diagram of the Andean forearc for Northern Chile, showing the effects and processes resulting from the extension of the South American plate at an oblique angle to its direction of absolute motion [Figure XX from Hartley et al., 1988].
interpreted to have been in retreat for much of this time, this period pre-dates the inferred age of many of the rotations determined from palaeomagnetic studies [Randall, 1996]. In addition, it is difficult to resolve the crustal rotations observed further inland with this model.

3.6 How are Crustal Rotations Driven?

Although quasi-continuum models are now believed by many to represent a valid description of the deformation of continental lithosphere, estimates of lithospheric viscosities made in conjunction with typical deformation rates for active mountain belts, imply substantial stresses over depth ranges of several tens of kilometres. This is much greater than the seismogenic depth range of 15-20 km considered, on the basis of seismicity, to deform in a purely brittle fashion [King et al., 1994].

This reasoning has also been cited to infer that the actual pattern of deformation is not determined by the brittle crust, but rather that the brittle crust is passively deformed in response to displacements or rotations in the deeper underlying material [e.g. Jackson & MacKenzie, 1988].

Using the present-day situation in California and Nevada, King et al., (1994), suggest that 60% of relative plate motion is accommodated through active seismicity, thus implying that 40% of the deformation occurs aseismically as creep along faults. This creep can be shown by instrumental measurements to be localised on very discrete surface faults, such as observed for the creeping section of San Andreas fault (Figure 3.6-1a), although other faults such as the Calaveras fault, display much more diffuse surficial deformation spread over several kilometres (Figure 3.6-1b), which in time is interpreted to accumulate and produce drag folding.
Figure 3.6-1  A-Distribution of deformation as a function of depth for block rotations. B-The effect of surface drag folding can be to cause near-surface deformation to be spread over a zone near to faults even though deformation is localised at depth. C-The model proposed by McKenzie & Jackson, (1983), to explain palaeomagnetic rotations. The major faults play a minor role. In each case the unshaded part represents the seismogenic crust with a nominal thickness of 15km [Figure 3 from King et al., 1994]
The models of King et al., (1994) differ to the interpretation of palaeomagnetic data from McKenzie & Jackson, (1983, 1986), who suggest that surface blocks rotate passively as a response to uniform flow in the lower crust & mantle (Figure 3.6-1c). Instead King et al., (1994) suggest the reverse scenario, with superficial ductile deformation processes diffusing block motion at depth (Figure 3.6-1b).

Several authors now contend that the continuous, ductile deformation of the lower crust and upper mantle is of vital importance in the understanding of how deformation is partitioned at the Earths' surface [e.g. Molnar, 1992; England & Houseman, 1986; Sonder & England, 1986]. As opposed to the more traditional view of the lower crust representing a weak layer that serves to decouple the upper mantle from the overlying brittle crust, it is now regarded to be strong enough to couple continuous deformation within the upper mantle to the brittle deformation of the upper crust and therefore to the relative motion of crustal blocks [Molnar, 1992].

Tikoff et al., (2002) describe the upper brittle crust/lower ductile crust, lower crust/lithospheric mantle and upper/lower mantle as possible 'clutch' zones (Figure 3.6-2). In contrast to detachment zones, these sub-horizontal shear zones are interpreted to represent partial attachment zones between layers, with deformation of the upper crust resulting from the bulk flow of the lithospheric mantle. Tikoff et al., (2002) also infer from the pervasive sub-horizontal shearing that deformation within orogenic margins is most likely bottom driven, as opposed to resulting from the horizontal movement of side-driven crustal blocks.

3.7 Summary

A range of potential rotation mechanisms have been suggested to explain the rotation pattern observed throughout the Central Andes with the differences
A-An example of differing response of the rheological layers in the lithosphere, for a transpressional deformation, with a clutch zone in the lower crust. B-Lithospheric strength profile (differential stress vs. depth) derived from extrapolation of experimental results. Clutch zones operate where transitions in rheology occur [Figure 1 from Tikoff et al., 2002].
between the models largely reflecting the scale at which palaeomagnetically determined rotations are described. At the scale of the overall orogen, oroclinal bending/differential shortening correctly explains the sense of rotation observed to the north and south of the Arica Deflection, but fails to predict the complexity of the rotation pattern resulting from deformation occurring at both regional and localised scales. Conversely, models of small block rotation are often well constrained with regard to the structural complexity of small areas but often cannot be extended to explain the large domains of homogenous rotation that extend beyond the effect of the system bounding faults. In the next three chapters the palaeomagnetic data from this study is presented, with the implications of the new data concerning the accumulation of crustal rotation in the Northern Chilean forearc region discussed in Chapter Seven.
Chapter Four

The Coastal Cordillera- Precordillera zone of northern Chile
29-30°S; The Tres Cruces & Condoriaco Areas

4.1. Introduction

This chapter deals first with a revision of existing data from the Condoriaco area and the reporting of a new data set from the Tres Cruces area (areas A and B respectively on Figure 4.1-1b).

The Condoriaco area lies to the east of La Serena, the capital of Chilean Province Region IV, between 29°30' & 30°00'S and has recently been remapped by the SERNAGEOMIN at a scale of 1:100,000 [Empan & Pineda, 1999]. Palaeomagnetic studies of the Cretaceous to earliest Tertiary strata by Palmer et al., (1980a) in the Rio Elqui Valley and wider Condoriaco-Rivadavia area were aimed at investigating the Cretaceous long normal polarity superchron. The stratigraphy used by Palmer et al., (1980a) during sampling has subsequently been modified and the new stratigraphy has been used in a reassessment of the original palaeomagnetic data.

The Tres Cruces field area is located approximately 50km to the northeast of La Serena (area B-Figure 4.1-1a) and is covered by the 1:250,000 scale, “Vallenar y parte norte de La Serena” mapsheet [Moscoso et al., 1980, currently undergoing revision by the SERNAGEOMIN].

Both areas straddle the Coastal Cordillera - Precordillera boundary, although the division between these two physiographic zones is much more cryptic, in some senses, than elsewhere in the Chilean Margin. This is largely due to the absence
Figure 4.1-1  A) Location of the Condoriaco (box A) and Tres Cruces (box B) field areas with respect to major population centres and topographic features (>3000m in grey). B) Outline map illustrating the principle tectonomorphic features and access routes (in red) through the Condoriaco and Tres Cruces field areas. Redrawn from Emparan & Pineda (1999), Pineda & Emparan, (1999) and Moscoso et al., (1982).
of the Central Valley between 28-30°S, meaning that there is no simple physiographic break corresponding to a geological break. As is the case elsewhere along the margin, these domains or physiographic zones are built upon a series of Jurassic to Eocene aged volcano-magmatic arcs which young progressively to the east (Figure 4.1-2).

Topographically the region encompasses a low altitude mountain belt in the west, the coast parallel Coastal Cordillera, where typical peak elevations are generally <1500m, whilst the eastern margin of the field area approaches the more elevated topography of the Chilean Precordillera (~2000-3500m typically). Palmer et al. (1980a) collected the majority of their samples from along Rio Elqui Valley and its subsidiary quebradas. During fieldwork the area was revisited for 5 days of field reconnaissance to establish the locations of the original sample sites and to determine whether there was sufficient material for a further collection.

The Trés Cruces area is divided by the main Pan Americana (Ruta 5) highway, allowing rapid travel both north and south. The majority of samples were collected along the E-W striking valley Quebrada Los Chores (Figure 4.1-1B), due mainly to the limited access afforded by the steep sided and narrow network of adjoining quebradas. The 1:250,000 scale map of Moscoso et al. (1980) includes sufficient detail to target specific magmatic intrusions which coupled with modern, reliable radiometric ages for the mapped plutons made these a prime target for investigation. These intrude a range of, in the main, volcanic- volcaniclastic units of Early Cretaceous age and these were extensively sampled (Figure 4.1-3).

The Coastal Cordillera - Precordillera boundary in this region is poorly defined, as elsewhere in northern and central Chile these two areas are separated by the Central Valley (or Central Depression) – a relatively modern, low lying, perched forearc basin, whose width is such that the older stratigraphy changes significantly
Figure 4.1-2 Geology of the Condoriaco and Tres Cruces field areas. Redrawn from Emparan & Pineda (1999), Pineda & Emparan, (1999) and Moscoso et al., (1982). Palaeomagnetic sampling localities indicated from this study (circles) and Palmer et al., (1980a-squares).

This study; 1) Arqueros Formation, 2) Quebrada de Los Choros dyke swarm, 3) Pluton Tilgo, 4) Pluton Los Colorados, 5) Pluton Corredores and 6) Pluton Las Campañas.

Palmer et al., (1980a); 1) Arqueros & Quebrada Marquesa Formations, 2) Quebrada La Totora Strata & Vinita Formation and 3) Los Elquinos Formation.
Figure 4.1-3  Generalised stratigraphic log of geology between 26-30°S in Coastal Cordillera-Precordillera region of northern Chile. The equivalent but more specific stratigraphy determined for the Condoriaco and Tres Cruces field areas as determined by Emparan & Pineda (1999) and Pineda & Emparan, (1999) is also shown (without repetition of the periods of volcanism).

Figure 4.1-4  Stratigraphic relationships proposed to explain the contact between the Arqueros and Quebrada Marquesa Formations in the Condoriaco area. A)-Structural relationship [Emparan & Pineda, 1997]. Quebrada Marquesa Formation fills half graben basin formed by normal movement by the La Liga listric fault.. B)-Stratigraphical relationship [Mourgues, 2000a, b & c; Perez & Reyes, 2000]. Quebrada Marquesa Formation is laterally equivalent to Arqueros Formation.
from one side to the other. For the sake of description here the line of the Paleocene magmatic arc in the north, which in effect becomes a caldera field in the south, is used as the boundary. In part this coincides with the southern extension of the Chañarcillo fold and thrust belt (La Liga Fault to the south) and is often marked by a distinct change in topography.

4.1.1 Geology of the Coastal Cordillera

The stratigraphic interpretation of the region is based on a combination of the recently published 1:100,000 scale maps for the Très Cruces area in the south [Emparan & Pineda, 1999, 2000] and the more detailed stratigraphic relationships established for similar, along strike, units observed further to the north in the Copiapó region (~27-28°S) [e.g. Marschik & Fontbote, 2001; Arévalo, 1994]. This is further refined by recent palaeontological and biostratigraphic analysis of the Cretaceous strata in the Condoriaco and Très Cruces areas [e.g. Perez & Reyes, 2000; Mourgues, 2000a, b & c], which provides a more precise age determination of the strata sampled for palaeomagnetic analysis along Quebrada Los Choros and the basis for the re-examination of the data of Palmer et al., (1980a).

Arc–backarc deposition and volcanism

The oldest (E. Jurassic) plutons intrude Paleozoic basement and/or Triassic sediments exposed at the coast (Figure 4.1-2). The majority of the area however comprises a thick sequence of (Jurassic? to) Early Cretaceous and earliest Tertiary aged volcanics, volcanioclastics and sediments, which record the easterly progression of the active volcanic/magmatic arc/back-arc. These units are intruded by granitic plutons ranging in age from approx 155-100 Ma. The principal stratigraphic units (Figures 4.1-2 and 4.1-3) are briefly described below. The original provenance of the name and the ages for the units are given in brackets.
**Bandurrias Group** (L. Valanginian-Barremian [Moscoso et al., 1982; after Segerstrom, 1960])

A continental volcanic/volcaniclastic sequence including intercalations of shallow marine limestones and sandstones, initially termed the Bandurrias Formation in the Copiapó area with an overall thickness of 2500-3000m. The general lack of (age-diagnostic) fossils in the predominantly volcanic strata, mean that there are difficulties in producing a definitive age for the Bandurrias Group. A laterally equivalent relationship with the predominantly marine sediments of the Chañarcillo Group, east of Copiapó [Segerstrom & Parker 1959], indicates a minimum age span of Late Valanginian to Barremian for the Bandurrias Group and it may include strata as young as Aptian-Albian in age. Between Vallenar (~28°S) and 29.5°S the Bandurrias Group’s stratigraphy is poorly constrained – south of 29.5°S it equates to the following two units:

**Arqueros Formation** (Hauterivian-Barremian [Aguirre & Egert, 1965])

An approximately 1,250m sequence of alternating volcanic and marine sedimentary beds with the presence of small ‘manto’-type (stratabound) manganese deposits in the upper part. The base of the Arqueros Formation is not documented, but the upper member was originally described as passing directly into the overlying Quebrada Marquesa Formation.

**Quebrada Marquesa Formation** (Aptian-Albian [Aguirre & Egert, 1970]) An approximately 1,900m sequence of clastic sedimentary rocks, predominantly of continental origin as well as andesitic lavas. The middle of the formation contains manto-type manganese deposits, which have been extensively mined in the area, with fossiliferous lenses of calcareous sandstones observed near the base.
However, *Pineda & Emparan* (1997) identified an erosional unconformity between the Arqueros and overlying Quebrada Marquesa Fms., in contradiction to the conformable relationship noted above by *Aguirre & Egert* (1965, 1970). Instead, they recognize a syn-extensional setting for the Quebrada Marquesa Fm. such that it partially fills an extensional half-graben formed by the east side down displacement of the Arqueros Fm. by the large La Liga (normal) fault (Figure 4.1.4a).

*Pineda & Emparan* (1997) state that the age of this extensional event must be Hauterivian, based on the fossil content of the Quebrada Marquesa Fm., implying therefore that the Arqueros Fm. is Hauterivian or older. On the most recent stratigraphic evidence this age is contradicted by the bio-stratigraphic analysis of the Arqueros and Quebrada Marquesa Formations in the Condoriaco area [*Mourgues* 2000a; *Perez & Reyes* 2000]. Palaeontological evidence suggests that the lowest marine limestone unit of the Arqueros Fm. and the calcareous marine intercalation at the base of the Quebrada Marquesa Fm. contain similar faunal assemblages of Hauterivian age. This suggests that the deposition of the Quebrada Marquesa Fm. was in fact contemporaneous with that of the Arqueros Fm., with *Mourgues* (2000a) proposing that the lower units interfinger with each other, with the Hauterivian marine horizon common to both (Figure 4.1-4b). The same authors also concluded that the upper marine limestone unit of the Arqueros Fm. contained abundant examples of *Agriopeura aff. blumenbachi*, a fossil indicating an upper Barremian age, comparable with examples from the Pabellón Formation (upper unit of the Chañarcillo Group). In short, the Bandurrias Gp, the Chañarcillo Gp, the Arqueros and Quebrada Marquesa Formations are, in essence, lateral equivalents of each other spanning (at least) the Hauterivian to Barremian period or approximately 135-125Ma and perhaps as young as Aptian (~115Ma) in places.
In Quebrada de Los Choros *Mourgues* (2000b & c) recognised an approximately 140m sequence of alternating limestones and volcaniclastics with volcanics increasing up sequence, (sampled for palaeomagnetic analysis in this study). Based on lithological and faunal similarities this package was directly correlated to part of the Arqueros Formation in the Condoriaco area. At both localities, distinctive porphyritic andesites with submarine pillow-type structures underlie the lowest calcareous clastic rocks. These 'pillow' lavas are significantly different to any stratigraphically higher lavas leading *Mourgues* (2000b) to conclude that the Arqueros Formation and the calcareous clastic sequence of Quebrada Los Choros were direct equivalents. In this study the interpretations of *Mourgues* (2000a, b & c) and Perez & Reyes (2000), concerning the age and extent of the Arqueros and Quebrada Marquesa Formations is adopted and hence the need for revision of the original Palmer et al. data.

**Arc magmatism**

North of 29°15' S, Jurassic plutons intrude Palaeozoic metamorphics and latest Triassic-earliest Jurassic Formations of the coastal region. Dioritic intrusions crop out along the coast for approximately 55 km. Currently this older plutonic suite is poorly known and was therefore not sampled during this study.

The Cretaceous Magmatic Arc is expressed as a NNE striking, margin parallel batholith, extending the full length of the region (28-30°S) [*Moscoso et al.*, 1982] and beyond. Contacts between separate plutonic bodies are observed and these have frequently been or are being mined as they are often the sites of major mineralisation. *Emparan & Pineda* (2000) recognised two plutonic events at 130-122 Ma and 115-110 Ma from the radiometric dating of the plutons, but were unable to discriminate these events petrologically or geographically. The eastern flank of the batholith is marked by a slightly younger suite of granitic-dioritic stocks.
(100-93 Ma, including Pluton Los Colorados) which intrude the Arqueros and Quebrada Marquesa Formations.

**Principal Structures**

The principal structures in the Condoriaco and Tres Cruces areas represent the continuation of the extensive, margin parallel fault systems recognised to the north. The most westerly of these structures is the El Romeral Fault System, which offsets the early-mid Cretaceous coastal batholith and represents the southern extension of the Atacama Fault. Displacement along this fault is predominantly of Jurassic-early Cretaceous in age, although there is evidence of later reactivation [Pineda & Emparan, 1999] and the fault system is associated with a number of economically significant iron mines (El Romeral (active), El Tofo (abandoned)). The fault zone (as mapped) undergoes a subtle change in strike at the latitude of Quebrada de Los Choros, from ~N-S to NNE-SSW and divides said quebrada from the coast to the west.

Whilst the El Romeral Fault System lies to the west of the majority of the sampling localities in the Condoriaco and Tres Cruces areas, the eastern boundary coincides with the Chañarcillo Fold and Thrust Belt (CFTB), that essentially separates the Coastal Cordillera-Precordillera to the north, but is observed to taper out at 29°30'S. Displacement along the CFTB has been variously ascribed to late Cretaceous [77 Ma-Arévalo, 1994; Arévalo & Grocott, 1997] or Paleocene [66 Ma-Matthews et al., 2001] deformation, with the younger age more closely associated with the resurgence of active arc magmatism/volcanism during the Paleocene in this area. Although the CFTB dies out southwards, the extensional La Liga Fault is observed to continue in its position c.30°S. This fault represents a significant break in topography and approximates the southern extension of the CC-PC boundary.
4.1.2 Geology of the Precordillera

The eastern sector of the field area is dominated by the plutons of the latest Cretaceous-Paleocene and Eocene arcs intruded into late Cretaceous-early Tertiary strata, which where seen, rest unconformably on the early Cretaceous in the west. The principal stratigraphic units of interest (Figures 4.1-3 & 4.1-4) are briefly described as follows;

*Intra-arc deposition & volcanism*

**Cerrillos Formation** (Albian-Cenomanian [Segerstrom & Parker, 1959]) (≡ Vinita Formation ([Aguirre & Egert, 1965]) ≡ Quebrada La Totora Strata ([Emparan & Pineda, 1999])

The Cerrillos Formation represents the initiation of continental deposition with principally andesitic volcanism. It marks the termination of the marine depositional environments of the lower Cretaceous and has been recognised as cropping out without interruption from Copiapo (c. 27°00') to 29°30'S and essentially without any significant changes in basic lithology [Segerstrom & Parker, 1959]. Aguirre & Egert (1965), described the equivalent Vinita Formation to outcrop south of 29°30'S with similar lithologies throughout, except for a middle member containing calcilutites and limestones with intercalations of andesite. Moscoso et al., (1982), state that it is impossible to define a regional boundary between the two formations due to their very similar nature. As such they treated the Vinita Formation as being part of the same unit as the Cerrillos Formation. Both formations rest, supposedly, unconformably on the Chañarcillo and Bandurrias Groups and are suggested to be post-Aptian in age.

At the base of the Vinita Formation, Emparan & Pineda (1999) noted the presence of a distinct conglomeratic unit comprising clasts derived from the
underlying Arqueros Formation (Figure 4.1-4). This distinctive unit, termed the Quebrada La Totora strata, is not recognised in other locations.

**Los Elquinos Formation** [Aguirre & Egert, 1965] ≡ Quebrada Yungay Strata


These units comprise lavas, tuffs and breccias of basaltic to rhyolitic composition, indicating intense bi-modal volcanic activity in a continental depositional environment. The Los Elquinos Formation (including the Quebrada Yungay Strata) are recognised south of 29°30'S, (Figure 4.1.3). They rest unconformably upon the Vinita Formation with the top of the Elquinos formation representing the present day erosional surface [Emparan & Pineda, 1999]. North of 29°30'S, no age equivalent strata is observed in the Chilean Pre-Cordillera until ~28°S, where the volcanic/sedimentary products infilling the Hornitos basin are observed (Abad, 1976a & b, 1980) and a similar age has been confirmed by radiometric dating [Emparan & Pineda, 1999] for these units.

**Arc volcanism**

**Late Cretaceous Volcanic Centres** [Emparan & Pineda, 1999]

Much of the central northern area of the Condoriaco-Rivadavia mapsheet [Emparan & Pineada, 1999] is dominated by the large collapse structure of Caldera Condoriaco (85Ma) and the much smaller, volumetrically, Cerro El Indio caldera (85-72 Ma), marking a switch to volcanism that is more focussed on distinct structures rather than the diffuse volcanism of previous times. Caldera Condoriaco developed upon the lower Cretaceous Quebrada Marquesa Formation and its' collapse-related deposits are probably represented by ignimbritic deposits within the Vinita Formation. Lacustrine-type sediments observed in the middle
section of the Vinita Formation are probably associated with the formation of a caldera lake.

Emparan & Pineda (2000) identified a suite of hypabyssal andesitic-dacitic intrusions, exposed in the western margin of Caldera Condoriaco and the underlying Vinita Formation. These high level intrusions probably represent the final remnants of ascending magma, resurgence, remaining after the caldera collapse event and occur at the easternmost locus of the 96-93 Ma magmatic arc with K-Ar dating indicating an age of 83-76Ma. The Condoriaco caldera at 30 km by 30 km was in effect a supervolcano and the published descriptions and our own field observations indicate that this was a period of widespread hydrothermal alteration and hence a possible remagnetisation event. The younger calderas described below are all much smaller, of localised effect and nest within the larger Condoriaco Caldera.

**Paleocene Volcanic Centres [Emparan & Pineda, 1999]**

The Paleocene was dominated by another phase of caldera formation including calderas Llano Perrada (c. 65Ma) and Tierras Blancas (Late Paleocene to Early Eocene), and the formation of the volcanic complexes La Corina (c. 60Ma) and Cerro El Inca (c. 60Ma), the products of which partially inter-finger with or rest paraconformably on the Los Elquinos Formation [Emparan & Pineda, 2000b]. The products of Caldera Llano Perrada outcrop separately in the extreme north-east of Caldera Condoriaco, upon which they rest paraconformably, indicating that little deformation occurred between the two volcanic phases in this area.

**Eocene Volcanic Centres [Emparan & Pineda, 1999]**

The mid Eocene is dominated by the volcanic complexes Cerro Blanco and Altos de Yaretal, which represent the final phase of volcanism recognised in the
Condoriaco area and again rest paraconformably on top of the Vinita Formation and collapse products of Caldera Condoriaco. K-Ar dates indicate an age of 45Ma [Emparan & Pineda, 2000b].

**Arc magmatism**

**The latest Cretaceous-Palaeocene Magmatic Arc**

*Abad* (1976a, b) reports two intrusive magmatic cycles in the latest Cretaceous and early Tertiary magmatic arc. The first is of very latest Cretaceous- Early Palaeocene age, contemporaneous with the Los Elquinos Formation and Quebrada Yungay Strata and comprises a suite of granitic to dioritic plutons intruding the Cerrillos (Vinita) Formation. As indicated on Figure 4.1-2, these plutons effectively mark the transition from Costal Cordillera to Precordillera as defined herein. This plutonic suite can be traced more or less continuously between Inca de Oro at 26°S [Taylor *et al.*, 2007] to the eastern margin of Caldera Condoriaco. Radiometric dating of plutons in this unit, which includes both plutons Las Campañas and Corredores which were sampled in the Los Choros area, has produced dates between 72-65 Ma and perhaps as young as 62Ma [Taylor *et al.*, 2007] in the Copiapó area. These plutons tend to form high ground immediately east of or intruding the Chañarcillo fold and thrust belt discussed below.

**The Eocene arc**

Outcropping to the east of the latest Cretaceous-Palaeocene arc are a sporadic series of plutons which can be traced in a discontinuous fashion from at least ~25°S to 30°S and beyond. These are closely associated with the Domeyko Fault system (see below) and ages range from 55 Ma-39Ma with the older granitic/dioritic plutons normally linked to discrete caldera structures.
Subsequent to the emplacement of the Eocene magmatic arc the active arc migrated eastwards towards its' present day position in the high Andes.

**Principal Structures**

The principal structure in the Precordillera region at the latitude of La Serena (c.30°S) is the transpressional Vicuna Fault system, which forms a southern strand of the larger Domeyko-West Fissure Fault Zone. In the far south-east corner of the Condoriaco area, *Emparan & Pineda* (1999) map the steep reverse faults belonging to the Vicuna Fault system as bringing areas of Palaeozoic aged basement to the surface (Figure 4.1-2). This is extremely similar to the situation observed in the La Guardia area (discussed in the next chapter). As all of the sampling localities in the Condoriaco and Tres Cruces area lie well to the west of the Vicuna Fault system, it is not considered further.

**4.1.3 Palaeomagnetic Sampling in the Condoriaco area**

*Palmer et al.,* (1980a) sampled what they believed was an approximately 6000m continuous section of predominantly Cretaceous aged volcanics in the Condoriaco area (Figure 4.1-2). A total of 36 sampling sites were collected along the E-W Rio Elqui and adjoining NE-SW Quebrada Marquesa valleys, with samples collected either as orientated cores or blocks. The data will be regrouped using the revised stratigraphical relationships outlined above and edited to remove data that is considered by modern standards to be unacceptable. The subsequent groupings will then need to be re-assessed for fold and reversal tests and the magnitude of rotations recalculated using modern reference poles [*Besse & Courtillot* 2002, 2003].
4.1.4 Palaeomagnetic Sampling in the Tres Cruces area

Palaeomagnetic samples were collected as orientated cores (see Chapter two for explanation), both along and to the north of Quebrada de Los Choros, immediately to the east of pueblo Punta Colorada (Figure 4.1-1B). Samples were also collected from Pluton Las Campañas near to the southern termination of the Chañarcillo Fold and Thrust Belt (c.29°00'S), in the north-east quadrant of the main field area and from Pluton Tilgo to the west of the main sampling area, close to the coast (c. 29° 30' S-Figure 4.1-2).

In total 51 sites of six or more samples were collected from a range of lithologies in 6 principal units as noted below and whose locations on Figure 4.1-2 correspond to the following:

1. Lower Cretaceous Arqueros/Quebrada Marquesa Formation
2. Quebrada Los Choros Dyke Swarm (80-90Ma)
3. Late Jurassic Pluton Tilgo (c.145Ma)
4. Mid-Cretaceous Pluton Los Colorados (c.90Ma)
5. Latest Cretaceous-Earliest Paleocene Pluton Las Campanas
6. Latest Cretaceous-Earliest Paleocene Pluton Corredores
4.2 Re-interpretation of the existing palaeomagnetic data from the Condoriaco area, c.30°S

4.2.1 Introduction

Palaeomagnetic data from the study of Palmer et al., (1980a) suggests that a primary, or at least pre-tilt remanence is isolated from the dominantly Cretaceous aged material in the Condoriaco area. A composite pole calculated using palaeomagnetic data from the three oldest formations (Arqueros, Quebrada Marquesa and Vinita Formations), appears to record clockwise crustal rotation of \(-15° \pm 11°\), when it was compared to the best estimate of the mean Cretaceous pole for South America at that time. Palmer et al., (1980a) ascribed this rotation to a localised deformation without defining a possible cause.

The groupings of the palaeomagnetic data from the study of Palmer et al., (1980a) have been re-interpreted following the stratigraphic revisions proposed by Emparan & Pineda (1999) and Mourgues (2000a) (Section 4.1). Site locations are shown displayed in Figure 4.1-2 based on the modern map and the localities listed in Table 4.2-1.

The reclassified stratigraphy of the Condoriaco area (Emparan & Pineda, 1999) primarily affects;

i) the age of the units sampled

ii) equates the Arqueros and Quebrada Marquesa formations which are now laterally equivalent (see section 4.1.2) and hence grouped together in further discussions.

iii) redefines the samples collected from the base of the Vinita Formation, as being from the the Quebrada La Totora strata
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<th>Long (°E)</th>
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Table 4.2-1: Site location and tilt correction of sampling localities from Palmer et al., (1980a). Allocation to particular stratigraphic units follows the stratigraphy of Emparan & Pineda (1999). The major significant change concerns samples originally sampled from the base of the Vinita Formation, now interpreted as forming the Quebrada La Totora strata. Arqueros & Quebrada Marquesa Formations now considered as being laterally equivalent following the stratigraphy of Mourgues (2000a, b & c), Perez & Reyes (2000).
The original palaeomagnetic data from Palmer et al., (1980a) did not include $\alpha_{95}$ values, and therefore these have been estimated using the standard approximation;

$$\alpha_{95} \approx \frac{140}{\sqrt{\kappa \times N}}$$ (Equation 4.2.1)

where $\kappa$ is the Fisherian precision parameter and $N$ the number of samples. This allows an easier comparison with other data and in particular that of this study. The original data were then filtered to remove poorly constrained directions, generally removing those block sample mean directions with angular dispersions $\gg 20^\circ$, and then new site and overall formation mean directions calculated.

4.2.2 Arqueros and Quebrada Marquesa Formations

Palmer et al., (1980a) sampled the Arqueros Formation at seven locations, six of which produced mutually consistent site mean directions, whilst the Quebrada Marquesa Formation was sampled at fourteen localities, with the mean directions of three sites rejected by the original authors due to incomplete demagnetisation (Table 4.2-2).

The Quebrada Marquesa samples are collected from both limbs of a broad anticline (Figure 4.1-2) and as such, the application of a fold test is a valid test of the origin of the isolated magnetisation. Whilst the scatter between site mean directions determined from the Quebrada Marquesa Formation is clearly decreased in tilt corrected coordinates, indicating a pre-tilt remanence, only a small improvement is noticed for the Arqueros samples (Table 4.2-2). This is, most likely, due to the consistently easterly dips at the sampling localities (Table 4.2.1).

Assuming that the Arqueros and Quebrada Marquesa Formations are lateral equivalents the two formations should record similar magnetisation directions.
Table 4.2-2  
Site mean directions for the Arqueros and Quebrada Marquesa Formations (recalculated after Palmer et al., 1980a). The highlighted direction (LS19) was not used to calculate the overall mean direction due to it being of reverse polarity. As the Arqueros Formation is interpreted to carry a primary remanence, acquired during the Cretaceous Long Normal Period, the reverse polarity direction is considered to be suspect.

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<td>10</td>
<td>12.5</td>
<td>-48.3</td>
<td>359.3</td>
<td>-53.5</td>
</tr>
</tbody>
</table>

Table 4.2-3  
Site mean directions for the Quebrada La Totora Strata and Vinita Formation (recalculated after Palmer et al., 1980a).
Given the ~6° difference in declination and inclination they are, well within error, the same. The minor discrepancy between them could be a function of localised tectonics, variation in the nature of the material sampled, or a function in the difference in numbers of sampled localities or some combination thereof. Both formations record only normal polarity directions (with exception of site LS19 from the Arqueros Formation), consistent with being acquired during the Cretaceous long normal polarity superchron, and record a similar amount of within group scatter suggesting that the effects of PSV are likely to have been averaged (Table 4.2-2). It is therefore entirely appropriate on both palaeomagnetic and stratigraphic grounds to combine the palaeomagnetic data from the two formations.

The combined data clearly passes a fold test, indicating a primary pre-folding remanence was isolated (Figure 4.2-1). Assuming a Barremian-Hauterivian age [Mourgues 2000b], the Arqueros and Quebrada Marquesa Formations record 17.9° (± 6.4°) of CW rotation when compared to the 130Ma reference pole of Besse & Courtillot (2002, 2003), whilst negligible inclination flattening is observed (0.8° (± 4.6°)), further supporting the observation that a Primary magnetisation has been recorded.

4.2.3 Vinita Formation and the Quebrada La Totora Strata

Originally thirteen sites were sampled from the Albian-Cenomanian Vinita Formation, seven of which would now be regarded as coming from the base of the Quebrada La Totora Strata (Emparan & Pineda 1999). The remaining six sites are still within the upper part of the Vinita Formation (Table 4.2-1, Figure 4.1-2).

Only homoclinal (westerly) dipping strata were sampled from both units and the application of a tilt correction has no significant effect on the overall dispersal of individual site mean directions and therefore does not help in the interpretation of
Figure 4.2-1 Stereonet plot of individual site mean directions and overall mean direction calculated for the combined palaeomagnetic data from the Arqueros & Quebrada Marquesa Formations (recalculated from Palmer et al., (1980a) after Mourges (2000a) & Perez & Reyes (2000)).
the primary or secondary nature of the magnetisation (Table 4.2-3). However the combined result from the underlying Arqueros and Quebrada Marquesa Formations, demonstrably record primary pre-tilt magnetisations, the amount of rotation recorded by these formations should therefore set an upper limit to the possible magnitude of rotation that may be recorded by the younger Vinita Formation and Quebrada La Totora Strata.

The mean in-situ magnetisation declination recorded by both the Quebrada La Tototora Strata and Vinita Formation, would appear to indicate that a large clockwise rotation of up to 30-40° in magnitude is recorded (regardless of the age of re-magnetisation). This is difficult to rationalise with the ~18° of CW rotation recorded by the older formations (Table 4.2-2) and therefore the magnetisation isolated from the Quebrada La Totora Strata and Vinita Formation are, more logically, interpreted to be primary and therefore record similar magnitudes of crustal rotation to that recorded by the Arqueros & Quebrada Marquesa Formations when compared to the appropriate reference poles.

Although the slightly older Quebrada La Totora Strata appears to record slightly greater CW rotation (more easterly declination) than the Vinita Formation (Table 4.2-3), the relatively small number of sites within each unit means this discrepancy may be an artefact of under-sampling. Both units record only normal polarity directions, consistent with a magnetisation associated the Cretaceous long normal polarity superchron and display similar scatter of site mean directions. This would indicate that although Emparan & Pineda (1999) discriminate the two units based on field characteristics, both units record a very similar magnetic field direction and it is therefore felt appropriate to combine the palaeomagnetic data and ascribe it a Albian-Cenomanian age (Table 4.2-3, Figure 4.2-2). The combined direction indicates 10.6° (± 10.2°) of clockwise rotation is recorded, when compared to the
Figure 4.2-2  Stereonet plot of individual site mean directions and overall mean direction calculated for the combined palaeomagnetic data from the Quebrada La Totora Strata and Vinita Formation (recalculated from Palmer et al., (1980a) after Emperan & Pineda (1999)).
100Ma reference pole of Besse & Courtillot (2002, 2003) with a statistically non-significant amount of inclination flattening (5.4° ± 11.6°).

4.2.4 Elquinos Formation

Palmer et al., (1980a), also presented palaeomagnetic data from six locations in the Elquinos Formation, which is now assigned a Maastrichtian age [Emparan & Pineda 1999, 2000]. The characteristic component of remanence (Table 4.2-4) appears to represent a secondary (post-tilting) remanence, with obvious dispersion of site mean ChRM directions on application of the tilt-correction (Figure 4.2-3). The restricted number of sampling localities and limited variation in bedding attitude however (Table 4.2.1) means the tilt-test is statistically indeterminate (Table 4.2-4). In addition, the amount of between-site scatter and the identification of a single reverse polarity direction, suggests that the recorded magnetisations were acquired over a reasonable length of time and certainly not the result of a single instantaneous remagnetisation event (Figure 4.2-3) and therefore does not fully preclude the possibility that a primary magnetisation is recorded.

Assuming a primary magnetisation age to be Maastrichtian, comparison with the 70Ma reference pole of Besse & Courtillot (2002, 2003), would indicate a statistically insignificant clockwise rotation (4.5° (± 15.3) and inclination flattening of -8.2° (± 12.2°). The shallowness of the inclination could be the result of inclination flattening.

Assuming a secondary magnetisation for the Elquinos Fm. brings with it the problems of determining timing of remagnetisation. Several younger periods of volcanism and magmatism are recognised in the Condoriaco area that could potentially be responsible for remagnetisation of the Elquinos Formation and
Table 4.2-4  Site mean directions for the Elquinos Formation (recalculated after Palmer et al., 1980a).

<table>
<thead>
<tr>
<th>Sampling Unit</th>
<th>Age of Sampling Unit</th>
<th>Inferred Age of Magnetisation</th>
<th>Mean Direction Dec. (°)</th>
<th>Mean Direction Inc. (°)</th>
<th>Reference Pole (B &amp; C, 2002)</th>
<th>Rotation ± eor (°)</th>
<th>Flattening ± error (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arqueros &amp; QM Formations</td>
<td>120-140 Ma</td>
<td>Primary-130 Ma</td>
<td>11.9</td>
<td>-44.4</td>
<td>130 Ma</td>
<td>17.9 ± 6.4</td>
<td>0.8 ± 4.6</td>
</tr>
<tr>
<td>QLT Strata &amp; Vinita Formation</td>
<td>110-95 Ma</td>
<td>Primary-100 Ma</td>
<td>7.0</td>
<td>-55.6</td>
<td>100 Ma</td>
<td>10.6 ± 10.2</td>
<td>5.4 ± 11.6</td>
</tr>
<tr>
<td>Elquinos Formation</td>
<td>71-65 Ma</td>
<td>Primary-70 Ma</td>
<td>356.0</td>
<td>-42.0</td>
<td>70 Ma</td>
<td>4.5 ± 15.3</td>
<td>-8.2 ± 12.2</td>
</tr>
<tr>
<td>Elquinos Formation</td>
<td>71-65 Ma</td>
<td>Secondary-60 Ma</td>
<td>11.2</td>
<td>-56.5</td>
<td>60 Ma</td>
<td>18.6 ± 12.2</td>
<td>-1.9 ± 7.6</td>
</tr>
</tbody>
</table>

Table 4.2-5  Overall mean directions recalculated from the palaeomagnetic study of Palmer et al., (1980a) and the magnitude of crustal rotation with respect to continental South America, using the reference poles of Besse & Courtillot (2002, 2003).
Figure 4.2-3  Stereonet plot of individual site mean directions and overall mean direction calculated for the combined palaeomagnetic data from the Elquinos Formation (recalculated from Palmer et al., (1980a)).
associated simple block tilting events could have caused the minor variation in dips recorded. These events include the intrusion of the latest Cretaceous-earliest Paleocene and Paleocene-Eocene magmatic arcs and two periods of volcanism during the Paleocene and Eocene.

As the palaeomagnetic data for the underlying strata record primary magnetisations (as previously argued), the remagnetisation of the Elquinos Formation must represent a very localised occurrence. One such event could be hydrothermal circulation associated with the Paleocene La Corina and Cerro El Inco volcanic centres (~60Ma) [Emparan & Pineda, 1999] whose deposits rest conformably upon the Elquinos Formation in the north of the sampling area (Figure 4.1-3). Comparison of the overall mean direction to the 60Ma reference pole of Besse & Courtillot (2002, 2003) indicates that the Elquinos Formation records 18.6° (± 12.2°) of CW rotation, with negligible inclination flattening of -1.9° (± 7.6°) observed. Little significant difference is noted in the magnitude of rotation recorded if a younger (50-40Ma) remagnetisation is assumed.

4.2.5 Summary

The clearly pre-tilt magnetisation isolated from the Arqueros & Quebrada Marquesa Formations gives a strong argument to suggest that the Quebrada La Totora Strata and Vinita Formation also record a pre-tilt magnetisation. Although the two groups of strata record CW rotations of similar magnitudes (Table 4.2-5), the older strata appears to record a greater amount of rotation, suggesting a possible early Cretaceous component of rotation (of the order of 5-10°) may have been isolated.

Palaeomagnetic data from the latest Cretaceous-Paleocene Elquinos Formation, does not pass or fail a fold test, predominantly due to the small number of
sampling localities, but does show significantly greater clustering in situ, suggesting a secondary (post-tilt) magnetisation has been isolated (Table 4.2-4, Figure 4.2-3). Assuming this to be correct, the Elquinos Formation records an identical amount of rotation to the Arqueros & Quebrada Marquesa Formations, irrespective of whether a 60, 50 or 40Ma reference pole is used (Table 4.2-5). This would suggest that the overall rotation of the Condoriaco area occurred subsequent to the remagnetisation of the Elquinos Formation, tentatively placed at 60Ma. An implication of this is that no early Cretaceous component is identified.

If a Primary magnetisation is interpreted from the Elquinos Formation, despite the pronounced dispersal of site mean directions after tilt correction (Figure 4.2-3), little or no significant crustal rotation is recorded when compared to the 70Ma pole of Besse & Courtillot (2002, 2003-Table 4.2-5). This would imply that all of the crustal rotation affecting the Condoriaco area occurred prior to the Paleocene and further that ~10° occurred during the early Cretaceous. The pattern could also result from an eastward reduction in the magnitude of rotation eastwards, representing a spatial rather than temporal gradient. Whilst the possibility that a Primary magnetisation is preserved cannot be dismissed however, the markedly greater clustering in situ, suggests that the Elquinos Formation has been remagnetised.
4.3 Tres Cruces-Early Cretaceous Arqueros Formation and the Quebrada Los Choros dyke swarm

4.3.1 Arqueros Formation

In the Quebrada Los Choros area (Figure 4.1-2), the Mid-Late Cretaceous magmatic arc intrudes the early Cretaceous aged strata, belonging to the Bandurrias Group [Moscoso et al., 1982]. As discussed in section 4.2, Mourgues (2000b) suggests that a ~140m sequence of calcareous clastic rocks observed to crop out along a part of Quebrada Los Choros is lithologically and biostratigraphically correlated to the Arqueros Formation in the Condoriaco area, ~50km to the south (units 2 & 4 of Aguirre & Egert, 1965; facies C of Emparan & Pineda, 1999-Figure 4.3.1). More broadly, this correlation firmly links the Bandurrias Gp. rocks in the Tres Cruces area [as per Moscoso et al., 1982] with the equivalent Arqueros & Quebrada Marquesa Fms. ~50km to the south.

The most prominent beds in Quebrada de Los Choros are thick grey-brown limestones that contain a variety of highly disarticulated mollusca, poorly sorted bioclastic material, detrital feldspars and lithic fragments of a volcanic origin (sites AT2-11 & 12-Table 4.3-1). The low sphericity of the extraclasts suggests limited transport consistent with the Arqueros Formation having been deposited in a marginal back-arc environment. The remaining sediments are mostly well-lithified, calcareously cemented, volcaniclastic red sandstones (Figure 4.3.2-site AT2-16), with andesitic-basaltic lavas and volcanic breccias, which become more common towards the top of the formation, before lavas of andesitic-basaltic composition become the dominant lithology overlying the Arqueros sediments to the east.

Eleven sites of 6-8 samples were collected from the overall sedimentary/volcanic sequence, centred on the Tres Cruces area of Quebrada Los Choros (Table 4.3-1).
Figure 4.3-1  A) Map showing the logged sections of the Arqueros Formation in the Llano de Arqueros and Quebrada de Los Choros areas (modified from Mourgues, 2000b), with sampling localities indicated in the Quebrada de Los Choros area. Light grey shading represents the extent of calcareous clastic rocks (Arqueros Formation), whilst the dark grey shading represents dominantly volcanic rocks belonging to the Arqueros/Quebrada Marquesa Formation. B) Stratigraphic logs of the Arqueros Formation observed in the Llano de Arqueros and Quebrada de Los Choros areas (modified from Mourgues, 2000b).
Table 4.3-1 Site number, location, lithology and tilt correction of sampling localities collected from the early Cretaceous Arqueros Formation in the Tres Cruces area of Quebrada de Los Choros. Also shown are the typical demagnetisation ranges at which the component of ChRM was typically demagnetised for each site (°C-thermal and mT-AF demagnetisation techniques). This table format will be used for the remainder of the thesis.

<table>
<thead>
<tr>
<th>Site</th>
<th>Sampling Site</th>
<th>Lat (°S)</th>
<th>Long (°E)</th>
<th>Lithology</th>
<th>Tilt Correction</th>
<th>°C</th>
<th>ChRM isolation</th>
<th>mT</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Max.</td>
<td>Min.</td>
</tr>
<tr>
<td>AT2-11</td>
<td>29.37</td>
<td>289.08</td>
<td>Biomicritic Limestone</td>
<td>290/09</td>
<td>100-150</td>
<td>580-580</td>
<td>4-6</td>
<td>40-70</td>
</tr>
<tr>
<td>AT2-12</td>
<td>29.36</td>
<td>289.08</td>
<td>Biomicritic Limestone</td>
<td>344/06</td>
<td>150</td>
<td>580</td>
<td>6-10</td>
<td>40-50</td>
</tr>
<tr>
<td>AT2-13</td>
<td>29.36</td>
<td>289.07</td>
<td>Green andesitic silt</td>
<td>Horizontal</td>
<td>150</td>
<td>580</td>
<td>4-10</td>
<td>60-90</td>
</tr>
<tr>
<td>AT2-14</td>
<td>29.35</td>
<td>289.08</td>
<td>Basaltic-Andesitic Lava</td>
<td>328/08</td>
<td>250</td>
<td>670</td>
<td>4-15</td>
<td>60-100</td>
</tr>
<tr>
<td>AT2-15</td>
<td>29.34</td>
<td>289.07</td>
<td>V. fine grained red sandstone</td>
<td>343/09</td>
<td>150</td>
<td>670</td>
<td>2-15</td>
<td>70-100</td>
</tr>
<tr>
<td>AT2-16</td>
<td>29.34</td>
<td>289.06</td>
<td>Med. grained red sandstone</td>
<td>344/16</td>
<td>150</td>
<td>670</td>
<td>4-15</td>
<td>60-100</td>
</tr>
<tr>
<td>AT2-17</td>
<td>29.38</td>
<td>289.10</td>
<td>Green andesitic lava</td>
<td>303/16</td>
<td>100</td>
<td>400</td>
<td>4-10</td>
<td>30-45</td>
</tr>
<tr>
<td>AT2-18</td>
<td>29.38</td>
<td>289.09</td>
<td>Fine grained red mudstone</td>
<td>331/03</td>
<td>200-350</td>
<td>670</td>
<td>4-10</td>
<td>45-100</td>
</tr>
<tr>
<td>AT2-19</td>
<td>29.37</td>
<td>289.04</td>
<td>Basaltic Lava</td>
<td>354/04</td>
<td>150</td>
<td>620</td>
<td>4-15</td>
<td>40-80</td>
</tr>
<tr>
<td>AT2-20</td>
<td>29.37</td>
<td>289.02</td>
<td>Coarse grained grey sandstone</td>
<td>089/10</td>
<td>150</td>
<td>670</td>
<td>2-6</td>
<td>70-100</td>
</tr>
<tr>
<td>AT2-21</td>
<td>29.38</td>
<td>289.02</td>
<td>Fine grained red sandstone</td>
<td>313/04</td>
<td>250</td>
<td>620</td>
<td>4-15</td>
<td>70-100</td>
</tr>
</tbody>
</table>

Table 4.3-2 Site number, number of samples used in the calculation of site mean direction (n) from the total number of samples demagnetised from a particular site (N), mean ChRM directions (both in-situ and tilt-corrected) as well as the overall mean direction from the early Cretaceous Arqueros Formation. This table format will used to present palaeomagnetic data for the remainder of the thesis.

<table>
<thead>
<tr>
<th>Site</th>
<th>n/N</th>
<th>Dec. (°)</th>
<th>Inc. (°)</th>
<th>Tilt-corrected Dec. (°)</th>
<th>Inc. (°)</th>
<th>k</th>
<th>α95 (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT2-11</td>
<td>6/6</td>
<td>15.0</td>
<td>-48.9</td>
<td>13.9</td>
<td>-57.8</td>
<td>618.9</td>
<td>2.7</td>
</tr>
<tr>
<td>AT2-12</td>
<td>5/6</td>
<td>18.4</td>
<td>-48.9</td>
<td>12.2</td>
<td>-52.0</td>
<td>411.5</td>
<td>3.8</td>
</tr>
<tr>
<td>AT2-13</td>
<td>6/6</td>
<td>6.6</td>
<td>-55.0</td>
<td>6.6</td>
<td>-55.0</td>
<td>354.4</td>
<td>3.6</td>
</tr>
<tr>
<td>AT2-14</td>
<td>6/6</td>
<td>354.7</td>
<td>-46.2</td>
<td>346.7</td>
<td>-49.5</td>
<td>154.0</td>
<td>5.4</td>
</tr>
<tr>
<td>AT2-15</td>
<td>7/7</td>
<td>18.7</td>
<td>-52.2</td>
<td>7.9</td>
<td>-56.8</td>
<td>212.4</td>
<td>4.2</td>
</tr>
<tr>
<td>AT2-16</td>
<td>6/6</td>
<td>16.8</td>
<td>-50.3</td>
<td>356.7</td>
<td>-56.6</td>
<td>398.3</td>
<td>3.4</td>
</tr>
<tr>
<td>AT2-17</td>
<td>4/6</td>
<td>11.9</td>
<td>-47.5</td>
<td>1.9</td>
<td>-62.0</td>
<td>70.1</td>
<td>11.0</td>
</tr>
<tr>
<td>AT2-18</td>
<td>5/6</td>
<td>13.6</td>
<td>-54.3</td>
<td>10.8</td>
<td>-56.2</td>
<td>240.1</td>
<td>4.9</td>
</tr>
<tr>
<td>AT2-19</td>
<td>7/7</td>
<td>5.1</td>
<td>-61.1</td>
<td>367.9</td>
<td>-61.6</td>
<td>85.6</td>
<td>6.6</td>
</tr>
<tr>
<td>AT2-20</td>
<td>7/7</td>
<td>340.4</td>
<td>-55.1</td>
<td>351.5</td>
<td>-49.8</td>
<td>246.3</td>
<td>3.9</td>
</tr>
<tr>
<td>AT2-21</td>
<td>8/8</td>
<td>6.6</td>
<td>-52.5</td>
<td>13.0</td>
<td>-44.2</td>
<td>166.7</td>
<td>4.3</td>
</tr>
</tbody>
</table>

Mean | 11/11 | 8.3 | -52.5 | 3.5 | -55.0 | 97.0 | 4.7 |

Figure 4.3-2 Photograph illustrating the shallow (easterly) dip of the red volcaniclastic sandstones (site AT2-16) that become more prolific towards the top of the Arqueros Formation in the Quebrada de Los Choros area. View to the north.
and Figures 4.1-2, 4.3-1b). Two sites were sampled from large, blocky outcrops of bioclastic limestone, five from volcaniclastic sandstones and four from basaltic-andesitic lava flows. Although the exact position of the sequence documented by Mourgues (2000b) is unclear (Figure 4.3-1), the sampled strata closely mirrors the stratigraphy described and for this reason is assigned a Hauterivian-Barremian age [Mourgues, 2000a & c; Perez & Reyes, 2000].

Three sites were apparently sampled from stratigraphically below the subaqueously erupted lavas (displaying pillow-type structures), upon which the calcareous clastic rocks of the Arqueros Formation are founded (dark grey unit below horizon marked as A in Figure 4.3-1a). Although Mourgues (2000b) appears to indicate that this strata represents the basal volcanic unit (Unit 1) of Aguirre & Egert (1965), the precise nature of this strata is not made clear. As Moscoso et al., (1982) simply correlate the overall volcano-sedimentary package with the Arqueros & Quebrada Marquesa Formations to the south, these sites are also considered to be of Hauterivian-Barremian age.

*Magnetic Mineralogy*

The magnetic mineralogy of the samples collected from the Arqueros Formation is, unsurprisingly, related to the sample lithology and are therefore discussed with respect to the lithology as follows.

*Limestone (AT2-11 & 12)*

Investigation of the variation of bulk susceptibility with temperature shows near reversible heating and cooling curves (Figure 4.3-3A), suggesting that heating produces only limited thermochemical alteration. Although bulk susceptibility is observed to decrease steadily up to temperatures of ~500°, suggesting that ti-rich magnetite may be present, the largest drop is observed between 500-580°, which
Figure 4.3-3  Magnetic mineralogy experiments on A) limestone samples (Site AT2-11) with titanomagnetite as the dominant carrier of magnetisation, B) red volcaniclastic sandstones subdivided into haematite and magnetite dominated lithologies, C) volcanic samples subdivided into green andesitic and dark basaltic-andesitic lavas, all from the Arqueros Formation in the Tres Cruces area (standard nomenclature).
would indicate that titanium poor magnetite is the dominant magnetic mineral. IRM acquisition experiments also indicate titanomagnetite to be the dominant magnetic carrier, with rapid acquisition of magnetisation observed in fields up to 100mT. The majority of samples failed to completely saturate in the maximum field of 800mT however, which would indicate the presence of a higher coercivity mineral such as haematite (Figure 4.3-3A), probably derived from the alteration of magnetite. During backfield coercivity experiments fields of ~40mT were required to oppose the initial IRM, indicating that the lower coercivity minerals (magnetite) dominate the magnetic mineral assemblage (Figure 4.3-3).

Thermal demagnetisation of a composite IRM also indicates that the hard (800mT) component does not significantly contribute to the overall magnetisation suggesting haematite contributes little to the overall remanence (Figure 4.3.3). The high coercivity behaviour observed during IRM acquisition could result from the presence of pyrrhotite (possibly derived from the alteration of pyrite, a process that is commonly associated with limestone).

The soft and intermediate components are almost identical in magnitude and are progressively destroyed up to temperatures of 540°C. This indicates that a range of compositions, or grain sizes were present, contributing equally to the overall remanence. The remanence is destroyed on treatment to 580°C, clearly indicating magnetite as the primary magnetic mineral. The presence of pyrrhotite is not obvious during thermal demagnetisation of a composite IRM, with the soft and medium coercivity components obviously carried by magnetite. To conclude the magnetic mineralogy of the limestones is dominated by magnetite, which may relate to the presence of volcanic detritus [e.g. Mourguès, 2000b] and suggests that the samples could be expected to record a reasonably intense DRM.

_Volcaniclastic sandstones (AT2-15, 16, 18, 20 & 21)_
All of the volcaniclastic sandstone (and mudstone) samples display very similar magnetic properties and indicate the presence of two-three magnetic minerals, albeit in varying amounts and are discussed under three headings below.

The volcaniclastic sandstones samples from Quebrada Los Choros demonstrate a range of rock magnetic behaviours, relating to the relative amounts of magnetite and haematite present. Figure 4.3-3b shows the results from rock magnetic experiments from two samples. The first, a sample from Site AT216, is dominated by haematite, whilst the second from site AT218 shows that magnetite is dominant magnetic mineral present.

High temperature (low-field) susceptibility experiments indicate that a significant drop in susceptibility is observed between 500-580°C for both samples (attributed to ti-poor magnetite), but this is more pronounced for sample AT218-05B, suggesting that magnetite dominates the magnetic mineral assemblage in this sample. Both samples indicate that haematite is also present (and dominant in the sample from site AT216), with the bulk susceptibility reduced to zero after heating to 670°C. As haematite is a much lower susceptibility mineral than magnetite (Chapter Two), the physical quantity of haematite in the sample from AT216 must be significantly greater than the actual amount of magnetite in the sample and therefore could perhaps be expected to provide a greater contribution towards carrying the observed magnetisation.

Sample AT218-05B also displays a marked peak in the heating curve between 150-350°C that is absent during cooling, suggesting that the sample has undergone some thermochemical alteration during heating. This is attributed to the presence of either maghemite or pyrrhotite, as both minerals are either transformed to haematite (maghemite), or reach Curie temperature (pyrrhotite) around this temperature.
IRM experiments reinforce the observation that magnetite and haematite are the dominant magnetic minerals within the sandstones, but again slightly different behaviours are noted reflecting the relative quantities of the two minerals. Magnetisation is rapidly acquired by both samples in fields up to 80-100mT, although this only accounts for ~20% of the total magnetisation acquired in the highest applied field by sample AT216-07B compared to 90% by sample AT218-05B (Figure 4.3-3B). Sample AT216-07B acquires the majority of the overall IRM between 80-800mT, without the sample becoming saturated, whilst sample AT218-05B becomes saturated by treatment to 300mT. Backfield coercivity experiments again reflect the relative amounts of magnetite and haematite present, with applied fields of 170-390mT required to oppose the initial applied IRM for haematite dominated samples, but only of 60-80mT to oppose that applied to magnetite dominated samples.

Thermal demagnetisation of a three component IRM applied to sample AT216-07B are dominated by the hard (800mT) and intermediate (300mT) magnetisations, which are destroyed by thermal demagnetisation at 670°C. A smaller soft component (50mT) is also observed and destroyed by treatment to 580°C, with the magnitude of this magnetisation related to the amount of magnetite present. In contrast, that applied to sample AT218-05B is dominated by the soft and intermediate coercivity components, both of which are significantly destroyed by treatment to 300°C. The soft component is then destroyed by treatment to 570°C (magnetite) whilst the intermediate component is destroyed (along with the hard component) at 670°C consistent with the Neel temperature of haematite (Figure 4.3-3b). The unblocking of the low and intermediate components is attributed to pyrrhotite, mainly because above this temperature the two components appear to be carried by different minerals.
Thin section analysis of these sandstones indicates the presence of large amounts of pigmentory haematite as well as large areas of intense haematisation. The actual development of (pigmentory) haematite is a complex secondary process in sandstones, but in this case it appears linked to the large amounts of volcaniclastic material and is likely to have evolved through the oxidisation of magnetite within pre-existing volcanic lithic fragments. This process may have initiated prior to sedimentary deposition, either during the initial volcanic activity or subsequent erosion. Pigmentory haematite has been shown in many cases to retain reliable records of the magnetic field and is therefore considered a potential carrier of magnetisation; its development however can occur many millions of years after the initial deposition, in which case the magnetisation will not be temporally associated with depositional processes. The presence of a large amount of extraclastic detritus of a volcanic origin also suggests that a reasonably intense DRM carried by magnetite, may be expected.

**Volcanics (AT2-13, 14, 17 & 19)**

The volcanic units sampled in Quebrada de Los Choros are comprised of two green andesitic and two dark lava flows of more basaltic composition. Rock magnetic experiments indicate that whilst the magnetic mineralogy of all of the lavas is dominated by magnetite, the accessory magnetic minerals observed change with the composition of the lava.

High temperature susceptibility experiments indicate that magnetite dominates the magnetic mineral assemblage of all of the lava flows sampled, with a Curie temperature of 580°C clearly defined (Figure 4.3-3c). The green andesitic lavas however display a significant (Hopkinson) peak between 150-350°C during heating (typified here by sample AT213-05), which is absent during cooling suggesting that some thermochemical alteration has occurred. Although the samples from the dark
basaltic lavas (sample AT219-07B) also display a substantially reduced susceptibility during cooling, only a very small peak is observed during the heating at ~350°C. The alteration products are clearly of lower susceptibility with the cooling curve clearly reduced between 400-35°C in comparison to the heating curve. This could either represent the alteration of pyrrhotite (of a high susceptibility composition such as Fe₇S₈ (Hunt et al., 1995)) to magnetite of a lower susceptibility form (e.g. Ti-rich), or of maghemite to haematite.

The IRM acquisition curve of sample AT213-05C (green andesite) indicates that the majority of magnetisation is acquired between 100-200mT, with magnetisation only acquired slowly in lower magnitude fields. In contrast, sample AT219-06 (dark basalt), rapidly acquires magnetisation in fields up to 100mT and is saturated in a 300mT applied field compared to the 500mT required to saturate the green andesite. Larger coercive (back-) fields are required to oppose the initial IRM remanence of the green andesites, suggesting that the magnetic mineralogy of these lavas comprises slightly higher coercivity minerals.

The observations from the IRM acquisition experiments are consistent with increased amounts of pyrrhotite in sample AT213-05C and this is confirmed by the thermal destruction of a composite IRM. The intensity of the initial remanence was so great that the sample saturated the Molspin magnetometer until it was treated to 150°C. After this point the intermediate (300mT) component clearly dominates the remanence and is destroyed by treatment to 300°C (consistent with pyrrhotite). The remaining remanence (carried by the low and intermediate components) is destroyed by treatment to 580°C suggesting that magnetite carries what remains of the composite remanence.

The dark basaltic samples are more obviously dominated by magnetite, with the low and intermediate coercivity components destroyed equally by treatment to
580°C (although a small reduction is noted in the intermediate component at 300°C may represent a slight contribution from pyrrhotite). Interestingly the hard (800mT) component appears to record a significant goethite contribution to the initial remanence, which is unblocked by treatment to 100°C. This may or may not be a genuine observation as the initial intensity of the magnetisation was close to saturating the magnetometer, but goethite could represent an accessory mineral within the lava.

Demagnetisation Behaviour

Very similar demagnetisation trends were observed from all of the samples collected from the Arqueros Formation in Quebrada de Los Choros, irrespective of lithology. The demagnetisation behaviour of each lithology relates very closely to the magnetic mineralogy however and therefore will be discussed separately for each rock type.

Limestone

The limestone samples demagnetised fully using both thermal and AF methods (Figure 4.3-4a). Generally only a single vector of magnetisation was observed and removed on application of >50-70mT or between 400-580°C when approximately 50% of the initial NRM intensity was unblocked (Figure 4.3-4a). Samples treated thermally from site AT211 also showed evidence of a high temperature (reverse polarity) direction up to 670°C that could be loosely constrained by great-circle analysis.

The wide thermal range and low-intermediate coercivities across which ChRM is demagnetised indicates that multi-domain (MD) or titanium rich magnetite is the primary carrier of magnetisation, consistent with the results from rock magnetic experiments. A very small amount of hematite is present within the samples, which
Figure 5.3.4  Representative Zijderveld plots for typical Thermal and AF demagnetisation behaviour of A) limestone samples, B) sandstone samples (haematite and magnetite dominated) and C) volcanic samples (green andesitic and dark basaltic-andesitic lavas) sampled from the Arqueros Formation in the Tres Cruces area. The shaded section on each Zijderveld plot represents the component of magnetisation identified as the ChRM direction and this convention will be used throughout the remainder of this thesis.
could be responsible for maintaining a separate (reverse) direction, but hematite also appears to be produced during heating. The higher temperature thermal demagnetisation steps were accompanied by an increase in the observed bulk susceptibility, perhaps indicating that the high temperature directions observed may well be artificially acquired during heating and the result of thermo-chemical changes.

**Volcaniclastic sandstones**

As would be expected from the mixed magnetite-hematite magnetic mineralogy established for the sandstone samples, whilst AF demagnetisation effectively removed the magnetite component of magnetisation complete demagnetisation of the hematite component wasn’t achieved until treatment to 670°C (Figure 4.3-4b). Both the magnetite and hematite components carry an identical direction, with the overall remanence observed to be univectorial and origin-bound (particularly clear from the magnetite dominated samples in Figure 4.3-4b). This suggests that both mineral fractions were magnetised contemporaneously.

**Volcanics**

As noted for the sandstone samples above, the mixed magnetic mineralogy (in this case magnetite & pyrrhotite) established for the two types of lava sampled from the Arqueros Formation is clearly evident in carrying the isolated remanence, with both minerals observed to carry the same direction of remanent magnetisation (Figure 4.3-4c). AF demagnetisation clearly demagnetises the magnetite component, to reveal an origin bound direction, although the overall remanence isn’t necessarily fully demagnetised due to the higher coercivity behaviour of pyrrhotite. Thermal demagnetisation clearly shows that a significant component is removed by treatment to 350°C, most obviously for the green andesitic samples,
with the magnetite component destroyed by treatment to 580°C (Figure 4.3-4c). Again the identical direction carried by both of the minerals suggests that they were magnetised simultaneously.

**Discussion of palaeomagnetic data from the Arqueros Formation**

Generally speaking a very small, low temperature/low coercivity direction was removed during the demagnetisation of all of the samples. This usually was only constrained by the initial NRM direction and first demagnetisation point and often the direction wasn’t far removed from that of the ChRM direction (Figure 4.3-5a). A stereonet plot of these directions shows that whilst they are reasonably widely scattered, some clustering is observed in the upper (hemisphere) NE quadrant, not too far removed from the present-day field direction. These low components may therefore represent a present-day overprint acquired in situ, or the reasonable scatter may reflect the acquisition of short-term viscous magnetisation acquired during sampling or subsequent storage.

Although all of the site-mean directions calculated were of normal polarity, they are not concordant with the present-day field direction (Table 4.3-2 and Figure 4.3.5b & c), indicating that an ancient record of the Earth’s magnetic field has been isolated. The observation that all of the lithologies sampled display a single component of magnetisation, which is carried by multiple magnetic minerals, suggests that the sediments and volcanics of the Arqueros Formation in Quebrada de Los Choros record only a single magnetisation event. This 'event' could either represent a Primary magnetisation, meaning that all of the magnetic minerals identified as carrying the ChRM, must have been present at the time of deposition/extrusion, or that the or the area has been subject to complete remagnetisation.
Figure 4.3-5 Stereonet projections of A) in situ low/coercivity/low temperature components (solid circles-upper hemisphere projection), B) in situ site mean ChRM directions and C) tilt corrected site mean ChRM directions from the early Cretaceous Arqueros Formation, Tres Cruces area.
Due to the shallow homoclinal E-SE dip of the outcrop (Table 4.3-1), tilt-corrections have little effect on the overall distribution of individual site mean directions (Table 4.3-2 and Figure 4.3-5b & c.). For this reason, no meaningful discussion of the age of the overall ChRM direction calculated for the Arqueros Formation may be made with respect to deformation/tilting of the Tres Cruces area. As a result, a tilt test of the data is indeterminate, although the slightly better clustering of the site-mean directions on application of the tilt-correction, as well as the fact that the equivalent strata ~50km to the south in the Condoriaco area clearly records a pre-tilt magnetisation, may suggest that it is marginally more likely that a primary, pre-tilt, magnetisation has been isolated in Quebrada de Los Choros (Table 4.3-2).

Although several areas of intense mineralisation were encountered in the Tres Cruces area, these were avoided for the purposes of palaeomagnetic sampling. There is no evidence at outcrop scale to suggest that the Arqueros Formation in Quebrada de Los Choros has undergone significant alteration and thin section analysis of some of the sampled units confirms this. The overall formation mean direction recorded by the Arqueros Formation is therefore tentatively interpreted as being pre-tilt in origin, recording an overall direction of D=003.5°, I=-55.0°, with an associated $\alpha_{95}$ error of 4.4° (Table 4.3-2, Figure 4.3-5b).

4.3.2 Quebrada Los Choros dyke swarm

To the east of the small village of Punta Colorada (Figure 4.1-1b), the base of the Arqueros Formation and the predominantly volcanic strata that underlies it, are intruded by a suite of NW-SE trending, dark green sub-vertical dykes of andesitic composition, of generally between 5-10m in width (Figure 4.3-6). Emparan & Pineda (1999) map a N-S orientated dyke swarm to intrude the Arqueros and Quebrada Marquesa Formations immediately to the south of Quebrada Los
Figure 4.3-6  Photograph of the Quebrada de Los Choros dyke swarm, taken facing to the north. Dykes are subvertical and trend NW-SE. Truck in foreground for scale.

Table 4.3-3  Site location, lithology and attitude of dykes sampled from the Quebrada de Los Choros dyke swarm intruding the Arqueros (Quebrada Marquesa) Formation to the east of Punta Colorada.

Table 4.3-4  Site mean ChRM directions and overall mean direction, Quebrada de Los Choros dyke swarm. Sites highlighted grey either produced no stable ChRM direction (AT2-41) or produced a clearly anomalous ChRM direction (AT2-46) that was excluded from the overall mean direction.
Choros. These dykes also intrude the Vinita Formation (105-90Ma), but do not appear to intrude the younger Elquinos Formation (78-65Ma-Figure 4.1-2), suggesting the dyke swarm was emplaced between 90-80Ma (Section 4.1). In total seven dykes were sampled including the chilled margins two cases (Table 4.3-3).

**Magnetic Mineralogy**

Magnetic mineralogy experiments indicate that all of the dykes sampled contain magnetite as a dominant carrier of magnetisation, but with varying amounts of what is interpreted to represent pyrrhotite. High temperature susceptibility experiments on all of the dyke samples tested indicate very tightly constrained Curie points at ~580°C, consistent with the presence of ti-poor magnetite (Figure 4.3-7). All samples also display a peak in bulk susceptibility between 250-350°C to a greater (Figure 4.3-7a) or lesser (Figure 4.3-7b) extent, as noted for the volcanic samples from the Arqueros Formation (discussed above).

The identification of this mineral as pyrrhotite (rather than maghemite) is supported by the observation that samples displaying a larger peak at 250-350°C during high temperature susceptibility experiments, also display slightly higher coercivity behaviour during IRM acquisition and back-field coercivity experiments. In addition these samples are only saturated in fields of 500mT or more (Figure 4.3-7a), as opposed to the 300mT field required to saturate those samples with magnetite dominated mineral assemblages (Figure 4.3-7b).

Thermal demagnetisation of composite IRMs show that pyrrhotite rich samples display significant unblocking of the dominant intermediate and soft coercivity components at ~300-350°C, with the remaining remanence unblocked at 580°C. This is clearly evident in the two peaks shown in the unblocking spectra of Figure 4.3-7a. Whilst the demagnetisation composite IRMs applied to magnetite rich
Figure 4.3-7  Magnetic mineralogy experiments carried out on samples from the Quebrada de Los Choros dyke swarm. A) Samples from site AT242, B) Samples from site AT246.
samples also indicate that the intermediate coercivity component dominates, little of the overall remanence is destroyed between 250-350°C, but instead is mostly unblocked by treatment to 580°C, indicating the majority of the remanence is carried by reasonably high coercivity magnetite (Figure 4.3-7).

Demagnetisation Behaviour

The general demagnetisation behaviour of the samples essentially reflects the relative quantities of magnetite and pyrrhotite present, with magnetite dominated samples generally demagnetising using both AF and Thermal techniques (Figure 4.3-8a). Those remanences also carried by substantial amounts of pyrrhotite respond more effectively to thermal demagnetisation (with AF demagnetisation often resulting in the acquisition of what is interpreted to be a GRM), with a clear component of magnetisation unblocked between 150-350°C, before the remaining remanence unblocked between 350-580°C (Figure 4.3-8b).

All of the dyke samples demagnetised record a reasonably straightforward, univectoral component of magnetisation subsequent to the removal of a very low temperature/low coercivity component (Figure 4.3-8a & b). Where both pyrrhotite and magnetite contribute significantly to carrying the overall remanence (e.g., sample AT24001A-Figure 4.3-8b), the two minerals clearly carry an identical direction, suggesting they were magnetised contemporaneously. Assuming the magnetite to be a primary magnetic mineral, associated with the initial composition of the intuded dykes, the origin of the identical magnetisation direction carried by by magnetite and pyrrhotite, can be explained in two ways. Either the overall remanence is a Primary TRM, meaning that the pyrrhotite is also associated with the initial magma composition, or the dykes have been remagnetised during an event responsible for producing pyrrhotite as a secondary mineral.
Figure 4.3-7  Zijderveld plots illustrating the AF and Thermal demagnetisation behaviour of A) magnetite dominated and B) pyrrhotite (& magnetite) dominated dykes from the Quebrada de Los Choros dyke swarm.
Discussion of palaeomagnetic data from the Quebrada de Los Choros dyke swarm

It is common practice to correct the in-situ palaeomagnetic data observed from dykes by assuming they were intruded vertically and restoring the present dip to a palaeo-vertical initial orientation. All the dykes sampled in this area are sub-vertical in orientation (Table 4.3-3, Figure 4.3-6), with dips of 80°+. With such steep dips, it is difficult to distinguish whether or not the dykes were intruded either in this orientation prior to, or after the deformation which led to the low angle dips (10° to the E-SE) affecting the area. In either case the orientation of the dykes suggests very little deformation has occurred since emplacement, assuming they were emplaced subvertically in a tensional/transtensional stress regime.

Of the nine dykes sampled, seven yield directions which might be expected, one is unstable and one carries a northerly positive remanence which is not easily explained other than as an aberration of the field e.g. a remanence acquired during a reversal (Table 4.3.4). Of the seven site-mean directions considered further one site AT2-42, carries a reverse polarity direction. Although poorly constrained, this direction passes the reversal classification of McFadden & McElhinny (1990) with a classification of Cl (1 representing the fact that there is only one reverse polarity direction and therefore the reversal test is performed as an isolated observation test). This single reverse direction is therefore considered a genuine record of an anti-parallel (reverse polarity) field direction and included in the overall mean and the overall mean direction recorded by the Quebrada de Los Choros dyke swarm is considered to represent a Primary direction.

The in-situ mean direction recorded by the Quebrada Los Choros dyke swarm (D=017.9°, l=-47.2 with an a95 error of 5.2°), records a significantly more easterly declination (~15°) than that recorded by the tilt-corrected strata belonging to the Arqueros Formation (or Quebrada Marquesa Formation) into which the dykes
intrude (Figure 4.3-9, Tables 4.3-2 & 4.3-4).

Although the estimated age of the dykes is 80-90Ma (approximately 30-40 Ma younger than the intruded strata), the reference direction during this period indicates that the ancient field maintained a near-constant orientation (Besse & Courtillot, 2002, 2003). The in-situ direction recorded by the dyke swarm would therefore indicate that the dyke swarm records a significantly larger CW crustal rotation than the Arqueros Formation into which it intrudes, which is difficult to rationalise given the fact that the intruded strata are also interpreted to carry a Primary (pre-tilt) magnetisation.

As discussed previously, the subvertical nature of the dykes makes it difficult to establish whether the dykes were intruded prior to- or after the period of observed deformation in the Tres Cruces area. Assuming the intrusion of the dykes pre-dates the deformation, a tilt correction based on the gross orientation of the intruded strata (000/10 (strike & dip)) can be applied to the mean direction isolated from the dykes. The tilt correction produces a mean direction of D=006.9°, I=-49.4° (Table 4.3-4), which within error is identical to that recorded by the Arqueros Formation (Figure 4.3-9). The dyke swarm is therefore concluded also to record a pre-tilt, Primary TRM.
Figure 4.3-9  Stereonet projections of A) in situ low coercivity/low temperature components, B) in situ site mean ChRM directions and C) tilt corrected site mean ChRM directions from the Quebrada de Los Choros dyke swarm.
4.4 The Early to Mid-Cretaceous Magmatic Arc

The palaeomagnetic data from both Plutons Tilge and Los Colorados are considered in this section, as they form part of the Early to mid Cretaceous magmatic arc. Only a very small number of sites were collected from pluton Tilge (due to drill failure in the field) and therefore only a limited interpretation of this data is possible.

4.4.1 Pluton Tilge

The pluton crops out over an area of $13\text{km}^2$, south of the town of Chungungo, approximately 40km west of the Los Choros sampling area (Figure 4.1-2), and four sites were collected from river cut sections. (Table 4.4-1, Figure 4.4-1). The overall pluton comprises several phases of granitic-dioritic magmatic activity, intruding an older Jurassic sub-volcanic/plutonic complex. Isotopic dates include K-Ar in biotite and whole rock & U-Pb in zircon and indicate the age of the intrusion to be between 121-130Ma [Emparan & Pineda, 2000].

*Magnetic mineralogy*

The majority of samples from the grey granodioritic groundmass of the main body of the pluton, proved to be dominated by near pure magnetite. $\chi$-$T$ experiments exhibit extremely sharply defined Curie temperature at $580^\circ\text{C}$, with near reversible heating curves except for the presence of an apparently ubiquitous peak at $\approx 300^\circ\text{C}$ (Figure 4.4-2a). This could either be due small quantities of pyrrhotite or maghemite.

IRM acquisition experiments clearly demonstrate that samples are saturated in applied magnetic fields of 200 mT (Figure 4.4-2a). This suggests that the peak noted during the heating phase of high temperature susceptibility experiments is
Table 4.4-1  Site localities, lithology and summary of demagnetisation behaviour of samples from the early Cretaceous Pluton Tilgo.

<table>
<thead>
<tr>
<th>Site</th>
<th>Sampling Site Lat (°S) Long (°E)</th>
<th>Lithology</th>
<th>Tilt Correction</th>
<th>°C Min.</th>
<th>ChRM Isolation Max.</th>
<th>mT Min.</th>
<th>mT Max.</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT2-48</td>
<td>29.45 288.73</td>
<td>Granite</td>
<td>-</td>
<td>150</td>
<td>580</td>
<td>2-8</td>
<td>35-50</td>
</tr>
<tr>
<td>AT2-49</td>
<td>29.45 288.72</td>
<td>Granite</td>
<td>-</td>
<td>100</td>
<td>250</td>
<td>4-10</td>
<td>15-40</td>
</tr>
<tr>
<td>AT2-50</td>
<td>29.45 288.72</td>
<td>Granite</td>
<td>-</td>
<td>100</td>
<td>400</td>
<td>4-6</td>
<td>25-70</td>
</tr>
<tr>
<td>AT2-51</td>
<td>29.45 288.71</td>
<td>Granite</td>
<td>-</td>
<td>100</td>
<td>400</td>
<td>NRM-4</td>
<td>20-30</td>
</tr>
</tbody>
</table>

Table 4.4-2  Individual site mean directions and overall mean direction calculated for the early Cretaceous Pluton Tilgo. The very large error associated with the mean direction calculated from site AT2-51 was not used in the calculation of the overall pluton mean direction, even though the actual direction was similar to that calculated for the remaining three sites.

Figure 4.4-1  Location of sampling localities within the early Cretaceous Pluton Tilgo—see Figure 4.1-2 for an overview of the Tres Cruces area.

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Figure 4.4-2  Magnetic mineralogy of samples from the early Cretaceous Pluton Tilgo. A) Grey facies Sample AT251-03C; B) Pink facies Sample AT248-02A. Symbols as per standard description.
likely to reflect the presence of maghemite, as pyrrhotite would tend to increase the observed coercivity of the sample (see Appendix B-magnetic mineralogy). The presence of maghemite is possibly related to the (deuteric) oxidation of magnetite late during emplacement, or developed subsequently through weathering or low temperature alteration for example.

Multicomponent IRM experiments show that the soft component of magnetisation dominates, although a significant intermediate component is also noted (Figure 4.4-2a). Both soft and intermediate components are progressively destroyed between 100-560°C with a small component of magnetisation removed at 250°C, again with no clear evidence of a significant contribution from pyrrhotite. These observations indicate that magnetite is the dominant (primary) magnetic mineral, with a small amount of maghemite also present.

Rock magnetic experiments carried out on samples from site AT2-48 (collected from a less common pinkish brown granite) indicate that a small amount of haematite may also be present within samples belonging to this granitic facies—as suggested by the reddened colour. χ-T experiments identified magnetite and maghemite as being dominant as elsewhere, but Curie temperatures in excess of 580°C were also observed (Figure 4.4-2b). IRM acquisition experiments suggest that the samples become fully saturated after application of fields between 200-300mT (slightly higher than for the grey samples), indicating that the amount and overall contribution of hematite is probably very small. This is supported by the observation that the contribution of the hard (800mT) component of a composite IRM is negligible (Figure 4.4-2b), whilst the soft and intermediate components are equal in magnitude and destroyed on treatment to 580°C. This indicates that magnetite is also the primary carrier of magnetisation for these (reddened) samples.
Demagnetisation behaviour

The initial NRM intensity of all samples was much lower in general than observed for samples from many of the other plutons sampled during this study. NRM intensities of up to 300 mA/m, but generally less than 100 mA/m are observed for most samples.

The majority of samples from Pluton Tilgo were more effectively demagnetised using AF than thermal demagnetisation, which often resulted in significantly more "noisy" demagnetisation plots. The ChRM of all of the samples is represented by a single generally origin bound component of magnetisation, often isolated subsequent to a very small low temperature/low coercivity component. Thermal demagnetisation of sample AT248-01 indicates that the ChRM is progressively destroyed by treatment between 150-580°C, although the bulk of magnetisation is unblocked between 540-580°C (Figure 4.4-3a). This suggests that ti-poor magnetite is the dominant magnetic carrier in these samples. The unblocked remanence at intermediate temperatures may be carried by maghemite or magnetite grains of larger size/greater proportions of titanium, but carries a similar direction to the ti-poor magnetite component and is therefore considered contemporaneous.

Thermal demagnetisation of samples from the remaining sites indicates that the entire remanence is carried solely by maghemite, being completely destroyed by treatment to between 250-400°C. This magnetisation is considered to be contemporaneous to that isolated from site AT2-48, due to the relationship between magnetite and maghemite observed for samples from that site and is interpreted to reflect the result of more complete oxidation of magnetite. No high temperature (>580°C) component of magnetisation was observed in any of the samples thermally demagnetised.
Figure 4.4-3 Examples of the typical Thermal (A) and AF (B) demagnetisation behaviour displayed by samples from the early Cretaceous Pluton Tilgo.
AF demagnetisation of samples from Pluton Tilgo indicates that the ChRM is of relatively low coercivity (indicated by the concave intensity plot in Figure 4.4-3b) and is consistent with the observation that magnetite and/or maghemite are the primary carriers of magnetisation. The ChRM is progressively destroyed in demagnetising fields of up to 30-70mT (Table 4.4-1) and the lower coercivity components may reflect the dominance of maghemite.

Discussion of palaeomagnetic data from Pluton Tilgo

The low-intermediate coercivity/temperature components of magnetisation isolated from the majority of samples are observed to be randomly dispersed and probably therefore represent a small short term induced VRM, acquired either during sample collection or during storage (Figure 4.4-4a). Three of the four sites produced well-constrained mean ChRM directions (Table 4.4-2 and Figure 4.4-4B) whose declinations are rotated CW from the present day field direction, indicating that an ancient remanence has been isolated. The fourth site (AT2-51) yields a poorly constrained direction (Table 4.4-2) and is therefore excluded from further consideration.

Although the three directions included in the overall mean direction have similar inclinations, the declinations are spread across an arc of 20° on the stereonet. This is quite conceivably an effect of under-sampling, leading to the insufficient averaging of the effects of PSV associated with the limited number of sites. There is however no reason to suggest that these are not genuine records of the ancient magnetic field direction. The overall mean direction calculated for Pluton Tilgo is \(D=018.6^\circ, I=-43.4^\circ\), with an \(\alpha_{95}\) error of 11.5° (Table 4.4-2) and will be considered a valid direction unless comparison with the remaining palaeomagnetic data from the Tres Cruces area suggests otherwise.
Figure 4.4-4  Stereonet plots of A) low and intermediate coercivity temperature components of magnetisation (not ChRM); B) individual site mean directions and overall mean direction calculated for the early Cretaceous Pluton Tilgo.
4.4.2 Pluton Los Colorados

Pluton Los Colorados forms the most westerly part of the early to mid-Cretaceous magmatic arc and is located approximately 70km NE of La Serena (~29° 30' S), at the boundary of the Coastal Cordillera/Pre-Cordillera (Figure 4.1-3). The outcrop covers approximately 8 x 20km (Moscoso et al, 1982), with the pluton long axis oriented ~N-S, reflecting the overall trend of the margin at this latitude. A single K-Ar date carried out on an amphibole fraction from a sample from the southernmost tip of the pluton produced an age of 96 ± 4Ma [Emparan & Pineda, 1999].

The pluton is dominantly grey to pink quartz-rich monzodiorite with amphibole and biotite or pyroxene [Emparan & Pineda, 1999] and in places there is obvious evidence of mixing of coeval magmas, with distinct contacts visible-differentiated primarily by colour (Figure 4.4-5a & b). Also evident in places within the pluton is the presence of enclaves/xenoliths of darker fine-grained material (Figure 4.4-5c & d) that are quite angular in appearance. These enclaves are generally observed to occur in the areas of the pluton that are of leucocratic, more granitic composition.

Palaeomagnetic core samples were collected from 10 sites along Quebrada Los Choros across the northern end of the pluton (Table 4.4-3, Figure 4.4-6). Eight of the ten sites were sampled from the main (monzodiorite) body of the pluton and two cross-cutting (basaltic/andesitic) dykes. Both dykes were essentially vertical, tentatively suggesting little post-emplacement tilting has affected the pluton since the dykes were intruded (assuming a near vertical original orientation at emplacement).
Figure 4.4-5  Field photographs illustrating some of the visible features of the mid-Cretaceous Pluton Los Colorados at outcrop scale (hammer for scale). A & B-Contact between pink and grey granitic magmas. C & D-Fine grained (and more mafic?) angular enclaves observed within the grey granitic unit.
Table 4.4-3 Site localities, lithology and summary of demagnetisation behaviour of samples from the mid-Cretaceous Pluton Los Colorados.

Table 4.4-4 Individual site mean directions and overall mean direction calculated for the mid-Cretaceous Pluton Los Colorados. The mean direction calculated for samples from site AT2-07 (shaded grey) is not used in the calculation of the overall pluton mean direction, due to the unacceptably high error (>>20°) associated.
Granitic groundmass of Pluton Los Colorados

Magnetic mineralogy

Rock magnetic data from site AT202 typifies the magnetic mineralogy as seen in all the granodioritic samples from Pluton Los Colorados (Figure 4.4-7a). Thermomagnetic experiments indicate that the dominant magnetic mineral present within the samples is magnetite, with sharply defined Curie temperatures indicated at 580°C (Figure 4.4-7a). The powdered samples underwent some slight thermochemical alteration during heating, with a slight increase in susceptibility between 300-350°C, which is absent on cooling and interpreted as the alteration of maghemite to haematite. Apart from this slight alteration, the thermo-magnetic curves appear to be effectively reversible.

The contribution of magnetite (and maghemite) is confirmed by IRM acquisition experiments, which clearly indicate that the samples acquire magnetisation extremely rapidly in fields of up to 100 mT before effectively saturating in fields of ~300mT (Figure 4.4-7a). The acquisition of magnetisation however is often observed to be quite noisy as the highest fields are applied, as is evident in Figure 4.4-7a. This is ascribed to instrument error associated with the measurement of the very high intensities acquired such that the induced magnetisations of some samples saturated the spinner magnetometers used. All of the samples indicate coercive fields of 30-60 mT are required to reverse the polarity of the acquired IRM during back-field coercivity experiments consistent with lower coercivity minerals dominating the observed magnetisation.

Thermal demagnetisation of a composite IRM indicates that that a soft (50mT) component of magnetisation, is unblocked between 100-350°C with the similar removal of an intermediate (300mT) component also noted (Figure 4.4-7a). This
Figure 4.4-7  Magnetic mineralogy of samples from the mid-Cretaceous Pluton Los Colorados. 
A) Granitic samples from site AT202 (representative of both grey and pink facies). 
B) Samples from the two dykes sampled from within the main body of the pluton.
suggests that a reasonable amount of maghemite is present and contributes significantly to the overall IRM. Above 350°C, both the soft and intermediate components are destroyed equally up to 580°C indicating the presence of magnetite. There is a complete lack of any hard component of magnetisation, suggesting that there is little or no haematite present.

The only slight differences noted between samples of the pink coloured granite and those from the pale grey granite is a less sharply defined Curie temperature for magnetite during k-T experiments, as well as evidence for the limited presence of some haematite as per Curie temperatures >580°C. IRM acquisition experiments confirm magnetite/maghemite as the dominant magnetic minerals (magnetisation rapidly acquired up to 100 mT), with very little evidence of a significant haematite contribution to the overall IRM (i.e. samples fully saturate in the maximum applied fields). The small haematite content is thought to be associated with the initial composition variation of the magma and not as a secondary mineral as such. As a consequence of this assumption therefore both magnetite and haematite (if present) are primary minerals and should record the same direction of magnetisation if the pluton records a primary TRM.

Demagnetisation behaviour

Initial NRM intensities range between 0.02-1.00 A/m, with the average intensity being <0.2 A/m. Those samples with Total NRM intensities >0.1 A/m, generally had a clearer demagnetisation trajectory than weaker samples. After removal of a low-intermediate coercivity/temperature component of magnetisation, the majority of samples simply displayed a single and generally origin-bound vector of magnetisation. This behaviour is observed when either normal or reverse polarity magnetisations are recorded, but the two polarities appear to be generally
associated with slightly different magnetic mineral assemblages and hence two
demagnetisation behaviours are identified.

The most common behaviour, termed here the A-type, is characterised by a
relatively weak resistance to AF demagnetisation, with the majority of NRM
destroyed on application of 30-40 mT (Figure 4.4-8a). The ChRM isolated from
some samples is carried by an even lower coercivity carrier mineral, where the
majority of NRM was destroyed after treatment to 15-20 mT and the direction of
magnetisation becomes unstable in fields >30-40 mT. This low coercivity
behaviour indicates that the ChRM recorded by the majority of the samples, is
carried either by magnetite of either large grain size, or ti-rich composition, and/or
maghemit (as suggested by rock magnetic studies).

Thermal demagnetisation of the A-type samples most commonly shows
substantial unblocking in the range 200-400°C (following the removal of a low
temperature random component between 100-200°C), with the intensity often
reduced to ~25-50% of the initial NRM intensity at temperatures of 400°C (Figure
4.4-8a). The remaining component of magnetisation is fully unblocked between
520-580°C, suggesting that (titanom-) magnetite is the principle carrier of this part of
the magnetisation. Thermal demagnetisation of the generalised A-type behaviour,
shows that only a single direction of magnetisation is observed across the
temperature range 200-580°C, implying that the multiple carriers of NRM observed
(maghemite and (titanom-) magnetite), acquired their magnetisation
contemporaneously.

The B-type behaviour is generally only observed in sites sampled from the pink
coloured granite, and then generally (but not exclusively) only in those samples
displaying reverse polarity magnetisations (Figure 4.4-8b). Similar to the A-type
magnetisation, the B-type behaviour is characterised by a single (origin bound)
Figure 4.4-8 Examples of A) 'A-type' (magnetite only) and B) 'B-type' (magnetite & haematite) demagnetisation behaviour (Thermal & AF) displayed by granitic samples from the mid-Cretaceous Pluton Los Colorados. C) Thermal and AF demagnetisation behaviour representative of samples from the two dykes sampled.
direction of remanent magnetisation, but the actual magnetisation is observed to be more resistant to AF demagnetisation. This component is directionally stable (Figures 4.4.8b) and progressively destroyed in fields up to ~70 mT. Thermal demagnetisation of this B-type magnetisation shows that after the removal of a low temperature component (~150°C), the magnetisation is evenly unblocked up to 670°C, with samples displaying only a single magnetisation vector, indicating both magnetite and haematite as a carriers of NRM (Figures 4.4-8b). The fact that the direction of remanent magnetisation carried by the haematite fraction is indistinguishable from that carried by magnetite, again suggests that the two minerals acquired magnetisation at the same point in time, interpreted to be during cooling from above 670°C.

*Pluton Los Colorados dykes*

*Magnetic mineralogy*

Rock magnetic experiments on samples from the two dykes collected from Pluton Los Colorados indicate that maghemite and magnetite are the dominant magnetic minerals (Figures 4.4-7b). The sample from site AT2-06 displays a large drop in susceptibility at ~400°C suggesting that maghemite (rather than pyrrhotite) dominates the magnetic mineral assemblage. The much-reduced susceptibility evident on cooling, shows that exceptional amounts of thermo-chemical alteration occur during the heating cycle, consistent with the production of a significantly lower susceptibility mineral, probably due to conversion of maghemite to haematite. Other dyke samples subjected to k-T experiments displayed more well defined Curie temperatures at 580°C (AT20303B-Figure 4.4-7b), indicating that magnetite is more dominant in these samples, with a far smaller contribution from maghemite.
IRM acquisition curves, although often quite noisy at higher field strengths, show that the samples saturate in magnetic fields of ~100 mT, similar to the behaviour expected for magnetite, but consistent with presence of maghemite (figures 4.4-7b). Thermal demagnetisation of composite IRMs indicates that the induced remanence is almost completely destroyed by treatment to 400°C (Figure 4.4-7b) for samples from site AT206, although a greater magnetite contribution is noted for samples from site AT203 (Figure 4.4-7b). The amount of maghemite interpreted to reflect the state of oxidation of the primary magnetite within the dyke samples, with the greatest oxidation represented by the increased dominance of maghemite.

**Demagnetisation behaviour**

The initial NRM intensities of samples from dyke AT203 (magnetite dominant) are three times larger than those exhibited by samples from dyke AT206 (maghemite dominant) and the stability of the isolated ChRM to demagnetisation reflects the established magnetic mineralogy. Samples from site AT2-03 demagnetised fully using either Thermal or AF techniques, with the remanent magnetisation completely destroyed on treatment to 540-560°C or 50-70 mT (Figure 4.4.8b). In contrast, samples from site AT2-06 displayed a significantly less stable, lower temperature/lower coercivity ChRM, which is destroyed after treatment to 400°C or ~40mT consistent with the majority of NRM being carried by maghemite.

As noted for samples from the main body of the pluton, all samples from both dykes carry only a single component of magnetisation, barring the presence of small low coercivity/low temperature components.

**Discussion of palaeomagnetic data from Pluton Los Colorados**

Most samples display a low coercivity/low temperature component that is generally removed by treatment to 2-5 mT or 100-200°C. These low components can
account for up 50%+ of the initial NRM intensity and their distribution suggests they were acquired as a short term induced VRM (Figure 4.4-9a).

Of the ten sites sampled from Pluton Los Colorados, six (four) yield normal (reverse) CHRM directions, although the direction calculated from site AT207 is rejected due to large a95 error (Table 4.4-4 and Figure 4.4-9b). It is clear that there is no significant difference between the site mean directions obtained from the two dykes and the remaining sites collected from the main body of the pluton, hence they are included in further analysis with the data from the main body of the pluton.

Application of a reversal test [McFadden and McElhinny, 1990], using the nine site mean directions results in a classification of C, indicating that the samples are likely to belong to populations with means that are 180° apart, and suggesting that the isolated remanence probably does represent an ancient record of the Earths' ancient magnetic field. The reasonable amount of scatter between the site mean directions also suggests that the pluton cooled during a reasonably long period of time and therefore the effects of PSV can reasonably be assumed to have been averaged (Figure 4.4-9b). The overall mean direction calculated for Pluton Los Colorados is therefore considered to represent an ancient TRM, with D=003.9°, I=-55.8°, with an a95 error of 7.8° (Table 4.4.4, Figure 4.4.9).
Figure 4.4-9 Stereonet plots of A) low to intermediate coercivity/temperature components of magnetisation isolated from all samples. Where these components are identified over more than two demagnetisation steps (including NRM), the associated error ellipse is indicated. B) Individual site mean directions and overall mean direction calculated for the mid-Cretaceous Pluton Los Colorados.
4.5 The latest Cretaceous-earliest Paleocene Magmatic Arc

The latest Cretaceous-earliest Paleocene magmatic arc intrudes the mid-late Cretaceous (Aptian-Cenomanian) volcanic, volcaniclastic and sedimentary sequences of the Cerrillos Formation. The latter was deposited in a continental arc-/back-arc basin [Marschik and Fontboté, 2001a], immediately east of the Bandurrias/Chanarcillo basin (Figure 4.1-3). Two plutons, Las Campañas and Corredores, were sampled in the Tres Cruces field area (Figure 4.1-3). Access to the interior of both plutons was, relatively, easy and sites were sampled from the interiors and margins to test for any positional variation e.g. alteration at the margins, slow cooling etc.

Ar$^{40}$-Ar$^{39}$ geochronology using hornblende separates from several margin parallel plutons belonging to this magmatic arc, a short distance to the north of Tres Cruces in the Vallenar area (c.28°30'S), produced ages of 65.4 Ma +/- 0.6 Ma (Pie de Gallo) and 66.8 Ma +/- 0.6 Ma (Chehueque) [Gipson, unpublished data]. Ar$^{40}$-Ar$^{39}$ ages on hornblende (71.2 ± 0.3 Ma) and biotite (71.0 ± 0.9 Ma) from the marginal shear zone and a biotite age (70.2 ± 0.9 Ma) from the interior of the pluton demonstrate rapid cooling and are, within reason, consistent with the ages from other plutons to the north (Truelove et al. 2005).

To the south in the Condoriaco-Rivadavia area, south of Quebrada Los Choros, the southern part of Pluton Corredores cuts and intrudes the north-eastern segment of the large collapse Caldera Condoriaco (c.85-72Ma-Figure 4.1-3). A suite of four K-Ar dates from the granodioritic intrusions in this area range between 71-68 Ma [Emparan & Pineda, 1999], which is compatible with those to the north. Once again, the locus of the magmatic arc responsible for the emplacement of this plutonic suite had shifted eastwards with respect to the 96-93 Ma plutonic event (which included the previously discussed Pluton Los Colorados).
The gross orientation of pluton long axes at the latitude of Pluton Corredores (29°30' S) is approximately N-S, as is also observed for plutons belonging to earlier magmatic arcs (e.g. Plutons Tilgo and Los Colorados-Figure 4.5-1 [Emparan & Pineda, 1999]). There appears to be a noticeable change in this direction of pluton elongation at the latitude of Pluton Las Campañas (c.29°S), to a more NNE-SSW orientation, which is also reflected in the general strike of the intruded strata. A further change in trend to a more NE-SW direction is noted further to the north at ~28°S. It was intended that the sampling of the two plutons with differently trending long axes would allow the investigation of the relationship between the observed regional trend of the margin and the magnitude of observed crustal rotation.

4.5.1 Pluton Las Campañas

At the latitude of Pluton Las Campañas (29° S) and to the north-east, the latest Cretaceous-earliest Paleocene magmatic arc lies within or immediately east of the Chañarcillo Fold and Thrust Belt (CFTB). The Early Cretaceous marine sediments of the Chañarcillo Group are separated from the continental volcanic deposits of the Cerrillos Formation by a distinct structural break, which in part appears to have been re-exploited during the emplacement of many of the plutons belonging to the latest Cretaceous-earliest Paleocene magmatic arc. This situation is clearly mapped as being the case for Pluton Las Campañas (Moscoso et al., 1982, Figure 4.1-3), with the western boundary of the Pluton in apparent thrust contact with the Chañarcillo Group, although recent remapping of the area [Truelove et al, 2003], suggests that the fault zone is a reactivated normal (basinBounding/growth?) fault.

Plutonic complexes of latest Cretaceous age, in the region to the north, are often bounded to the west by steeply dipping, syn-plutonic faults and ductile shear
Figure 4.5-1 Extent and location of the (easterly younging) magmatic arcs intruding the present day forearc region in northern Chile (between 27-30°S). The plutons sampled in the Tres Cruces field area are circled, with arrows indicating the orientation of pluton long axes. See text for discussion.
zones (Grocott & Taylor, 2002). They suggest that such plutons were emplaced through roof uplift/floor subsidence mechanisms, exploiting pre-existing steeply dipping faults that were reactivated during transtension. In the case of Pluton Las Campañas Truelove et al., (2005) suggest a combination of both roof uplift and floor subsidence for the emplacement mechanism due to the parallelism of the contacts and regional dip of the host/roof rock of the Cerrillos Formation (Figure 4.5-2).

Ten sites (AT2 22-31) were collected from Pluton Las Campañas (Table 4.5-1, Figure 4.5-3) where excellent access to the interior of the pluton was afforded along a road leading to the Las Campañas (astronomical) Observatory situated on top of Cerro Las Campañas in the east. The presence of a sub-horizontal basaltic sill observed at site AT2-23 in pluton Las Campañas, may suggest that there has been little if any tilting, assuming a near horizontal intrusion originally. Other than this sill, there were no directly measurable surfaces evident within the pluton that would serve as reliable palaeohorizontal indicators and as a consequence all palaeomagnetic data will be discussed in geographic (in situ) coordinates.

In some sampling sites, up to 60% of the outcrop was composed of melanocratic enclaves of finer grained, darker and apparently more dioritic material, with one site (AT2-28) also displaying fine grained angular xenoliths of volcanic appearance, up to 10cm in size. Approximately 50% of the samples collected were from these enclaves when they were suitably large enough to be collected for palaeomagnetic analysis, with the remaining samples taken from the more leucocratic and coarser, surrounding groundmass. At site AT2-24 the enclaves were accompanied by a stratified and distinct horizon (sill like) differentiated layering. The presence of more melanocratic enclaves and magma stratification is interpreted as resulting from the intrusion of two coeval (but immiscible) magmas.
The altered country rock around the observatory dips conformably to the regional dip - potential pluton roof?

Apparent roll over of stratigraphy to the Las Campanas Pluton, similar to 'roll over' structures associated with roof uplift / floor depression emplacement models

Floor depression emplacement model (Grocott & Taylor 2002)

The current preliminary model suggests a component of both floor depression & roof uplift due to the dipping roof

Figure 4.5-2 Pluton emplacement mechanism suggesting a combination of roof uplift and floor subsidence to explain the intrusion of Pluton Las Campanas (Truelove et al., 2005).

<table>
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<tr>
<th>Site</th>
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<th>Lat (°S) Long (°E)</th>
<th>Lithology</th>
<th>Tilt Correction</th>
<th>°C Min.</th>
<th>°C Max.</th>
<th>ChRM Isolation</th>
<th>mT Min.</th>
<th>mT Max.</th>
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<td>560-580</td>
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<td>70</td>
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<tr>
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<td>289.25</td>
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<td>400</td>
<td>560</td>
<td>15-35</td>
<td>70-100</td>
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</tr>
<tr>
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<td>No ChRM Recovered</td>
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<td>Enclave rich Granodiorite</td>
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<td>540-580</td>
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<td>520</td>
<td>8-20</td>
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Table 4.5-1 Site location, lithology and summary of sample demagnetisation behaviour from the latest Cretaceous-earliest Paleocene Pluton Las Campanas.
Figure 4.5-3  Map showing site localities within Pluton Las Campañas—see Figure 4.1-2 for an overview of the Tres Cruces area.

Table 4.5-2  Site mean ChRM directions from the latest Cretaceous-earliest Paleocene Pluton Las Campañas.
and hence both were sampled in order to investigate possible mineralogical/palaeomagnetic differences between the layering, which might have affected the acquisition of NRM and/or the palaeomagnetic behaviour within the pluton.

**Magnetic Mineralogy**

The rock magnetic experimental data indicates that the magnetic mineralogy of the plutons is comparatively simple being, essentially, nearly pure/low Ti magnetite. K-T experiments (Figure 4.5-4a) show an extremely well defined Curie temperature at 560-580°C, indicating the presence of pure or nearly pure magnetite, with the cooling curve observed to be almost identical to the heating curve, i.e. reversible. This suggests that the samples have undergone very little thermochemical alteration during heating, and that magnetite is the dominant carrier of magnetisation within the samples. There is no evidence of higher temperature fractions being present at all in these samples.

IRM acquisition and back-field coercivity experiments (Figure 4.5-4a), indicate that the samples are completely saturated in magnetic fields of between 200-500 mT, with rapid acquisition of magnetisation in fields up to 100 mT. Subtle variations in the acquisition of magnetisation in fields between 100-500 mT may indicate fluctuation in the proportion of SD (and/or PSD) relative to MD sized magnetite particles. Applied fields of between 20-60mT (30mT average) are required during back-field coercivity experiments to destroy the initial IRM, again indicating that magnetite is the only significant carrier present.

Thermal demagnetisation of a composite IRM indicates that the soft (50mT) and intermediate (300mT) components dominate the overall IRM, with no hard (800mT) component of magnetisation observed. Both the soft and intermediate
Figure 4.5-4 Typical results of rock magnetic experiments from the latest Cretaceous-earliest Paleocene Plutons Las Campañas (A) and Corredores (B).

Figure 4.5-5 Variation of NRM intensity and Bulk Susceptibility between samples from the latest Cretaceous-earliest Paleocene Plutons Las Campañas and Corredores. Enclave samples circled. No significant differences between plutons but greater spread of Las Campañas samples noted.
components are progressively demagnetised until they are completely unblocked at 580°C (i.e. Tc for magnetite), but the more pronounced demagnetisation of the soft component at temperatures of ~350-400°C may reflect that a small proportion of the magnetisation is carried by a lower temperature mineral such as maghemite (Figure 4.5-4a). In short, pure or near pure magnetite dominates in all these samples.

Rock magnetic experiments indicated no discernable (magnetic) mineralogical difference between samples collected from melanocratic enclaves and the leucocratic groundmass of the plutons (nor indeed between samples from the two plutons). Figure 4.5-5 illustrates the relationship between initial NRM intensity and sample susceptibility and this confirms that there is also no consistent difference in the magnetic properties of the enclaves and pluton groundmass, or between the plutons, but greater within site variation is noted for samples from Pluton Las Campañas.

It is notable that the data of Figure 4.5-5 show some samples with exceptionally high total NRM intensities with respect to their otherwise normal susceptibilities and this occurs in three particular sites from Las Campañas. The magnetisation of these sites (and a few other sporadic samples) was heavily overprinted although not always totally, by large magnitude, univectorial magnetisations of up to ~14 A/m which display variable within site directions (Figure 4.5-6). The example data from site AT2-27 show the rapid reduction in the intensity of the magnetisation with applied AF field and marked differences in the direction recorded depending on where they were collected from the outcrop. The overall behaviour is consistent with this exceptionally large, low coercivity component having been acquired as a result of lightning strike [e.g. Butler, 1992]. Given the high, open aspect of the
Figure 4.5-6 Examples of the demagnetisation of samples from Site AT2-27 (Pluton Las Campañas) that display anomalously large initial NRM intensities and unusually orientated magnetisation directions interpreted as lightning strike induced magnetisations.
sampling area this was not completely unexpected. Two of the three sites thus affected failed to produce any discernible/reliable ChRM.

Demagnetisation Behaviour

Only a limited range of demagnetisation behaviours are observed during Thermal and AF treatment of samples from Pluton Las Campanas (Figure 4.5-7). Most commonly, subsequent to the removal of a low coercivity/low temperature component of magnetisation, only a single, univectorial (origin bound) component of magnetisation is observed (Figure 4.5-7a). This component of magnetisation is observed to be reasonably evenly demagnetised between temperatures of 100°C (-250°C) and 580°C, or AF fields of between 10-70mT, suggesting that Ti-poor magnetite is the principle carrier of magnetisation.

Discussion of palaeomagnetic data

Low coercivity/temperature (<20mT or <200°C) component directions isolated during demagnetisation, but excluding those from lightning strike sites previously discussed, are widely scattered, with limited clustering about the present day direction (Figures 4.5-8a). The generally random scattering noted for these components is therefore interpreted to be representative of minor short term VRMs.

Magnetisation directions recovered from individual samples are noted to often be relatively inconsistent within many of the sampling localities, consequently resulting in the calculation of poorly constrained site mean ChRM directions, although overall the site mean directions are reasonably consistent (Table 4.5-2, Figure 4.5-8B). It is clear that the majority of directions determined were acquired during a period of reverse magnetic polarity (Figure 4.5-8, Table 4.5-2) and therefore must represent ancient records of the Earth's magnetic field. Given the
Figure 4.5-7 Examples of the typical demagnetisation behaviour (AF & Thermal) of samples from Pluton Las Campañas. A) A-type behaviour (origin-bound & univectorial). B) B-type behaviour (multi-component, GC).
Figure 4.5-8  A) Low-intermediate coercivity/temperature components of magnetisation that are not ChRM from individual samples from Pluton Las Campañas. Where these components are identified over more than two demagnetisation steps (including NRM), the associated error ellipse is indicated. B) Individual site mean directions and overall mean direction (in situ) from Pluton Las Campañas.
fact that only a single normal polarity site is observed, the lack of evidence for low
temperature alteration products, geochronology suggesting rapid cooling and the
lack of any direct evidence for post-emplacement large scale tilting it is therefore
assumed that the remanence is primary and was acquired at the time of
emplacement in the latest Cretaceous – earlist Palaeocene (~70Ma). The overall
mean direction, is $D=203.9^\circ$, $I=46.1^\circ$ with an associated $\alpha_{95}$ error of 13.3° (Table
4.5-2, Figure 4.5-8).

4.5.2 Pluton Corredores

Pluton Corredores was sampled towards the eastern end of Quebrada Los Choros
where it intrudes the Cerrillos Formation (Figure 4.1-2). Seven sites of 6-8
samples were collected from between 5-10 km east of the c.100 Ma Pluton Los
Colorados, and ~50 km to the south of Pluton Las Campañas (Table 4.5-3, Figure
4.5-9). Access to the interior of the pluton was possible along the Los Choros
valley, which cuts E-W across the approximately N-S striking long axis of the
pluton (Figure 4.1-2). As one moves eastward the valley becomes increasingly
tightly confined/steep-sided and sampling was limited to outcrops of at valley floor
level.

Visibly the granodiorite is very similar in appearance to that previously described
for pluton Las Campañas i.e. it is typified by enclaves (normally 5-40cm across) of
melanocratic, finer grained material, set within a more leucocratic (coarser
grained) granodioritic groundmass. As with the sample collection at pluton Las
Campañas, where possible, samples were taken equally from both the enclaves
and the groundmass, to test any mineralogical differences that may have affected
the acquisition of magnetisation within the pluton.
Table 4.5-3 Site location, lithology and summary of sample demagnetisation behaviour from the latest Cretaceous-earliest Paleocene Pluton Corredores.

Table 4.5-4 Site mean ChRM directions from the latest Cretaceous-earliest Paleocene Pluton Corredores.

Figure 4.5-9 Map showing site localities within Pluton Corredores—see Figure 4.1-2 for an overview of the Tres Cruces area.
Magnetic mineralogy

Rock magnetic experiments were similar to pluton Las Campanas (Figure 4.5-3B) and hence the same conclusion is reached that the magnetic carriers are predominantly low Ti - pure magnetite. In some samples however a much larger fraction of the magnetisation was carried by maghemite.

Demagnetisation behaviour

The ChRM direction progressively unblocked generally between 100-200°C and 560-580°C (Table 4.5-3) and over a wide range of coercivities. Demagnetisation behaviour is consistent with the rock magnetic behaviour in suggesting titanomagnetite to be the principle carrier of the NRM, although the variation in behaviour may suggest that there is a wider range of Ti substitution and/or grain size in comparison to pluton Las Campañas.

The dominant demagnetisation behaviour displayed is characterised by the progressive destruction of a single linear and generally origin bound, direction of magnetisation (following removal of a low coercivity/low temperature secondary VRM component as discussed previously). This is referred to as the ‘A-type’ behaviour. This relatively simple magnetisation is clearly demagnetised by both AF and Thermal methods, and is evident in samples displaying both normal and reverse polarity remanent magnetisation directions (Figure 4.5-10a).

Thermal demagnetisation of the A-type remanence reveals two forms in terms of behaviour—one in which the unblocking occurs in the upper end of the spectrum and one where it is more distributed with significant loss of intensity between ~350-400°C suggesting a significant amount of maghemite was present. Thermal demagnetisation of several samples displaying the univectorial A-type magnetisation indicates that the overall ChRM is carried predominantly by a higher
Figure 4.5-10  Examples of the typical demagnetisation behaviour (AF & Thermal) of samples from Pluton Corredores (as described for Pluton Las Campañás). A) A-type behaviour. B) B-type behaviour.
temperature magnetic mineral, with the majority unblocked between 500-580°C (Figure 4.5-10a). This indicates that ChRM, in these samples, is almost entirely carried by (titano-) magnetite, with only a small contribution from maghemite.

Many of the samples did not demagnetise fully to reveal a ChRM direction with a stable end-point, but instead the demagnetisation path followed a great circle path, suggesting the presence of two components of remnant magnetisation with overlapping demagnetisation spectra. This behaviour is referred to as the 'B-type' magnetisation and samples can become unstable at low to very low AF demagnetisation field (Figure 4.5-10b). Only reverse polarity directions/great circle segments can be resolved/constrained with this type of behaviour as it is impossible to distinguish an ancient normal polarity from the present day (unrotated) direction. The magnetic carrier minerals in these samples are unable to sustain stable demagnetisation trends above 50mT or 400°C, suggesting maghemite is the dominant carrier.

**Discussion of palaeomagnetic data**

Demagnetisation of samples from Pluton Corredores demonstrated the removal of a low coercivity/low temperature component of magnetisation in the majority of cases. Whilst there appears to some slight clustering about the present day field direction, the majority of the directions are scattered suggesting that they represent a short term induced VRM (Figure 4.5-11).

In comparison to Pluton Las Campañas, lightning induced IRMs were rare in samples from Pluton Corredores, but several samples did display this behaviour. The ChRM directions include three reverse and four normal polarity directions, with those of reverse polarity notably more precisely defined than those of normal polarity (Table 4.5.4 and Figure 4.5.11).
Figure 4.5-11 A) Low-intermediate coercivity/temperature components of magnetisation that are not ChRM from individual samples from Pluton Corredores. Where these components are identified over more than two demagnetisation steps (including NRM), the associated error ellipse is indicated. B) Individual site mean directions and overall mean direction (in situ) from Pluton Corredores.
The normal polarity site mean directions are very close to the present day field direction and therefore there may be some unresolved contamination of the ChRM. The normal and reverse groupings do however pass the reversal test (McFadden & McElhinny 1990) with a C-classification. The preservation of anti-parallel magnetisation directions indicates that the magnetisation recorded by the pluton was acquired over a period of at least one reversal event, and that the normal polarity directions are also probably an ancient record of the Earths' magnetic field. The overall magnetisation of Pluton Corredores is therefore interpreted to represent a primary 70Ma TRM, with $D=001.8^\circ$, $I=-49.6^\circ$, with an $\alpha_{95}$ error of 11.1° ($k=30.5$) (Figure 4.5-11).
4.6 Summary

The Tres Cruces field area was chosen primarily to investigate the presence of a N-S spatial gradient that appears to define the southern limit of the Central Andean Rotation Pattern (CARP Figure). A previous study ~50km to the south of Quebrada de Los Choros [Palmer et al., 1980a], suggests that the predominantly Cretaceous strata observed in the Condoriaco area, to the east of La Serena (c.30°S), appears to record significantly less CW rotation than is evident further to the north in the Vallenar area, c.28°S [Gipson, unpublished data].

Palaeomagnetic data from both the Tres Cruces and Condoriaco areas suggests that the majority of sampling units record simple, univectorial magnetisations, that are generally interpreted to be primary in origin, with only the Los Elquinos Formation, sampled by Palmer et al., (1980) clearly recording a post-tilt remagnetisation (Section 4.2). The only exceptions to this observation concerns the ChRM isolated from the Arqueros Formation and Pluton Los Colorados, both of which were sampled along Quebrada de Los Choros. In both of these sampling units the age of the isolated magnetisation is unclear, with several inconsistencies noted.

Arqueros/Quebrada Marquesa Formation

The laterally equivalent Arqueros and Quebrada Marquesa Formations have now been sampled in both the Condoriaco and Tres Cruces study areas. Assuming both localities record a similar magnetisation history, the strata should therefore act as a convenient marker with which to compare the pattern of crustal rotation recorded in the two areas. Whilst Palmer et al., (1980a) were able to sample the Arqueros and Quebrada Marquesa Formations in both limbs of a broad anticline in the Condoriaco area, the equivalent strata along Quebrada de Los Choros are
observed to dip homoclinally towards the E-NE by ~10° and therefore the age of magnetisation could not be unambiguously determined through application of a tilt-test. The palaeomagnetic data from the Condoriaco area obviously passes a fold test, with significantly greater clustering of site mean directions observed in tilt corrected coordinates (Figure 4.2-1). A slight improvement in the clustering of site mean directions from the Tres Cruces area is observed after tilt correction (Figure 4.3-5), suggesting that the Arqueros Formation in the Tres Cruces Area also records a Primary (pre-tilt) magnetisation.

A primary (or at least pre-tilt) magnetisation from the Arqueros Formation in Quebrada de Los Choros is supported by palaeomagnetic data from the dyke swarm sampled at the western end of the valley. Although the in-situ direction recorded by the dykes is more in accord with the existing palaeomagnetic data in the Condoriaco area [Palmer et al., 1980a], it implies a significantly greater amount of rotation is recorded than evident from the Arqueros/Quebrada Marquesa Formation into which the dykes intrude. The simplest answer therefore is to assume that the dykes were intruded prior to the observed deformation of the Arqueros Formation and tilt correction of the palaeomagnetic data to correct for the gentle deformation observed in the Tres Cruces area restores the overall direction recorded by the dykes closer to that recorded by the Arqueros Formation.

Direct comparison of the overall (tilt-corrected) mean directions from the Arqueros/Quebrada Marquesa Formations from the Tres Cruces and Condoriaco sampling localities, indicates that the two areas record slightly different directions, the most obvious difference being a marked inconsistency in the observed inclination, the sedimentary/volcanic sequence sampled along Quebrada de Los Choros recording a much steeper overall mean direction (Figure 4.6-1). Initially assuming that the strata in both areas were magnetised during a similar time
Figure 4.6-1 Overall mean directions calculated for the Arqueros/Quebrada Marquesa Formations in the Quebrada de Los Choros and Condoriaco sampling localities. The expected palaeo-direction corresponding to the time of magnetisation is indicated in yellow.

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Table 4.6-1 Expected palaeofield directions calculated for the Tres Cruces study area using the reference poles of Besse & Courtillot, (2002, corrected 2003).
period, the inclination recorded in the Condoriaco area coincides with that of the expected direction at the latitude of La Serena (within error), using the 130Ma pole of Besse & Courtillot (2002, 2003-Table 4.6-1, Figure 4.6-1). The mean direction recorded by the equivalent strata along Quebrada de Los Choros however, is ~11° steeper than the expected direction calculated using the same pole. This suggests a number of possibilities:

1. The Arqueros Formation in the Tres Cruces area has undergone a reasonable amount of northward transport (which is not supported geologically).
2. The isolated magnetisation was not acquired c.130Ma (i.e. Arqueros Formation has been subject to a remagnetisation in the Tres Cruces area).
3. For some reason the Arqueros Formation in the Tres Cruces area records a field direction that deviates from the time-averaged field direction c.130Ma.
4. (Different) Localised structures may influence the magnitude of crustal rotation in each area.

Whilst the steeper than expected direction in the Quebrada de Los Choros area could indicate that the Arqueros Formation records a younger magnetisation (and indeed records a similar direction to that noted for Pluton Corredores to the west), the appreciable improvement in clustering after application of the small tilt correction (Table 4.3-2, Figure 4.3-5) and lack of obvious alteration of the sampled strata, means it is difficult to reconcile the observed magnetisation as anything other than a pre-tilt magnetisation.

The mixed lithologies sampled along Quebrada de Los Choros (sediments and volcanics), record a reasonable range of directions, suggesting that the effects of PSV have been averaged to some extent, although there is still the possibility that the steeper than expected inclination may result from a temporal perturbation in
the magnetising field. Palmer et al., (1980a) sampled only from the volcanic members of the Arqueros and Quebrada Marquesa Formations and so the observed disparity in overall direction may be associated with lithological differences influencing the direction of magnetisation originally acquired.

As discussed in Sections 4.1 & 4.2, the slight (but appreciable) differences in the overall mean direction observed from each area may reflect differences in the localised structural setting, with several sites from the Condoriaco area being sampled from areas bordering the extensional La Liga Fault (Figure 4.1-3), where deformation of the strata may produce rotations that do not occur around a vertical axis (e.g. MacDonald, 1980).

Despite the pronounced difference (in inclination) between the overall mean directions, comparison of individual site mean directions determined from each locality indicates that data from the two sampling areas are reasonably consistent (Figure 4.6-2), with only normal polarity directions observed in both areas. This is consistent with both areas acquiring magnetisation during earliest Aptian times, at the start of the Cretaceous long normal polarity superchron. The strata sampled along Quebrada de Los Choros is therefore summarised as retaining a Primary magnetisation that is contemporaneous (albeit not identical) to that recorded in the Condoriaco area. The age of the magnetisation recorded by the Arqueros Formation along Quebrada de Los Choros will be discussed further with relation to the magnitude of rotation recorded by the remaining sampling units from the Tres Cruces area in Chapter Eight.

Pluton Los Colorados

The single K-Ar date from the south of the pluton, indicates that the age of the pluton is 96 ± 4 Ma [Emparan & Pineda, 1999], placing the timing of emplacement
Figure 4.6-2 Individual site mean directions from the Arqueros/Quebrada Marquesa Formations from the Tres Cruces and Condoriaco areas.
within the Cretaceous Long Normal Polarity Superchron. The observation that four of the ten sites recorded reverse polarity directions is clearly at odds with this and could be interpreted in a number of ways:

1. The pluton retains a Primary remanence, recording both normal and reverse polarity sub-chrons. The K-Ar date of 96 ± 4 Ma however [Emparan & Pineda, 1999], if correct, falls within the Cretaceous Long Normal Polarity superchron, in a period when no sub-chrons are currently suspected [Ogg & Gradstein, 2004].

2. The single K-Ar age is significantly wrong and the pluton is either considerably older or younger. Such erroneous K-Ar ages are not unknown, with the error normally considered a consequence of the closed system assumption being violated—i.e. Ar has been lost or gained at some point post mineral formation. However, the age is consistent with the position of the granite in the extreme west of the Coastal Batholith [e.g.; Grocott & Taylor, 2002; Moscoso et al., 1982].

3. A dual polarity remagnetisation event occurred, whose age is difficult to constrain.

As there is no direct evidence to suggest that the pluton has undergone significant remagnetisation, and site mean directions pass a reversal test (suggesting that the site mean directions were acquired in antiparallel fields), the overall direction is considered to be a primary TRM. This will be discussed further with regard to the magnitude of crustal rotation recorded by the sampling units along Quebrada de Los Choros.
The La Guardia area lies between 27°30'-28°00'S and 69°30'-70°00'W, with its centre some 70 km southeast of the city of Copiapó, the regional capital of Chilean province Region III, Atacama (Figure 5.1-1). The area studied corresponds closely to that recently mapped by the SERNAGEOMIN at a scale of 1:100,000 [Iriarte et al., 1999], but extends to encompass part of the main Copiapó valley and adjacent Los Loros map sheet [Arévalo, 1994].

Topographically the area lies within the Chilean Precordillera and is divided into a series of distinct mountain ranges by often, deeply incised valleys, which, although allowing access also impose distinct limitations on the accessibility of much of the area (Figure 5.1-1). The altitude ranges between approximately 2200 and typically 3700m (the maximum is 4300m in the extreme SE) with the overall heights of the peaks generally rising to the east. The principal valley is the NW-SE trending Copiapó valley off which are the east-west Quebrada Carrizalillo and the NE trending valleys of the Rio Jorquera and Rio Viscachas de Pulido. The only small village (a few 10s of people) is located at La Guardia in the upper reaches of the Rio Jorquera (Figure 5.1-1).

5.1.1 Stratigraphy

The stratigraphic units of the La Guardia area range from Triassic to Eocene in age and in part these rest unconformably above, or in tectonic contact with, a series of Permian granitoid intrusions (280-260 Ma), which are the local basement
Figure 5.1-1 A) Location of the La Guardia field area with respect to major population centres and topographic features (>3000m in grey). B) Outline map illustrating the principal tectonomorphic features and access routes (in red) through the La Guardia field area. Redrawn from Iriarte et al., (1999).
(Figure 5.1-2). The gross stratigraphy (Figure 5.1-3) of the area (excluding the Permian) can be broadly divided into two geological packages that mark the evolution of the La Guardia area from a back arc setting in the Mesozoic to an active arc setting in the Cainozoic. Throughout the La Guardia area, any of the above units may be locally overlain (especially in the narrower valleys) by unconsolidated Atacama gravels and younger alluvial/fluvial deposits which reflect the area's present day tectonic setting as being part of the modern forearc region of northern Chile.

5.1.2 The Late Triassic-mid Cretaceous backarc

The principal units (Figures 5.1-2 & 5.1-3) are described below following the work of Iriarte et al., (1999):

**La Ternera Formation** (U. Triassic-Lias [Bruggen, 1950; Jensen, 1976])

A clastic sedimentary and volcanic sequence approximately 1,200m thick which unconformably overlies a basement of Paleozoic Granitoids. The basal section consists of ~120m of predominantly reddened polymict conglomerates and coarse-grained sandstones that form lenticular deposits over several of the Paleozoic plutons, the majority of the La Ternera Formation, ~1,000m, comprises andesitic-basaltic lava, pyroclastic/ignimbritic flows and volcanioclastics (Figure 5.1-4).

**Lautaro Formation** (Sinemurian-Bajocian [Segerstrom, 1959])

Resting conformably on the La Ternera Formation (Figure 5.1-4), the Lautaro Formation is composed of sandstones, calcarenites, calcilutites and limestones that are predominantly marine in origin, with a maximum thickness in the south western quadrant of the La Guardia area of ~800m. The formation is observed to thin dramatically to the east where the entire thickness is reduced to as little as
Geology of the La Guardia field area (redrawn after Iriarte et al., 1999 & Arevalo, 1994). As there is no recent mapping to the immediate south and east of the La Guardia mapsheet, sampling localities are projected on to a section of LandSAT 7 ETM+ imagery, displayed using Red (Band 7), Blue (Band 4) and Green (Band 2) channels. Only the La Ternera and Lagunillas Formations were sampled away from the La Guardia mapsheet and generally speaking the La Ternera Formation appears as dark blue to purple in colour, with the underlying Lagunillas and Quebrada Monardes sandstones appearing dark green in colour. Landsat 7 (ETM+) image, downloaded from the Global Land Cover Facility (http://glcf.umiacs.umd.edu/data/landsat). Palaeomagnetic sampling localities from this study (circles) are identified as: 1-La Ternera Formation including a-Fundo Santa Rosa, b-Manflas Plantation and c-Banderitas areas; 2-Lagunillas Formation including a-Jorquera Valley and b-Figueroa Valley; 3-Quebrada Monardes Formation; 4-Hornitos strata (Taylor et al., 2007); 5-Caldera Jorquera; 6-Pluton El Gato. Palaeomagnetic sampling localities from the study of Riley et al., 1993 (squares) are identified as: A-La Ternera Formation (Rio Aguas Blancas); B-Lagunillas Formation; C-Cerrillos Formation; D-Quebrada Seca Formation.
Figure 5.1-3  Generalised stratigraphic log of the La Guardia Area.
Figure 5.1-4 Photograph illustrating the conformable relationship between the La Ternera, Lautaro and Lagunillas Formations in the La Guardia field area. Photo taken along the Rio Jorquera.

Figure 5.1-5 Cross section constructed through the La Guardia mapsheet [from Iriarte et al, 1999].
The reduction in the overall thickness of the formation is accompanied by transition to a more continental character.

**Lagunillas Formation** (Lower Jurassic [Jensen, 1976]) + **Quebrada Vicunita Strata** [Hillebrandt, 1973]

Although in some areas these continental sediments & volcanics are deposited above the Lautaro Formation (Figure 5.1-4), the lower red sandstones of the Lagunillas Formation are also probably in part laterally equivalent to the marine limestones. Where the Lautaro Formation is observed to thin eastwards, the Lagunillas Formation is observed to rest approximately conformably over the La Ternera Formation. The upper part of the Lagunillas Formation is dominated by andesitic to basaltic-andesitic volcanics that are time-equivalent to the volcanics of the Quebrada Vicunita Strata, observed to overly the Lautaro Formation in the west of the study area.

**Quebrada Monardes Formation** (U. Jurassic-L. Cretaceous [Mercado, 1972])

This comprises an almost purely continental clastic series of red bed sediments with increasing volcanics towards the top of the sequence. The sandstones and conglomerates belonging to the Quebrada Monardes Formation are typically thickly bedded and well stratified.

This overall Mesozoic volcanic and sedimentary package is easily recognised throughout the La Guardia area, with the pale cream-buff carbonate beds of the Lautaro Formation clearly separating the much darker continental sediments and dark basaltic-andesitic lavas of the La Ternera and Lagunillas Formations (Figure 5.1-4). The discrimination between the continental red-beds of the Lagunillas and Quebrada Monardes Formations within the field however, proved to be much less obvious.
Cerrillos Formation (Albian-Cenomanian [Segerstrom & Parker, 1959])

Continental sedimentary and volcanic sequence comprising mainly andesitic lavas and volcanic breccias, deposited in a basin whose subsidence was related to crustal extension as documented in the Tierra Amarilla area and the Sierra Fraga extensional complex to the north of the La Guardia area [Mpodozis & Allmendinger, 1993; Arevalo, 1999]. The Cerrillos Formation overlies the older Triassic-L. Cretaceous strata with an angular unconformity including strata belonging to the Bandurrias and Chañarcillo Groups to the south and west of the La Guardia area. The Cerrillos Formation only crops out in the extreme west of the La Guardia area, but it is inferred to underlie younger units in the western part of the mapsheet area.

5.1.3 The Latest Cretaceous-Paleocene-Eocene intra-arc

The Mesozoic, essentially backarc, package described above, is unconformably overlain by a second (younger) package interpreted to represent the initiation and continuation of Paleocene-Eocene intra-arc volcanism/magmatism, as the locus of the active volcanic/magmatic arc migrated eastwards from its' early-mid Cretaceous position in the Coastal Cordillera. The principal units are displayed in Figures 5.1-2 & 5.1-3 and briefly described (after [Iriarte et al., 1999]) as follows:

Hornitos Formation (inc. Sierra La Dichosa lavas/La Higuera strata, Quebrada Seca Formation & Estratos Los Leones) (U. Cretaceous-earliest Tertiary)

Terrestrial sedimentary deposits of the La Higuera strata interfinger with andesitic-basaltic volcanics Sierra La Dichosa lavas, and infill an extensional volcanotectonic half-graben referred to as the Hornitos Basin. The basin evolved during the late Cretaceous to earliest Paleocene, immediately west of the La Guardia
area (78-63 Ma [Arevalo et al., 1994]). Very similar terrestrial volcanic and clastic sequences (Quebrada Seca Formation and Sierra Los Leones Strata) are recognised and infill smaller basins to the east of the Hornitos Basin that resulted from the inversion of the U. Triassic-L. Cretaceous strata in the La Guardia area [Iriarte et al., 1999].

The unconformity at the base of the Sierra Los Leones Strata and Quebrada Seca Formation (which generally rest on the Quebrada Monardes Formation) suggests that this event may be temporally equivalent to the Chañarcillo fold and thrust belt, exposed to the west in the Rio Copiapo Valley and formed by sinistral transpression during the upper Cretaceous (93-78 Ma; [Arevalo & Grocott, 1997; Arevalo, 1999]). ‘Hornitos-type’ deposits have been described as far south as the Huasco valley, some 150km south of Copiapo, with the Elquinos Formation in the Condoriaco area (c.30°S) also interpreted to represent a lateral equivalent of the ‘Hornitos-type’ deposits [Emperan & Pineda, 1999], in the La Serena area.

Megacaldera Carrizalillo, associated volcanic/plutonic complexes & Quebrada El Romero Strata (lower Paleocene-upper Paleocene, [Rivera & Mpodozis, 1994]

Much of the north west quadrant of the La Guardia area is dominated by 1500m of rhyolitic welded tuffs and lavas, topped by ~70m of lacustrine limestones of lower Paleocene age (65-60Ma), identified as the intra-caldera deposits (Quebrada El Romero Strata) of a very large volcanic structure of ~80 x 50 km known as Megacaldera Carrizalillo [Rivera & Mpodozis, 1994]. Nested within or close to the circumference of this larger structure are several smaller calderas, dated between 62-55Ma, which include; Calderas Lomas Bayas, El Durazno, Agua Neuva and Jorquera (Figure 5.1-2). These smaller volcanic centres are noted for their tuffs, lavas, ignimbrites, associated dykes, domes and intracaldera sediments. All of
these volcanic edifices appear to be in part founded unconformably on earlier 'Hornitos-type' deposits.

**Eocene Volcanics**

The final period of caldera construction in the La Guardia area is represented by Caldera Bellavista (Figure 5.1-2), which has ages between 48-45Ma (Figure 5.1-3). After this period, the active volcanic arc migrated eastwards towards its present locus in the Altiplano/Puna.

5.1.4 Plutonism

The single most important pluton, in terms of outcrop area, is the mid-Eocene Pluton El Gato which cuts through much of the older stratigraphy and has an elongate sinuous form in plan (Figure 5.1-2). Four K-Ar dates (biotite) define an age range of 46-41 Ma, with a further K-Ar date (plagioclase) indicating a younger age of 36.5 ± 1.1 Ma, interpreted as probably representing a thermally reset age [Iriarte et al., 1999]. Its map and field relationships strongly suggest that it has a sill like form and the roof is well exposed as being a near horizontal to shallowly eastward dipping contact north of Q. Carrizalillo (Figure 5.1-5). Numerous small intrusions ranging in age from late Cretaceous to Eocene in age crop out in the area in the form of stocks, sills and dykes.

5.1.5 Structural Geology

The geology of the La Guardia area is segmented into a number of structural blocks, divided by a series of N-S to NE-SW trending reverse faults (Figures 5.1-1 & 5.1-2), with average lengths of between 40 & 100 km [e.g. Jensen, 1976; Jensen & Vicente, 1976; Soffia, 1989]. These fault strands form part of the La Ternera fault system, itself part of the 800+km long Domeyko Fault system, which
also includes the well-known West Fissure Fault [Tomlinson & Blanco, 1997a; 1997b].

The westerly verging, medium to low angle (reverse) La Ternera Fault, represents one of the major faults at the extreme edge of the Domeyko Fault System. The fault extends for 65 km to the north of the map [Comejo et al., 1993; Iriarte et al., 1996] where it is observed to have undergone sinistral displacement with a contractional component during a phase of Eocene sinistral transpressive deformation, c. 42-36 Ma [Mpodozis et al., 1994; Tomlinson et al., 1993].

Further to the east, the sub-parallel high angle, inverse, Pauna-La Estancilla, La Iglesia, Vizcachas-La Guardia and Quebrada Aranguiz Faults, verge west and east and separate the area into series of up- and down-thrown blocks. This process has exhumed the Paleozoic basement in places (e.g. Pilar de Montosa—Figure 5.1-2), whilst passively deforming the cover and preserving, in structural depressions, synclines composed of the Cretaceous and Tertiary sequences that pre-date the deformation [Iriarte et al., 1999].

The extreme north of the Pauna-La Estancilla Fault, is cut by the granodiorites of Pluton Carrizalillo (45-41 Ma) which shows no evidence of being offset by the fault. However, further to the east, the Vizcachas-La Guardia Fault displaces granitoides dated c. 39 Ma, which probably indicates that these faults are individually reactivated at different periods during the mid to late Eocene.

Inverse faults of medium to high angle, easterly vergent (Quebrada Las Vaca & La Mona Faults), affect the Quebrada Monardes Formation in the La Guardia Area,

---

1This structural style of "pilares y zanjas" (horst & graben) "thick-skinned" tectonics, is also observed to the south, in the High Cordillera at the latitude of Ovalle (31° S [Godoy & Davidson, 1976; Moscoso & Mpodozis, 1988]) and is probably present in, or to the east of the Los Choros study area.
without inverting the Paleocene tuffs of Caldera Jorquera. Although it is not known how old the faults are, the faults are proposed to result from a phase of deformation occurring during the latest Cretaceous-Paleocene period (c. 72-57 Ma). A similar angular unconformity exists between the Upper Cretaceous and Paleocene sequences in Quebrada Paipote [Iriarte et al., 1996] and El Salvador [Comejo et al., 1997], to the north and north-west respectively.

5.1.6 Palaeomagnetic Sampling

The majority of samples were collected from the area covered by the Hoja La Guardia 1:100,000 map sheet [Iriarte et al., 1999], centred around Caldera Jorquera (Figure 5.1-3). Samples were also collected from the Rio Copiapo valley, shown on the adjacent Hoja Los Loros map [Arevalo, 1994] to the south and west of the La Guardia area. The more detailed description of the localised stratigraphy and geological structures in the La Guardia area provided far greater information in comparison to the 1:250,000 map sheet available for the Tres Cruces field area.

In total six geological units were sampled during January 2003, with individual locations shown in Figure 5.1-2 (Geology & Sampling Localities):

1. Upper Triassic La Ternera Formation (sampled in three areas)
2. Jurassic Lagunillas Formation
3. Upper Jurassic-lower Cretaceous Quebrada Marquesa Formation
4. Uppermost Cretaceous Sierra La Dichosa Lavas
5. Paleocene intra-caldera pyroclastics and lavas belonging to Caldera Jorquera
6. Eocene Pluton El Gato
5.2 Previous Palaeomagnetic sampling in the La Guardia field area

5.2.1 Introduction

Riley et al., (1993), reported evidence of clockwise crustal rotations preserved by a number of Mesozoic formations in northern Chile based on samples collected from the La Guardia area. The sampling area was approximately the same as in this study, covering ~80x30km south and west of Copiapo, with the majority of samples collected along Quebrada Carrizalillo. The samples were collected from what were originally interpreted as the La Ternera, Quebrada Monardes and Cerrillos Formations [Sepulveda & Naranjo, 1982] (Figure 5.1.3 and Table 5.2-1).

When the locations of collected samples are plotted on the more recent 1:100,000 map sheet of the La Guardia area (Iriarte et al., 1999), the stated sample locations/units do not always correspond to the re-interpreted stratigraphy. The palaeomagnetic data from the study of Riley et al., (1993) have therefore been re-interpreted utilising the new stratigraphy of Iriarte et al., (1999) where appropriate.

5.2.2 Implications of the Revised Geology

Outcrop patterns of the La Ternera and Cerrillos Formations (in the Elisa de Bordos area) have changed relatively little from that of Sepulveda & Naranjo (1982) and therefore there is no effect on the interpretation of the original palaeomagnetic data from these formations. The nature of the Quebrada Monardes and Cerrillos Formations (cropping out in the in Cuesta El Gato area) have however been substantially altered. This directly affects the way in which the palaeomagnetic data from these formations can be interpreted, most importantly with respect to the age/nature of the isolated magnetisations.
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<th>Reclassified Stratigraphy/Area</th>
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Table 5.2-1 Location of sampling localities from the study of Riley et al., (1993). The effects of the reclassified stratigraphy [after Iriarte et al., 1999], is shown, with site localities from the La Ternera Formation divided into two separate areas.
Quebrada Monardes Formation (as described by Sepulveda & Naranjo, 1982)

Whereas many previous studies described the entire (~1700m thick in places), transitional continental red bed sequence (+ lavas), generally observed to rest conformably on the carbonate and terrigenous marine Lautaro Formation in the La Guardia area as the Quebrada Monardes Formation [e.g. Bell, 1982, 1989; Sepulveda & Naranjo, 1982]. Iríarte et al., (1999) split the overall sequence into two, with the lower section represented by the Lagunillas Formation, the upper by the Quebrada Monardes Formation (Figure 5.1.3 and 5.1.4).

The Jurassic Lagunillas Formation [originally identified by Jensen, 1976], comprises ~600m of reddened conglomerates, fine to medium grained sandstones, and, critically, an upper largely volcanic member of ~500m of andesitic-basaltic lavas. Iríarte et al., (1999), recognised that the red-bed sequence in the La Guardia area was divided by a series of andesitic and basaltic lavas, below which the strata is dominated by the reddened conglomerates and sandstones characteristic of the basal member of the Lagunillas Formation, with the more truly continental red sandstones of the Quebrada Monardes Formation resting conformably on the lava sequence. Therefore the lavas themselves represent the upper member of the Lagunillas Formation and are interpreted as being stratigraphically equivalent to lavas of similar composition which overlie and interfinger with the Lautaro Formation (Quebrada Vicunita Strata), suggesting that the Lagunillas sandstones and conglomerates are the lateral, marginal marine/continental equivalent of the marine carbonates dominating the Lautaro Formation. This would imply that the sandstones sampled by Riley et al., (1993), are actually older than originally interpreted (Early Jurassic as opposed to latest Jurassic-earliest Cretaceous), and possibly of a similar age to the marine limestones of the Lautaro Formation (Sinemurian-Bajocian).
Cerrillos Formation

At Cuesta El Gato in the La Guardia area, Sepulveda & Naranjo, (1982), interpreted the volcanic and volcaniclastic deposits overlying the Quebrada Monardes Formation to represent a continuation of the Cerrillos Formation from the west. However the Cerrillos Formation as sampled by Riley et al., (1993) is now regarded as volcanic deposits that preceded the formation of Caldera Jorquera [Iríarte et al., 1999]. These volcanic and volcaniclastic deposits, form the Quebrada Seca Formation and rest unconformably upon the Quebrada Monardes Formation and are latest Cretaceous in age, which is some 20-25Ma younger than the Albian-Cenomanian age suggested for the Cerrillos Formation in the Rio Copiapo valley [Arevalo, 1994].

Where the new stratigraphy has had no effect, the overall Formation mean directions are taken directly from Riley et al., (1993), otherwise the original data has been subdivided based either on the new stratigraphy, or in the case of the La Ternera Formation, the data is separated based on geographical location.

5.2.3 Reinterpretation of Palaeomagnetic Data

The palaeomagnetic data from the study of Riley et al., (1993), will be discussed in relation to the originally identified sampling formations.

La Ternera Formation

The oldest strata sampled by Riley et al., (1993), was collected from the late Triassic (to earliest Jurassic) La Ternera Formation, predominantly from along the Rio Aguas Blancas valley to the north and east of the La Guardia area sampled herein (Figure 5.1-3). The palaeomagnetic results from the La Ternera Formation (Table 5.2-2, Figure 5.2-1) show that the rocks have recorded both normal and
### Table 5.2-2

Site mean directions for the La Ternera Formation [from Riley et al., 1993]. Data is divided into two areas (Rio Aguas'Blancas-14 localities, Banderitas-4 localities), with overall mean shown representing that calculated from the original study.

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Mean 14/14 | 33.0 | -51.9 | - | - | 40.0 | 6.4 |

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Mean 4/4 | 36.7 | -41.1 | - | - | 69.2 | 11.1 |

All Mean 18/18 | 33.9 | -49.5 | - | - | 39.9 | 5.5 |

### Table 5.2-3

Site mean directions for the Quebrada Monardes Formation [from Riley et al., 1993], with overall mean shown representing that calculated from the original study. (1)Samples now interpreted to be sourced from the Lagunillas Formation [after Iriarte et al., 1999]. Remaining samples interpreted to be sampled from the lower sedimentary member of the Quebrada Seca Formation [after Iriarte et al., 1999]. Site 90qm51 (shaded) excluded from calculation of overall mean direction due to the near horizontal inclination.

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<tr>
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<td>41.1</td>
<td>21.9</td>
<td>20.0</td>
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</tbody>
</table>

All Mean 8/9 | 15.5 | -40.9 | - | - | 23.7 | -36.7 | 15.5 | 14.5 |

Mean (1) 4/4 | 20.1 | -48.5 | - | - | 6.3  | 39.6 |

(1) Samples now interpreted to be sourced from the Lagunillas Formation [after Iriarte et al., 1999]. Remaining samples interpreted to be sampled from the lower sedimentary member of the Quebrada Seca Formation [after Iriarte et al., 1999]. Site 90qm51 (shaded) excluded from calculation of overall mean direction due to the near horizontal inclination.
Figure 5.2-1 Stereonet plot of individual site mean directions and overall direction calculated for palaeomagnetic data from the La Ternera Formation (from Riley et al., 1993). Circles represent the Rio Aguas Blancas locality, squares the Banderitas locality.
reverse polarity site-mean directions, shown to be demonstrably anti-parallel at the 95% confidence level, passing the reversal test of McFadden & McElhinney (1990) with a classification of C. This would suggest that a stable ancient remanence direction has been isolated.

As shown in Table 5.2-2 and Figure 5.2-1, there is very little difference between the in-situ and tilt-corrected La Ternera directions, although the statistics are marginally poorer after tilt correction. Although the site level tilt corrections are small in magnitude, the bulk of the La Ternera Palaeomagnetic data actually fail a tilt-test, although Riley et al., (1993) suggested there was no geological reason to suspect remagnetisation of the La Ternera Formation in this area and interpreted the magnetisation to be primary and pre-tilting in origin (Table 5.2.2).

Riley et al., (1993) sampled the La Ternera Formation in two areas, the Rio Aguas Blancas (14 sites) valley and the Banderitas area (4 sites) of the Rio Jorquera valley (Figure 5.1-3 & Tables 5.2-1 & 5.2-2). In the Rio Aguas Blancas a post-tilting remanence is recorded, with the data failing a tilt test as the individual site mean directions become more dispersed after application of a tilt correction (Table 5.2-2). Sites in the Banderitas area appear to record a pre-tilt magnetisation, with the application of site level tilt corrections resulting in a marginal increase in clustering, although the tilt-test is indeterminate (Table 5.2-2).

Given the more recent mapping of the La Guardia area by Inarte et al., (1999) it is clear that the two sampling areas are located in different structural blocks separated by the major Viscachas La Guardia reverse fault (Figure 5.1.2). The palaeomagnetic data from the La Ternera Formation from the study of Riley et al., (1993) will be discussed further with respect to data collected from the same formation in this study.
Quebrada Monardes Formation

Riley et al., (1993), established that the red volcanioclastic sandstones retained a stable ancient remanence (Table 5.2-3), with normal and reverse polarity site mean directions observed to pass a reversal test [McFadden & McElhinney, 1990]. The mean site ChRM directions from the Quebrada Monardes Formation are clearly more tightly clustered after structural correction (Figure 5.2-2, Table 5.2-3), indicating that they record a primary, pre-tilting magnetisation. This positive fold test however, is clearly most influenced by four of the site directions, which cluster quite significantly after the application of a tilt-correction (Figure 5.2-2 (circles) and Table 5.2-3).

When the sampling localities are plotted on the La Guardia map and a LandSat 7 (ETM+ image), it is clear that many of the Quebrada Monardes samples collected by Riley et al., (1993), were in fact sampled from red beds situated very close to the top of the La Ternera Formation. (which is characteristically very dark in the LandSat image while the Lagunillas and Quebrada Monardes red beds appear to be green in colour (7-4-2 R-G-B)). This suggests that four of the Quebrada Monardes sites, 90qm14, 18, 19 & 85, are in fact sampled from the Lagunillas Formation and it is these sites which yield a positive fold test (Table 5.2-3, Figure 5.2-3).

Riley et al., (1993), noted that clasts from three conglomeratic beds from within the lower Quebrada Monardes Formation produced randomised directions, further suggesting a primary pre-tilt origin for the formation mean direction, the majority of the clasts were volcanic andesites of presumably Triassic or early Jurassic age. These conglomerate beds are located just above the Lautaro limestones, and correspond to the Lagunillas Formation as defined by Iriarte et al., (1999). The positive fold test (Figure 5.2-3, Table 5.2-3) and failed conglomerate tests noted
Figure 5.2-2  Stereonet plot of individual site mean directions and overall direction calculated for palaeomagnetic data from the Quebrada Monardes Formation (from Riley et al., 1993). Circles represent samples now classified as being sampled from the Lagunillas Formation; squares represent samples from the lower sedimentary member of the Quebrada Seca Formation.
Lagunillas Fm. (Riley et al., 1993) +
in situ

Lagunillas Fm. (Riley et al., 1993) +
tilt corrected

Figure 5.2-3 Stereonet plot of individual site mean directions and overall direction calculated for palaeomagnetic data from the Lagunillas Formation (recalculated from Riley et al., 1993, after Iriarte et al., 1999).
for these samples, here interpreted to have been collected from the Lagunillas Formation, indicate that the overall formation records a pre-tilt primary magnetisation.

Of the remaining four site mean directions originally used to calculate the overall Quebrada Monardes Formation mean direction displayed in Figure 5.2-2 (squares) and Table 5.2-3, site 90qm51 has been omitted from further analysis, due to a near horizontal inclination in both in-situ and tilt-corrected coordinates. The remaining three sites display little obvious clustering either in situ or after tilt correction. From Figure 5.1-3, it is believed that these three sites should now be considered to have been collected from sandstones belonging to the lower sedimentary member of the late Cretaceous to earliest Palaeocene Quebrada Seca Formation, which overlies the Quebrada Monardes Formation with an angular unconformity. The palaeomagnetic data from these three sites belonging to the Quebrada Seca Formation will be discussed further in a later section, as will those from the Lagunillas Formation.

Cerrillos Formation

Riley et al., (1993) sampled strata they interpreted as belonging to the Cerrillos Formation (post-Aptian /110 Ma) at two separate localities (Figure 5.1-3). The first sampling locality the Elisa del Bordos area is situated at the western end of Quebrada Carrizalillo. Here the palaeomagnetic data were interpreted as recording a post-tilt magnetisation, comprising solely of normal polarity directions, which are obviously dispersed by the application of a tilt correction (Table 5.2-4, Figure 5.2-4).

Although only normal polarity site mean ChRM directions were observed, the overall direction isolated is clearly not the Present-day field due to the fact the
### Table 5.2-4

Site mean directions for the Cerrillos Formation in the Elisa de Bordos locality [from Riley et al., 1993]. Overall mean shown represents that calculated from the original study, which excluded the direction from site 90eb70 (shaded) due to the much larger associated error calculated in comparison to the remaining sites.

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<th>Inc. (°)</th>
<th>Dec. (°)</th>
<th>Inc. (°)</th>
<th>k</th>
<th>a95 (°)</th>
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### Table 5.2-5

Site mean directions for the Cerrillos Formation in the Cuesta El Gato locality [from Riley et al., 1993], with mean direction shown (for CS localities) calculated from the original study. These samples are now interpreted to belong to the upper volcanic member of the Quebrada Seca Formation [after Iriarte et al., 1999] and are therefore combined with samples interpreted to belong to the lower sedimentary member of the Quebrada Seca Formation, originally sampled as the Quebrada Monardes Formation (Table 6.2-3). Sites 90qm51 and 90qm54 (shaded) were excluded from the calculation of the overall mean direction due to the near horizontal inclination (51) and much larger associated error (54).

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<td>49.3</td>
<td>-</td>
<td>-</td>
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Figure 5.2-4 Stereonet plot of individual site mean directions and overall direction calculated for palaeomagnetic data from the Cerrillos Formation in the Elisa de Bordos locality (from Riley et al., 1993).
formation obviously records a substantial amount of CW rotation (Figure 5.2-4). *Riley et al.,* (1993) loosely attribute the timing of remagnetisation to a normal polarity period during the early Tertiary. However the tight clustering of site-mean directions (Figure 5.2-4) brings in to question whether or not palaeo-secular variation has been completely averaged and if the apparently large amount of CW rotation recorded is in fact just a (geologically) instantaneous record of a field, deviating substantially from the time-averaged dipole field of the time.

Assuming the ChRM direction recorded by the Cerrillos Formation at the Elisa de Bordos locality has adequately sampled PSV, a secondary remagnetisation with an overall direction of \( D=048.6^\circ, I=-44.9^\circ \), with an \( \alpha_{95} \) of 5.2\(^\circ\), is recorded.

The second sampling locality within the Cerrillos Formation as sampled by *Riley et al.,* (1993), is in the Cuesta El Gato located toward the eastern end of Quebrada Carrizalillo (Figure 5.1-3). As observed in the Elisa de Bordos area, the predominantly andesitic lavas sampled record an obvious post-tilt remagnetisation, with the characteristic site mean directions tightly clustered in-situ (Table 5.2-5, Figure 5.2-5). Unlike the Elisa de Bordos area however, all of the sites are characterised by a reverse as opposed to normal polarity. This would indicate once again, that the isolated remanence is an ancient direction unrelated to that of the present day magnetic field.

The obvious difference in polarity in the Elisa de Bordos and Cuesta El Gato areas indicates that the two areas were remagnetised at different times, or over a period spanning at least one reversal of the magnetic field. The two areas also record very different amounts of clockwise rotation. The Cerrillos Formation in the CEG area clearly records significantly less clockwise rotation than is observed in the EDB area, only 50km to the west.
Figure 5.2-5 Stereonet plot of individual site mean directions and overall direction calculated for palaeomagnetic data from the Cerrillos Formation in the Cuesta El Gato locality (from Riley et al., 1993). Now re-interpreted as the upper volcanic member of the Quebrada Seca Formation (after Iriarte et al., 1999).
Riley et al., (1993) suggest that, as long as the effects of palaeosecular variation were adequately averaged, the two areas record different magnetisation histories, and are also probably situated within different structural blocks. However, according to the most recent mapping of the La Guardia area, the samples collected from the Cerrillos Formation in the CEG area by Riley et al., (1993), were actually collected from the upper volcanic member of the Quebrada Seca Formation, deposited unconformably over the Cerrillos Formation, immediately to the east of the El Gato intrusion (Figure 5.1-3).

As discussed previously, Riley et al. (1993) also sampled the lower sedimentary member of the Quebrada Seca Formation identified by Iriarte et al., (1999). As the two members of the Quebrada Seca Formation were sampled reasonably close together (Figure 5.1-3), it is considered appropriate to combine the palaeomagnetic data to produce a mean overall ChRM direction for the Quebrada Seca Formation (Table 5.2-5, Figure 5.2-6). Although the upper volcanic member records only reverse polarity directions, the lower sedimentary member (originally identified as belonging the Quebrada Monardes Formation), records both normal and reverse polarity directions, which when all the data is combined, pass the reversal test of MacFadden & McElhinney (1990), indicating that the isolated ChRM was acquired over a period of one or more reversals. The combined individual site-mean ChRM directions isolated from within the Quebrada Seca Formation are clearly dispersed on the application of a tilt-correction, indicating that the Formation records a post-tilt (Secondary) magnetisation (Figure 5.2-6, Table 5.2-5) with D=196.3°, I=49.3°, with an associated α₉₅ of 12.9°.
Figure 5.2-6 Stereonet plot of individual site mean directions and overall direction calculated for palaeomagnetic data from the Quebrada Seca Formation (recalculated from Riley et al., 1993, after Iriarte et al., 1999).
5.3 The Late Triassic La Ternera Formation

5.3.1 Sampling Localities

This unit was sampled in three areas (Table 5.3-1), Fundo Santa Rosa and Manflas (probably within the same structural block) and Banderitas situated in the NE of the Jorquera valley, clearly separated from the others by major strands of the La Ternera Fault system (see Figures 5.1-2, 5.1-3, Section 5.1). In addition Riley et al., (1993) collected four sites in the Banderitas area and a further 14 in the Rio Aguas Blancas area, again in a separate structural block immediately east of the Banderitas area (Section 5.2, Figure 5.1-3). In summary the late Triassic strata have been sampled in three separate blocks within the La Ternera fault system. This should permit some insight concerning the overall effect of Eocene-Oligocene deformation over the observed pattern of crustal rotations in the La Guardia area.

Fundo Santa Rosa

This area is located in the SW of the study area where the La Ternera Formation rests unconformably on the Palaeozoic La Estancilla Pluton (Figures 5.1-2 & 5.3-1a). The presence of a distinctive, cream coloured package of marine calcareous sediments (Figure 5.3-1b), suggests the sampled strata are situated in the lower part of the volcanic facies of the La Ternera Formation [Iriarte et al., 1999].

Most sites were sampled from individual basaltic lava flows, although site AT3-27 was from a metasediment and AT3-29 a coarse conglomeratic horizon (probably belonging to the basal sedimentary member of the La Ternera Formation). In the conglomerate the clasts are predominantly granitoid in composition, although sandstone and heavily altered volcanic clasts were also common. Although the
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<th>ChRM Isolation Min.</th>
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<td>NRM</td>
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**Table 5.3-1** Site location, lithology and summary of sample demagnetisation behaviour from the latest Triassic-earliest Jurassic La Ternera Formation. (1*)-Sampling Localities from the study of Riley et al., (1993).
Figure 5.3-1  A-Map illustrating the location of sampling sites in the Fundo Santa Rosa-Manflas sampling areas from the La Ternera Formation. Geology and satellite image as displayed in Figure 5.1-2 with the dark blue/purple area corresponding to the La Ternera Formation. B-Photograph illustrating the pale cream coloured marine sediments observed within the predominantly volcanic facies of the La Ternera Formation in the Fundo Santa Rosa area. The sediments appear to be thrust out towards the north [Iriarte et al., 1999]. View to the NW.
Figure 5.3-1. Map illustrating the location of sampling sites in the Banderitas (west)-Rio Aguas Blancas (east) sampling areas from the La Ternera Formation. Geology and satellite image as displayed in Figure 5.1-2 with the dark blue/purple area corresponding to the La Ternera Formation. Sampling sites marked in yellow are from the study of Riley et al., (1993).
matrix was too unconsolidated to sample, sufficient clasts were collected without significant disturbance in order to carry out a conglomerate test.

**Manflas**

The Manflas area lies just west of the main Copiapó valley, some ~5km north of, Fundo Santa Rosa (Figure 5.1-3) and is assumed to be in the same structural block. In this area, the middle volcanic member of the La Ternera Formation crops out as a series of homoclinally dipping lava flows that were sequentially sampled over a ~1.5km traverse. The individual flows were easily identified with clearly defined upper and lower surfaces. The lava flows in the Manflas area are considered to be of a similar late Triassic age to those sampled in the Fundo Santa Rosa area.

**Banderitas**

The most northerly sampling area lies within the Jorquera river valley (Table 5.3-1, Figures 5.1.2 & 5.3-1c). The area is part of a 'pop-up'-type structure, delimited to the east and west by lengthy reverse faults belonging to the wider Domeyko (La Ternera) Fault System (Figures 5.1-2 & 5.3-1c).

The majority of sites are volcanic flows and lie directly below the conformable contact with the overlying marine Lautaro Formation. The proximity to the overlying Lautaro Formation, suggests an earliest Jurassic age for the uppermost La Ternera Formation sampled in this area. The westernmost sampling locality belongs to an andesitic breccia that was sampled close to the bounding thrust zone, immediately NE of Caldera Jorquera and was collected with the intention of testing the age of magnetisation in relation to the observed thrust faulting.
5.3.2 Magnetic Mineralogy

Samples from the La Ternera Fm. exhibit a wide range of behaviours. The majority of samples clearly exhibit indications of two or more magnetic minerals, with haematite often the dominant magnetic mineral in many of the samples (e.g. site AT3-72-Figure 5.3-2a). In haematite dominated samples, a characteristic 'Hopkinson's' peak in the bulk susceptibility at approximately 600-650°C is commonly observed, with a Curie (Neel) temperature of ~700°C. IRM acquisition experiments demonstrate that little magnetisation is acquired in fields <200mT, with samples clearly not saturating in the maximum field applied (800mT-Figure 5.3.2a). Three–component IRM experiments are dominated by the hard component, which is not destroyed until after treatment to 670°C (Figure 5.3.2a). Much smaller magnitude soft and intermediate coercivity components are also observed, but these too are apparently carried by haematite being unblocked at temperatures >600°C.

Many of the rock magnetic experiments indicate that magnetite coexists with haematite, with inflexion points in k-T heating curves noted at ~580°C (e.g. site AT359-Figure 5.3-2b). Such samples display much higher bulk susceptibilities than the haematite dominated samples, but haematite is still probably the dominant mineral when the much higher magnetic susceptibility of magnetite is taken in to consideration. IRM experiments demonstrate much steeper acquisition curves between 0-200mT, consistent with magnetite. The observation of the acquisition of a significant component of IRM in fields >200mT, combined with the lack of saturation by 800mT, confirms that haematite is still present in significant quantities (c.f. Figure 5.3-2b and 5.3-2a). Thermal demagnetisation of composite IRMs' indicates the intermediate coercivity component (300mT) dominates the IRM and is unblocked by treatment to 580°C, as is a significant soft coercivity
Figure 5.3-2 Magnetic mineralogy experiments on representative lava/volcaniclastic samples from the overall La Ternera Formation. A) Haematite dominated samples, B) Magnetite dominated samples, C) Pyrrhotite dominated samples. Diagrams use the standard nomenclature.
fraction. The remainder of IRM (including a hard component of similar magnitude to the soft component) is fully destroyed by treatment to 670°C. This is interpreted to represent a mixture of magnetite and haematite, with single domain or ti-poór magnetite interpreted to dominate the IRM, due to the observation of a magnetite dominated intermediate coercivity component (Figure 5.3-2b).

Other behaviours include the presence of pyrrhotite and what appears to be pure magnetite in some samples.

5.3.3 Demagnetisation Behaviour

Generally the demagnetisation behaviour of the La Ternera samples can be split into three groups, based on the relative contributions of haematite, magnetite and another low temperature mineral in carrying the characteristic direction of remanence, as discussed below.

A-type remanence

The most common demagnetisation behaviour is of a single univectorial, origin bound component of magnetisation of either normal (dominant) or reverse polarity (Figures 5.3-3a-d). Such samples are extremely resistant to AF demagnetisation, with the ChRM direction only clearly isolated using the thermal technique. Normally a low temperature component of magnetisation is not observed, except where the ChRM direction is of reverse polarity and a small component of magnetisation of normal polarity is often removed in temperatures up to 400°C (e.g. Figure 5.3-3c). In the case of normal polarity ChRMs, the vector of magnetisation remains static around the NRM direction up to these temperatures (Figures 5.3-3a & b).
A-Type Behaviour-Univectorial Haematite component

B-Type Behaviour-Univectorial Magnetite & Haematite components

C-Type Behaviour-Overlapping Magnetite & Haematite components

Figure 5.3-3 Representative Zijderveld plots for typical Thermal and AF demagnetisation behaviour of samples collected from the late Triassic-early Jurassic La Ternera Formation in the La Guardia Area. A-D) A-type behaviour (Thermal only) characterised by a univectorial, origin bound ChRM direction, carried almost exclusively by haematite (with exception of D - see text for explanation). E-H) B-type behaviour characterised by a univectorial ChRM direction, carried predominantly by magnetite, but also haematite. I-J) Site AT3-54 displays a dual component remanence carried by pyrrhotite, magnetite and haematite, producing curved demagnetisation paths.
The A-type ChRM direction begins to unblock at ~400°C, with a small component removed by treatment to 580°C suggesting that magnetite carries a minor component of the ChRM direction. The majority of remanence is destroyed between 640-670°C (Figures 5.3-3a,b & d), with a small number of samples not fully demagnetised at the highest temperature (Figure 5.3-3c). This suggests that haematite (with subordinate magnetite) is the principal carrier of magnetisation in these samples.

**B-type remanence**

This differs from Type-A in that a lower temperature/lower coercivity component (with regard to haematite) contributes more significantly to carrying the ChRM direction (Figures 5.3-3e-h). A consequence of this is that AF demagnetisation is as successful in demagnetising these samples as Thermal demagnetisation. Generally a small, low temperature normal polarity component is removed from all samples (typically <400°C), but is again only clearly distinguishable when the ChRM direction is of reverse polarity (Figure 5.3-3h). Further demagnetisation, AF or thermal, progressively removes a single origin-bound component of magnetisation. This suggests that the B-type ChRM direction is carried by both magnetite and haematite, with the observation that AF demagnetisation almost completely destroys the overall remanence suggesting that magnetite carries a greater proportion of the ChRM.

**C-type remanence**

The final, C-type, demagnetisation behaviour was only observed in site, AT3-54, although Sites AT3-30 & 67 are interpreted to retain similar (if poorly defined) magnetisations. In the B-type magnetisation both magnetite and haematite carry the same single direction of magnetisation, samples from site AT3-54 suggest that
these minerals carry antiparallel components of magnetisation (Figures 5.3-3i & j). This is most clearly shown by thermal demagnetisation, which produces a curved demagnetisation path.

A low temperature/coercivity component is identified as the normal polarity linear segment visible on most Zijderveld plots of the demagnetisation data from this site. This component approximates the present day field direction (or an unrotated ancient direction-Figure 5.3-4a). The higher coercivity/temperature component is never defined by a stable end-point but can be estimated using constrained great circles. All of the samples from Site AT3-54 exhibit great circle demagnetisation paths of almost identical trajectory and produce a well-defined mean direction (Figure 5.3-4b). This direction is clearly rotated CW and is therefore unrelated to the low coercivity/temperature (present day) normal polarity direction identified, suggesting that it is an ancient remanence. The high temperature/coercivity direction is therefore considered to represent the ChRM direction, equivalent to that carried by the A- and B-type remanences.

**Distribution of demagnetisation behaviours**

The demagnetisation data from all three sampling areas is summarised in Tables 5.3-2 & 3, with the dominant demagnetisation behaviour of each site also indicated. Demagnetisation behaviour appears to be directly related to lithology as discussed in Section 5.3-1, with A-type behaviour is typically associated with conspicuously reddened lava flows in all sampling areas, but these are noted to dominate in the Fundo Santa Rosa and Manflas areas (Figure 5.3-2). B-type behaviour is dominantly recognised from pyroclastic units sampled in the Banderitas area, although a few of the lava flows sampled also display either similar behaviour or a mixture of A- & B-type behaviour. C-type behaviour is the least common observed, and is neither confined by locality nor lithology.
Figure 5.3-4 Individuai sample magnetisation directions and overall site mean direction calculated for Site AT3-54, which displays a C-type remanence. A) Low temperature/coercivity components record the present day field direction or an unrotated ancient direction. B) High temperature/coercivity components identified using constrained great circles. The site mean direction is clearly rotated CW and unrelated to the low temperature/coercivity component.
Table 5.3-2  Site mean ChRM directions and overall mean direction, from the La Ternera Formation, Fundo Santa Rosa and Manflas sampling areas. Superscript letter refers to the dominant demag. behaviour identified at each site. FSR-sites AT3-27 & 28 (shaded) are located in a separate structural block to the remaining sites and are therefore excluded from the overall mean direction. Manflas-site AT3-67 excluded from the overall mean direction due to the highly unusual orientation.

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<td>6.8</td>
<td>-49.6</td>
<td>124.3</td>
<td>5.4</td>
</tr>
<tr>
<td>AT3-60 A</td>
<td>8/8</td>
<td>148.2</td>
<td>56.9</td>
<td>195.0</td>
<td>61.3</td>
<td>96.6</td>
<td>6.6</td>
</tr>
<tr>
<td>AT3-61 A</td>
<td>8/8</td>
<td>7.2</td>
<td>-74.1</td>
<td>48.8</td>
<td>-55.0</td>
<td>55.0</td>
<td>7.5</td>
</tr>
<tr>
<td>Mean AT3-62</td>
<td>8/8</td>
<td>2.1</td>
<td>-47.1</td>
<td>-</td>
<td>-</td>
<td>150.4</td>
<td>10.1</td>
</tr>
</tbody>
</table>

Table 5.3-3  Site mean ChRM directions and overall mean direction, from the La Ternera Formation, Banderitas sampling area. Superscript letter refers to the dominant demag. behaviour identified at each site. (1*)-Localities from Riley et al., (1993).
5.3.4 Discussion of Palaeomagnetic Data

Combining these new data with that of Riley et al., (1993) the La Ternera Formation has been sampled at four localities in the La Guardia area. On reflection subsequent to fieldwork, the Fundo Santa Rosa and Manflas Plantation areas are considered to be situated in the same structural block, due to similarity of dips, stratigraphy and lithology coupled with a lack of mapped evidence for an intervening fault and are therefore treated as a single locality hereafter. The Banderitas and Rio Aguas Blancas localities are separated by strands of the DFZ and are therefore considered to be structurally separate from one another (Figures 5.1-2 & 5.1-3). Systematic analysis of the data should permit the detection of any localised rotations due to the influence of the La Ternera Fault Zone, as well as regional rotation that may be recorded.

Fundo Santa Rosa and Manflas Areas

ChRM directions from the Fundo Santa Rosa and Manflas localities (Table 5.3-2, Figure 5.3-5) are dominantly carried by A-type remanences. Individual site mean directions are well defined and typically of reverse polarity, but are collectively dispersed, showing a distinct, strung distribution in situ. The site mean directions fail a tilt test (Table 5.3-2, Figure 5.3-5), suggesting a post-tilt remanence has been isolated despite the essentially uniform dip of the sampled strata, throughout both sampling localities (with exception of sites AT3-27 & 28-Table 5.3-1, Figure 5.3-5).

Tilt correction produces unfeasibly high inclinations, which would imply a palaeolatitude of nearly 60°, which is absurd when compared to the majority of palaeomagnetic data from elsewhere in the region (Table 5.3-2, Figure 5.3-5). The only explanations for such a steep inclination being of primary origin would
Figure 5.3-5  Stereonet projections of site ChRM directions and overall mean direction from the La Ternera Formation in the Fundo Santa Rosa and Manflas Localities. A) In situ. B) Tilt corrected.
therefore be that either it does not reflect a simple dipole field at the time of acquisition, or that the tilt of the strata is the result of compound deformation and requires some complex correction; meaning the incorrect application of a tilt-correction may produce an apparent (fictitious) tectonic rotation about a non-vertical axis [e.g. MacDonald, 1980].

Further evidence for the origin of the A-type remanence in this area comes from site AT3-29 which was collected from a ~70m long road cutting through thick conglomerates, containing well-rounded, predominantly granitic clasts with a smaller percentage of andesitic and sedimentary/meta-sedimentary clasts. Twelve individual clasts including Permo-Triassic granitoids, sandstones, meta-sediments and highly weathered volcanics were sampled. Individual clasts directions display two different behaviours, with in-situ directions measured from granitoid and other magnetite-dominated clasts being scattered (Figure 5.3-6a) while sandstone and (reddened) haematite-dominated clasts cluster about a reverse polarity direction (Figure 5.3-6b). This result indicates selective remagnetisation has taken place and it would appear A type magnetisations are a result of this, at least in the Fundo Santa Rosa localities.

Taken together the results from Manflas and Fundo Santa Rosa localities would suggest that a post tilting overprint has remagnetised both lavas and sandstones, resulting in the growth of haematite and of magnetite (or the contemporaneous resetting of the magnetite component), such that both minerals carry the same direction. A post tilting remanence is therefore interpreted to be recorded by the La Ternera Formation in this area (although this will be discussed further in the wider context), with a mean direction of $D=156.8^\circ$, $I=53.1^\circ$ and an $\alpha_{95}$ of $9.8^\circ$ (Table 5.3-2).
Figure 5.3-6 Stereonet projections of magnetisation directions isolated from individual clasts sampled from the polymict conglomerate at Site AT3-54. A). Randomly orientated directions (in situ) from granite clasts (unless otherwise stated). B) Sedimentary clasts displaying a secondary haematite overprint with mean direction indicated (in situ).
Palaeomagnetic data from this locality include four sites from the study of Riley et al., (1993) (Table 5.3-3). Both haematite dominated (A-type) and magnetite dominated (B-type) remanences were observed, with B-type remanences common in the pyroclastic units sampled. Sites collected from the Banderitas area have a much greater range of bedding attitudes implying that a tilt-test of the site mean ChRM directions will be more meaningful with regard to determining the nature of the magnetisation isolated.

Isolated directions display low within site scatter, with a greater between site scatter interpreted to represent the complete averaging of the effects of PSV (Table 5.3-3, Figure 5.3-7). Both a positive overall fold test (Table 5.3-3, Figure 5.3-7), as well as a positive conglomerate test sampling discrete blocks from an andesitic breccia (Figure 5.3-8), indicate that a pre-tilt magnetisation is recorded. This is in direct contrast to the overall magnetisation identified from the Fundo Santa Rosa / Manflas area. In addition, although the magnetisation is dominantly of normal polarity, the two reverse polarity directions identified, pass a reversal test with a classification of C (McFadden & McElhinny, 1988), again indicating that an ancient direction has been isolated.

The final piece of evidence pointing to the preservation of an ancient magnetisation direction, concerns the fact that a low temperature/coercivity direction recovered from three sites, appears to record the present day (or at least an unrotated normal polarity) field direction (Table 5.3-3, Figure 5.3-9). Application of the tilt-correction to these directions causes pronounced scattering and indicates a clearly post-tilting origin for this component of magnetisation (Figure 5.3-9). The fact that this component is clearly different to the ChRM direction
Figure 5.3-7  Stereonet projections of site mean ChRM directions and overall mean direction from the Banderitas locality in the La Ternera Formation A) In situ. B) Tilt-corrected. Circles-this study, squares-Riley et al., (1993).
Figure 5.3-8 Stereonet projection of individual sample magnetisation directions from an andesitic breccia from the Banderitas locality in the La Ternera Formation.

Figure 5.3-9 Stereonet projections of low-intermediate coercivity/temperature mean magnetisation directions from the Banderitas locality in the La Ternera Formation. A) In situ. B) Tilt corrected.
identified from the overall area, re-enforces the ancient origin for the ChRM direction.

Assuming a Primary (pre-tilt) magnetisation is recorded, the La Ternera Formation in the Banderitas area records an overall mean direction of $D=036.9^\circ$, $I=-54.2^\circ$, with an $a_{95}$ error of $7.4^\circ$ (Table 5.3-3). The inclusion of the four sites from the study of Riley et al., (1993) has little effect on the overall ChRM direction, but significantly decreases the overall error. The fact that two independent studies identify near identical ChRM directions, displaying similar demagnetisation behaviours, suggests that the overall magnetisation can be considered a genuine carried record of the ancient field.

**Aguas Blancas**

The remaining palaeomagnetic data to be discussed is that from the Aguas Blancas river valley collected by Riley et al., (1993) (see Section 5.2, Table 5.2-2 and Figure 5.2-1). Both normal (8 sites) and reverse (6 sites) polarities were present and pass the reversal test (McFadden & McElhinny 1988) with a classification of C, suggesting that a stable (ancient) magnetisation, spanning at least a single reversal event was recorded. The application of a tilt-correction has little effect on the overall observed directions as strata of only of limited range of attitudes (with small dip) were sampled, although the individual site-mean directions are slightly more clustered in-situ, hence the fold-test was statistically indeterminate. Assuming a primary origin for the observed ChRM (as proposed by Riley et al., (1993)), the Aguas Blancas area records an overall mean direction of $D=030.2^\circ$, $I=-51.0^\circ$, with an $a_{95}$ error of $8.2^\circ$. This direction is very similar to that observed in the Banderitas area immediately to the west.
Palaeomagnetic data suggests that the La Ternera Formation records a diachronous magnetisation history in the three areas in which it has been sampled. To the south, the Fundo Santa Rosa-Manflas area records a post-tilting (and anticlockwise rotated) remanence, while the northern sampling areas are suggested to retain a pre-tilting remanence (despite the similarities in demagnetisation behaviour). The ChRM directions, determined for each of the three areas are displayed in Figure 5.3-10, which shows that the mean direction from the Fundo Santa Rosa-Manflas area is clearly removed from that recorded by the northerly sampling areas. (In addition it can be seen, as argued previously, if the Fundo Santa Rosa-Manflas result is considered in tilt corrected coordinates the unacceptably steep inclination would need to be explained).

The apparently strung distribution of site mean directions from the Fundo Santa Rosa-Manflas area, in-situ, tends to suggest either that two components of magnetisation have not been separated during demagnetisation, or that the localities have been affected by small-scale structures, possibly associated with displacement along the La Ternera Fault system, that may represent a localised component of deformation. This will be discussed further with regard to the overall pattern of rotation recorded by the La Ternera Formation in the La Guardia area.
La Ternera Fm.
Sampling Area MDs
*in situ*

- Lower
- Upper

Fundo Santa Rosa
& Manflas

Banderitas

Rio Aguas Blancas

60Ma

200Ma

La Ternera Fm.
Sampling Area MDs
tilt corrected

Figure 5.3-10 Stereonet projection of the overall mean magnetisation directions from the four sampling localities in the La Ternera Formation. A) In situ. B) Tilt corrected.
5.4 The Quebrada Monardes and Lagunillas Formations

5.4.1 Stratigraphic relations and sampling

Mid-Jurassic to early Cretaceous continental red sandstones of the Lagunillas and Quebrada Monardes Formations crop-out extensively north and east of the Copiapó river and were sampled along the Rio Jorquera and Figueroa valleys (Table 5.4-1, Figures 5.1-3 & 5.4-1a). Outcrops are NE striking, typically forming thrust delimited packages in the east of the area and represent the remains of a N-S elongated back-arc basin. Originally this thick, continental, red volcanioclastic sequence was considered to represent a single continuous sequence of terrigenous origin (e.g. Bell, 1989) derived from the Jurassic-Cretaceous arcs to the west. They were consequently treated as a single formation for palaeomagnetic analysis by Riley et al., (1993).

More recently Iriarte et al., (1999) have divided this sequence into two formations. The Lagunillas Formation of mid-Jurassic age comprising a lower member of reddened conglomerates and sandstones (Cocambico member) and an upper predominantly volcanic member (Penasco Largo member-Figure 5.1-4). The younger (upper Jurassic-early Cretaceous) Quebrada Monardes Formation is defined as those red-beds deposited above the upper Jurassic lavas of the Penasco Largo member of the Lagunillas Formation or the stratigraphically equivalent lavas of the Quebrada Vicunita Strata (Figure 5.1-4). (A further (younger) sequence of red-beds was also identified as the Quebrada Seca Formation, which is of late Cretaceous age).

Our field strategy therefore aimed to identify these stratigraphic units in the field rather than simply sample "red-beds". In some areas the Penasco Lago member of the Lagunillas Formation is either much reduced in thickness, or not mapped as
Table 5.4-1 Site location, lithology and summary of sample demagnetisation behaviour from the Jurassic Lagunillas Formation. (1*)-Sampling Localities from the study of Riley et al., (1993).

<table>
<thead>
<tr>
<th>Site</th>
<th>Lat (°S)</th>
<th>Long (°E)</th>
<th>Lithology</th>
<th>Tilt Correction</th>
<th>(^\circ)C</th>
<th>ChRM isolation</th>
<th>mT</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT3-05a</td>
<td>27.74</td>
<td>290.34</td>
<td>Med.-fine grained red sandstone</td>
<td>230/52</td>
<td>430-520</td>
<td>560-670</td>
<td>-</td>
</tr>
<tr>
<td>AT3-05b</td>
<td>27.74</td>
<td>290.34</td>
<td>Med.-fine grained red sandstone</td>
<td>225-230/51-57</td>
<td>150-460</td>
<td>560-670</td>
<td>-</td>
</tr>
<tr>
<td>AT3-05c</td>
<td>27.74</td>
<td>290.34</td>
<td>Med.-fine grained red sandstone</td>
<td>225/51</td>
<td>NRM-600</td>
<td>600-620</td>
<td>-</td>
</tr>
<tr>
<td>AT3-05d</td>
<td>27.74</td>
<td>290.34</td>
<td>Med.-fine grained red sandstone</td>
<td>225/46</td>
<td>NRM-250</td>
<td>670</td>
<td>-</td>
</tr>
<tr>
<td>AT3-05e</td>
<td>27.74</td>
<td>290.34</td>
<td>Med.-fine grained grey sandstone</td>
<td>230/45</td>
<td>150-232/51-57</td>
<td>430-640</td>
<td>-</td>
</tr>
<tr>
<td>AT3-09</td>
<td>27.76</td>
<td>290.48</td>
<td>Laminated fine grained sandstone</td>
<td>212/27</td>
<td>200-520</td>
<td>600-640</td>
<td>-</td>
</tr>
<tr>
<td>AT3-10</td>
<td>27.68</td>
<td>290.48</td>
<td>Laminated fine grained sandstone</td>
<td>219/15</td>
<td>560-600</td>
<td>670</td>
<td>-</td>
</tr>
<tr>
<td>AT3-11</td>
<td>27.67</td>
<td>290.48</td>
<td>Laminated fine grained sandstone</td>
<td>259/24</td>
<td>150-350</td>
<td>460-670</td>
<td>-</td>
</tr>
<tr>
<td>90qm14 (1*)</td>
<td>27.66</td>
<td>290.55</td>
<td>Red Sandstone</td>
<td>055/50</td>
<td>420</td>
<td>630</td>
<td>-</td>
</tr>
<tr>
<td>90qm18 (1*)</td>
<td>27.65</td>
<td>290.56</td>
<td>Red Sandstone</td>
<td>290/23</td>
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<td>-</td>
</tr>
<tr>
<td>90qm19 (1*)</td>
<td>27.64</td>
<td>290.52</td>
<td>Red Sandstone</td>
<td>019/14</td>
<td>350</td>
<td>610</td>
<td>-</td>
</tr>
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<td>290.42</td>
<td>Red Sandstone</td>
<td>015/40</td>
<td>550</td>
<td>650</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 5.4-2 Site mean characteristic directions and overall mean direction from the Jurassic Lagunillas Formation. (1*) Sampling Localities from the study of Riley et al., (1993).
Figure 5.4-1  A-Map illustrating the location of sampling sites from the Lagunillas and Quebrada Monardes Formations. Geology and satellite image as displayed in Figure 5.1-2 with the dark green/blue area corresponding to the Lagunillas/Quebrada Monardes Formation. B- Photograph illustrating the conformable contact between the pale cream beds belonging to the marine Lautaro Formation (bottom right & foreground) and the continental red beds belonging to the Lagunillas Formation. Taken along the Rio Jorquera valley with view to the north.
being present (Figure 5.1-3). In these areas the red sandstones of the Cocambico member and the Quebrada Monardes Formation present a continuous inseparable sequence and were therefore avoided in such circumstances. In the east, the Lautaro Formation provides a convenient marker horizon such that the lower Cocambico Member rests conformably on the conspicuous pale-cream coloured limestone beds (Figure 5.4-1b) and hence it was therefore straightforward to identify the red beds of the Lagunillas Formation for sampling in this area.

5.4.2 Cocambico Member of the Lagunillas Formation

The Cocambico Member was sampled at four sites along the Rio Jorquera and Rio Figueroa valleys (Figure 5.1-3, Table 5.4-1). Site AT3-05 is exceptional in that 45 samples were collected from a continuous ~50m thick sequence of well bedded units with the aim of determining whether the section might record magnetostratigraphic changes. This section lies immediately east of Caldera Jorquera (Figure 5.1-3), with samples coming from the lower to middle part of the Cocambico member (the equivalent strata being well exposed (Figure 5.4-1) if a little inaccessible, on the opposite side of the Jorquera river valley). The remaining three sites were sampled towards the top of the section, a little below the lavas of the Penasco Largo member.

Lithologies in site/section AT3-05 range from fine-grained muddy sandstones to coarse-grained volcaniclastic lithic arkoses and vary in colour from cream/grey via purple to red, with red pigments observed to coat the dominantly angular and often poorly sorted clasts in thin section. Sites AT3-09 to 11, are well-sorted lithic-arenites, stained bright red, but with distinctive bedding parallel laminations of unstained medium grained quartziferous bands. All of the Lagunillas sandstones react with HCL, indicating the presence of carbonate, but samples from the base of the formation react most vigorously. This carbonate may either be a secondary
cement associated with the influx of meteoric or magmatic fluids, or could be primary in origin as these sandstones are interpreted as representing transitional near shore deposits of a sabkha-littoral (beach) environment e.g. Bell (1989).

5.4.21 Magnetic Mineralogy of the Lagunillas Formation

The sandstones have very low bulk susceptibilities and k-T experiments generally demonstrate reversible heating & cooling curves, indicating that only very minor alteration occurs during heating (Figures 5.4-2a & b). In the majority of samples, susceptibility is typically observed to decrease between ~550-700°C, with a Curie (Neel) temperature just below 700°C (Figures 5.4-2a & b), suggesting that, not unexpectedly, haematite is the dominant magnetic mineral present.

The dominance of haematite is also reflected in IRM acquisition experiments, where the majority of remanence is acquired in fields in excess of 100mT, indicating that lower coercivity minerals (such as magnetite) contribute relatively little to the induced remanence (Figures 5.4-2a & b). None of the samples subjected to rock magnetic experiments were observed to approach saturation in the highest applied fields and thermal demagnetisation indicates that a composite IRM is dominated by the hard and intermediate components, which often are not fully removed by treatment to 670°C. This would suggest either that the oven has not actually reached the set temperature, or that particularly thermally resistant (exceptionally fine grained, single domain) haematite is present (Figures 5.4-2a & b).

A small number of samples show a variation on the above in that they do appear to contain appreciable amounts of lower temperature/lower coercivity magnetic minerals. These samples show, for example, significant thermo-chemical alteration during heating (Figure 5.4-2c) including a distinct increase in bulk susceptibility
Figure 5.4.2 Magnetic microscopy experiments on representative samples from the Languedoc formations. A & B: Hematite-dominated samples. C: Samples with appreciable magnetite.

- **Intensity of Magnetisation (mA/m)**
- **Normalised Magnetisation**
- **Normalised Susceptibility**
- **Unblocking Spectrum**
- **Acquisition Spectrum**

**A**
- Hematite-dominated Sandstone Sample

**B**
- Hematite-dominated Sandstone Sample

**C**
- Magnetite Magnetochemistry Sample with appreciable Magnetite

---

**Legend:**
- A170204
- A170103B
- A170103B
between 350-500° which then drops between 500-580°C, indicating a Curie point at ~540°C (inflexion point on heating curve) consistent with titanomagnetite of intermediate Ti content [e.g. Hunt et al., 1995]. Haematite is still observed to be present, indicated by the prolonged 'tail' of heating curves extending to temperatures approaching 700°C (Figure 5.4-2c).

IRM acquisition/demagnetisation experiments show that haematite remains the dominant magnetic mineral in such samples however although an increased amount of remanence is acquired in fields <100mT (Figure 5.4-2c). Again the hard and intermediate components dominate the composite IRM, but a significant soft (50mT) component is clearly destroyed by treatment to ~580°C, consistent with the presence of (titano)magnetite. Generally speaking, those samples with apparently more abundant magnetite are typically much less reddened than the remaining samples, being more yellow to grey in colour.

5.4.2ii Demagnetisation Behaviour of the Lagunillas Sandstone Samples

The samples record complex magnetisation histories with multiple components of magnetisation that often overlap substantially. Despite this and the fact that average site NRM intensities were generally low (<5 mA/m) virtually all samples yielded consistent vectors of magnetisation. In most cases the ChRM direction was recognised either as a linear segment on a Zijderveld plot or from demagnetisation paths that were amenable to interpretation using great-circle analysis. Generally only thermal demagnetisation was effective in destroying remanence and therefore this was the demagnetisation technique of choice.

Demagnetisation revealed that the NRM generally consisted of one or more, low to intermediate temperature components, in addition to the ChRM, which itself often overlapped, an albeit poorly constrained, high temperature direction. Variation in
the demagnetisation behaviour of the sandstones largely reflects the degree of overlap between these differing components of magnetisation overlap and the polarity of the principle ChRM and high temperature components.

During demagnetisation behaviour (Figure 5.4-3) the first demagnetisation step (100°C) often removes a very small component of magnetisation, which is widely scattered between samples and is therefore likely to represent a short-term VRM, acquired post sampling. The first consistent component of NRM isolated (normally between 100°C and 250-350°C) typically corresponds to the present day field direction (as indicated in the inset of each of the Zijderveld plots illustrated (geographic coordinates-Figure 5.4-3). The present day (or at least unrotated normal polarity) nature of this low-intermediate component has a mean (in situ) direction of $D=347.7^\circ$, $I=-38.8^\circ$ ($\pm 4.6^\circ$) (Figure 5.4-4a).

Above 350°C, a variety of behaviours associated with the components of magnetisation, carried by what is interpreted to be separate magnetite and haematite fractions, are observed. Often a single component of remanence is observed to be carried at temperatures in excess of 350°C (Figure 5.4-3a) or the magnetisation is seen to be static up to temperatures of ~540°C before the remanence is fully unblocked at higher temperatures (Figure 5.4-3b). This behaviour is attributed to magnetite and haematite carrying the same/parallel components of magnetisation, suggesting they probably acquired their magnetisation simultaneously.

The ChRM direction is not always so clearly defined and magnetite and haematite are interpreted to retain different components of magnetisation with overlapping thermal unblocking spectra, with neither component constrained by a stable magnetisation endpoint (Figures 5.4-3c & d). After removal of the present day direction (which often overlaps with the ChRM remanence carried by magnetite -
Figure 6.4-3  Representative demagnetisation plots for typical Thermal demagnetisation behaviour of samples collected from the Jurassic Lagunillas Formation in the La Guardia Area. A & B) Parallel magnetite and haematite components of magnetisation. C-E) Antiparallel magnetite and haematite components of magnetisation. F) Unexpected 'intermediate' directions.
Figures 5.4-3c & d), the magnetisation vector becomes static in the lower hemisphere after treatment to higher temperatures, indicating a reverse polarity direction. A magnetite fraction is clearly removed by treatment to 540-580°C, with an increase in intensity suggesting that an anti-parallel component of magnetisation is recorded at the highest temperatures i.e. a poorly defined normal polarity is recorded by haematite.

The haematite carried component of magnetisation was never fully resolved, due either to a substantial amount of remanence not being unblocked (despite sufficient experimental time and calibration of the oven –see Chapter 2), or to directional instability at the highest temperature demagnetisation steps (Figures 5.4-3c & d). It may be possible that the haematite fraction was exceptionally fine-grained and perhaps more resistant to thermal demagnetisation such that even longer hold times were required in the oven than were actually used.

In Figure 5.4-3e the magnetite and haematite components are clearly separated, if not fully resolved. A significant reversal in the observed polarity is observed after treatment to 560°C, with the magnetite component in the lower (reverse polarity) hemisphere and the haematite component in the upper hemisphere. Again the magnetite direction is not defined by a stable endpoint, with the haematite component also not fully destroyed and the final demagnetisation steps are observed to be static about a single point. The magnetite and haematite components in this example are shown to carry antiparallel directions of magnetisation (as suggested by Figures 5.4-3c & d), suggesting they were not acquired simultaneously.

Finally anomalous intermediate temperature directions that cannot be reasonably explained as a record of the ancient magnetic field (Figure 5.4-3c) but may result from the simultaneous removal of two non-parallel components of magnetisation
Figure 5.4-4 Steronet projections of components of magnetisation isolated from the Lagunillas sandstones that do not represent the ChRM direction. A) present day (or unrotated normal polarity) field directions and B) NW orientated 'intermediate' directions that cannot be explained by field directions expected from the La Guardia area.
were also observed. An example of such a clearly defined but anomalously orientated linear demagnetisation path is shown in Figure 5.4-3f, which illustrates the removal of a NW orientated vector of magnetisation between 300-520°C. The NW (or SE) orientation of this 'intermediate' direction is commonly observed in samples from the Lagunillas Fm. (Figure 5.4-4b). Whatever the actual nature of these directions, they are not considered to represent the ChRM direction and are therefore not believed to be of tectonic significance.

5.4.2iii Summary of Demagnetisation Data from the Lagunillas Formation

Although site AT3-05 was collected with magnetostratigraphic intentions the complex behaviour is such that it is better considered more simply in terms of isolating reliable remanence directions. In order to weight the data from site AT3-05 in the overall mean direction calculated for the Lagunillas Formation, individual 'site' means were calculated by subdividing Site AT3-05 into five ~10 m sections, so that similar numbers of samples were included in all site means (Table 5.4.2).

The demagnetisation behaviour of the Lagunillas sandstones indicates the ChRM direction is carried by a combination of magnetite and haematite, either of the same (Figures 5.4-3a & b) or opposing polarities (Figures 5.4-3c-e). The thermal unblocking spectra of site mean ChRM directions from this study are consistent with those established in four sites by Riley et al., (1993-Chapter 5.2, Table 5.4-2), who state that the ChRM isolated from these samples is unblocked across a similar range of temperatures, typical of magnetite and haematite. The mixed polarity of the ChRM indicates an acquisition period spanning at least one reversal of the Earths' magnetic field. If the isolated magnetisation carried by magnetite is interpreted to represent a primary DRM, the development of secondary haematite may have occurred at a later time, within anti-parallel magnetic fields.
The formation mean direction is very well defined, using magnetite and haematite borne components of magnetisation discussed above (Table 5.4-2, Figure 5.4-5), with normal (8 sites) and reverse (4 sites) polarity directions passing both a reversal test (with C-classification—MacFadden & McElhinny, 1990) and a fold test. In addition, the tilt corrected inclination is consistent with the expected inclination for the sampling latitude (Table 5.4-2, Figure 5.4-5a). This indicates that the ChRM direction is a pre-tilting (Primary) magnetisation.

Although the highest temperature (haematite carried) component of magnetisation was present in most sites it can only be well defined from site AT3-05 (Table 5.4-2, Figure 5.4-5b). This high temperature mean direction is almost exactly antiparallel to the ChRM direction discussed above, suggesting that both components are of pre-tilting origin (Figure 5.4-5b). Assuming the characteristic and high temperature directions represent a pre-tilt DRM carried by magnetite (± haematite) and an early secondary CRM haematisation of the sandstones, the anti-parallel components of magnetisation are interpreted to have been acquired within a relatively short period (geologically).

The ChRM magnetisation direction is $D=205.5^\circ$, $I=49.8^\circ$, with an $\alpha_{95}$ error of 9.5° (Table 5.4-2, Figure 5.4-5a) and will be discussed further in a tectonic context.

5.4.3 Quebrada Monardes Formation

The Quebrada Monardes Formation was sampled at seven locations, in two areas located in the northeast quadrant of the mapsheet [Iriarte et al., 1999] (Figure 5.1-3, Table 5.4-3). Samples were collected from the base of the formation (Sites AT3-06, 07 & 08), where the rocks are well lithified, but often highly fractured. The sediments are generally well-sorted, fine-grained, grey to reddish brown muddy sandstones with lamination and cross bedding or more massive beds, with no
Figure 5.4-5 Steronet projections of A) Site mean directions and overall mean ChRM direction calculated for the Lagunillas sandstones, and B) high temperature directions isolated from individual samples and overall mean high temperature direction. The high temperature component is almost exactly anti-parallel to the ChRM direction (yellow), suggesting that both components were acquired prior to CW rotation.
Table 5.4-3  Site location, lithology and summary of sample demagnetisation behaviour from the late Jurassic-early Cretaceous Quebrada Monardes Formation.

<table>
<thead>
<tr>
<th>Site</th>
<th>Sampling Site</th>
<th>Lat ('S)</th>
<th>Long ('E)</th>
<th>Lithology</th>
<th>Tilt Correction</th>
<th>°C Min.</th>
<th>°C Max.</th>
<th>ChRM isolation</th>
<th>mT Min.</th>
<th>mT Max.</th>
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<td></td>
<td>27.69</td>
<td>290.45</td>
<td>Med. grained dark red sandstone</td>
<td>196/36</td>
<td>100</td>
<td>640-670</td>
<td></td>
<td></td>
<td></td>
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<td>AT3-07</td>
<td></td>
<td>27.69</td>
<td>290.45</td>
<td>V. fine grained sandstone</td>
<td>210/30</td>
<td>300-450</td>
<td>640-670</td>
<td></td>
<td></td>
<td></td>
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<td>290.45</td>
<td>Med.-fine grained red sandstone</td>
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<td>640-670</td>
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<td></td>
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<td></td>
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<td>Fine grained red sandstone</td>
<td>243/16</td>
<td>NRM-520</td>
<td>460-670</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 5.4-4  Site mean characteristic directions and overall mean direction from the late Jurassic-early Cretaceous Quebrada Monardes Formation.
internal structure, interbedded with much finer grained (silty) red-brown material of a friable nature. Samples were also collected from the upper part of the formation just below the overlying Quebrada Seca Formation (Sites AT3-12 to AT3-15, Figure 5.1-3). These beds were very similar to those observed at the base, but had a more pinkish-brown general colouration. In places, lags of coarser gravel sized material were observed, often set in a much paler coloured matrix/cement that reacted violently with HCl, indicating the presence of calcium carbonate.

Thin-section analysis of a range of samples revealed calcite infilling voids/pore spaces, and the main constituents are angular to sub-angular plagioclase feldspar, andesitic lithic fragments and relatively small amounts of quartz grains, hence these sandstones are arkosic arenites. The coarsest units were clast supported with long axes generally parallel to bedding. In addition clasts were coated with a red iron oxide (haematite?) pigment, which also formed opaque patches between grains/clasts.

5.4.3i Magnetic Mineralogy of the Quebrada Monardes Formation

The magnetic mineralogy is very similar to that observed for the Lagunillas sandstone, except for a more variable behaviour depending upon the amount of magnetite present. The rock magnetic behaviour of most samples is consistent with haematite being the dominant magnetic carrier (Figure 5.4-6a). High temperature susceptibility experiments on powdered samples indicate the bulk susceptibility of most samples is very low (often at or near the noise level of the KLY-3 kappabridge), susceptibility was generally only seen to decrease between temperatures of 600-700°C, consistent with the Neel temperature of haematite. Haematite dominated samples acquire only very limited IRM's in applied fields of <100mT, and fail to saturate in the maximum fields available, with coercive fields of 300mT or more required to reverse the initially imparted IRM (Figure 5.4-6a).
Figure 5.4-6 Magnetic mineralogy experiments on representative samples from the Lagunillas Formation. A) Haematite dominated samples, B) Magnetite & Haematite mineral assemblages C) Magnetite dominated samples.
Thermal demagnetisation of a three component IRM shows that whilst a small intermediate coercivity component was removed between 100-300°C, possibly indicating that a small amount of pyrrhotite is present, the hard and intermediate components were progressively destroyed between 300-620°C and are completely unblocked by treatment to 670°C. Although a very small soft coercivity (50mT) component was observed and removed by treatment to ~500°C, the magnitude of this component demonstrates the subordinate nature of magnetite with respect to haematite in these samples (Figure 5.4-6a).

Some samples exhibited more significant quantities of magnetite (in addition to haematite), with high temperature susceptibility experiments indicating a Curie temperature at ~580°C and IRM experiments displaying slightly lower coercivity behaviour. Samples containing more significant quantities of magnetite acquire magnetisation more rapidly in fields up to 100mT, although still fail to saturate in the highest applied field and required coercive fields of ~300mT to reverse the initial IRM (Figure 5.4-6b). The increased amount of magnetite was also evident during three-component IRM analysis, with a pronounced soft component unblocked at 570°C. The hard component of IRM still dominated the overall magnetisation with a lesser intermediate component both destroyed on treatment to 670°C (Figure 5.4-6b). Rock magnetic studies therefore indicate that haematite is still the dominant constituent of the magnetic mineral assemblage for these samples.

A very small number of samples, collected mainly from the coarsest grained, yellow coloured, sandstones from the top of the Quebrada Monardes Formation, display magnetite dominated mineralogies, with haematite as the subordinate magnetic mineral. Thermomagnetic experiments showed a pronounced peak between ~300-500°C during heating, that was absent during cooling, suggesting
that the samples underwent thermochemical alteration (Figure 5.4-6c). The reason for this peak is unclear. Although magnetite apparently dominates these samples they did not saturate in the maximum available field (800mT) but do show substantial acquisition of magnetisation in applied fields up to 100mT (Figure 5.4-6c). Composite IRMs are dominated by the soft and intermediate components, which are destroyed by thermal demagnetisation at 580°C while the smaller magnitude hard component is destroyed by treatment to 670°C. In summary this behaviour is consistent with magnetite as the principal carrier of remanence but with a small amount of haematite also present.

5.4.3ii Demagnetisation Behaviour of the Quebrada Monardes Sandstones

Thermal demagnetisation was the preferred demagnetisation method for the majority of samples. Most samples were relatively weakly magnetised with total NRM intensities typically in the range 5-10 mA/m (25 mA/m max), and initial bulk susceptibility ranged between 100-500 x 10^{-6} SI units, as might be expected for haematite dominated (sedimentary) samples. Neither the initial intensity nor bulk susceptibility, were obviously correlated to site location, general lithology, grain size, colour or the presence/absence of carbonate.

Thermal demagnetisation identified a number of components of magnetisation contributing towards the overall NRM, with each component carried by a magnetic fraction that is unblocked across a characteristic temperature range. These components are identified as follows:

1. a low temperature (<150°C) VRM magnetisation of random direction,
2. a low-intermediate (100-400°C) temperature direction (Figure 5.4-7-light grey shading on intensity plots) and typically in the present day direction,
Figure 5.4-7 Representative demagnetisation plots for typical Thermal demagnetisation behaviour of samples collected from the late Jurassic-early Cretaceous Quebrada Monardes Formation in the La Guardia Area. A-C) Parallel magnetite and haematite components of magnetisation. C-F) Antiparallel magnetite and haematite components of magnetisation. Shades of grey on intensity plots indicate the unblocking temperatures of the various components of magnetisation contributing towards the overall NRM.
3. the ChRM between 400-620°C, spanning both magnetite and haematite ranges (Figure 5.4-7-mid grey shading on intensity plots), and
4. a high temperature component, corresponding to a direction carried purely by haematite (up to 670°C), that may or may not relate to the ChRM (Figure 5.4-7-dark grey shading on intensity plots).

Although each of these components was probably present in the majority of the specimens demagnetised, the overall demagnetisation behaviour is determined by the polarity and relative intensity of each component. Two generalised behaviours were identified: the more straightforward comprised univectorial and often origin bound ChRM directions where the ChRM and high temperature components are of the same polarity (Figures 5.4-7a-c); the other behaviour is more complex and consisted of up to three, directionally meaningful², directionally bipolar, segments in their demagnetisation trajectories (Figures 5.4-7d-f).

Most demagnetisation diagrams are dominated by the ChRM component of magnetisation, which is generally unblocked somewhere between 400-620°C, indicating that it is carried by both magnetite and haematite (Figure 5.4-7). This component was usually well grouped within site and predominantly of reverse polarity (Figure 5.4-8b). The observation that magnetite contributes significantly to carrying the ChRM direction, is slightly at odds with the outcomes of the rock magnetic experiments which showed haematite to be more dominant of the measured characteristics. This suggests that haematite is less important in carrying the ChRM remanence. Haematite does however carry a component of

² i.e. the isolated component of magnetisation is considered to represent a genuine record of the ancient (or modern) geomagnetic field and not a short-term secondary VRM or otherwise spurious direction.
Figure 5.4-8 Stereonet projections of A) Low temperature components of magnetisation considered not to represent ChRM in the Quebrada Monardes samples (interpreted to represent a mixture of present day field directions and short term VRMs), and B). ChRM (and high temperature) directions isolated from individual samples and overall mean direction calculated for samples from Site AT3-07 The high temperature component is almost exactly anti-parallel to the ChRM direction (yellow), suggesting that both components were acquired contemporaneously.
remanence that is apparently separate to the ChRM direction, as with the Lagunillas data, suggesting that two haematite fractions exist within the samples, one associated with ChRM and the other carrying the highest temperature direction. Where the haematite component is of the same polarity as the ChRM the two cannot be separated (Figures 5.4-7a-c). Many samples however demonstrate high temperature directions that are clearly antiparallel to ChRM and are therefore more easily identified (Figures 5.4-7d-f) if not quantified. Typically this was recognised as a great-circle demagnetisation path between the two opposing polarity directions at the highest temperature demagnetisation steps (Figure 5.4-7d).

Most of these directions were too poorly resolved in order to constrain reasonable individual site mean directions for this component, but there were sufficient examples from three sites to calculate a mean direction using constrained great circles (Table 5.4-4). An example of this behaviour from Site AT3-07 indicates that the high temperature direction is almost exactly anti-parallel to the characteristic direction carried by magnetite ± haematite (Figure 5.4-8c). As in Fm. Lagunillas the high temperature 'haematite only' component is interpreted to represent a secondary CRM carried by a haematite fraction that developed subsequent to deposition.

5.4.3iii Summary of Demagnetisation Data from the Quebrada Monardes Formation

As with previous studies (Randall, 1996; Riley et al., 1993) the magnetisation history of the Quebrada Monardes Formation was found to be complex with up to four components of magnetisation present. The majority of samples displayed a dominant ChRM direction, which is carried by both magnetite and haematite although often a very high temperature (haematite-only) component was
recognised. In contrast to the Lagunillas Fm., site mean directions of the ChRM Formation clearly fail a fold test, indicating that the Quebrada Monardes Formation has undergone a post tilting remagnetisation (Table 5.4-4, Figure 5.4-9). The fact that the majority of the sites are dominated by reverse polarity magnetisations indicates that the isolated remanence is ancient in origin. The individual site-mean directions are reasonably tightly clustered (Figure 5.4-9), which may suggest that PSV has not been entirely averaged, but the two normal polarity directions pass the reversal test of McFadden & McElhinny (1990) with a classification of C, indicating they were acquired in a closely related, anti-parallel magnetic field, indicating the remagnetisation event affecting the Quebrada Monardes Formation spanned at least a single reversal of the geomagnetic field.

Overall therefore, the ChRM of the Quebrada Monardes sandstones is considered to record a Secondary (post-tilt) magnetisation, with a mean overall direction of \( D=197.7^\circ, I=52.0^\circ \), with an \( \alpha_{95} \) error of 8.1° (Table 5.4-4, Figure 5.4-9). In contrast the high temperature, haematite only, component direction is better clustered after tilt correction (Table 5.4-4) suggesting it pre-dates tilting of the formation. The mean tilt corrected direction of the haematite only component (calculated using all constrained GC's) is \( D=37.3^\circ, I=-53.0^\circ \), with an \( \alpha_{95} \) error of 7.8° (Table 5.4-4, Figure 5.4-9), suggesting a CW rotation ~20° greater than recorded by the ChRM direction.

5.4.4 Discussion of Palaeomagnetic Data from Red Beds in the La Guardia Area

The main issue with using red sediments for palaeomagnetic work is the fact that the timing (and indeed the precise mechanism) of the formation of the red pigment(s) is often unclear, and could occur many millions of years after the sediments were deposited and such magnetisations, may, therefore represent much later secondary CRMs. This implies that magnetisation directions carried by
Figure 5.4-9  Stereonet projections of A) Individual site mean directions and overall direction calculated for palaeomagnetic data from the Quebrada Monardes Formation, and B) Mean high temperature direction calculated from great circle analysis of individual samples from the Quebrada Monardes Formation. The high temperature component is almost exactly anti-parallel to the ChRM direction (yellow), suggesting that both components were acquired prior to CW rotation.
continental red beds, even in the absence of any remagnetisation event as such, may not be related to the orientation of the magnetic field at the time of deposition.

Magnetic mineralogy experiments indicated, that as expected from the distinctive red appearance, haematite is present in all of the samples (Figures 5.4-2 & 5.4-6). Thin section analysis showing the majority of this haematite probably forms the red pigmentary staining of the quartz, feldspar and lithic fragments in the sandstones. Also evident were larger grains of haematite (or altered magnetite), possibly of detrital origin or derived from the oxidation of large detrital (volcanic) grains. Magnetite was also shown to be present in varying amounts possibly reflecting that the majority of the detrital grains/lithics contain very small amounts of magnetite as well as individual, larger, MD grains which do not generally record stable magnetisations as effectively as fine-grained haematite [e.g. Butler, 1992; Turner, 1981].

The sub-sampling of the two formations according to the stratigraphy of Iriarte et al. (1999), as undertaken during the 2003 field season, has shown that in the La Guardia area, the continental sandstones of the Quebrada Monardes Formation are palaeomagnetically distinct from those of the older, underlying Lagunillas Formation. The most notable feature of the palaeomagnetic data from the Quebrada Monardes and Lagunillas Formations is the Primary remanence isolated from the Lagunillas Formation with a positive fold-test indicating a pre-tilt magnetisation is clearly preserved (Figure 5.4-5, Table 5.4-2). In contrast, site mean directions from the younger Quebrada Monardes Formation fail a fold test, indicating that a secondary remanence has been isolated (Table 5.4.4, Figure 5.4-9).

The timing of the growth of the red (haematite) pigment present within the Lagunillas Formation appears to geologically post-date the acquisition of ChRM,
which is interpreted to be carried by a combination of magnetite and haematite. Whilst the highest temperature (haematite-only) direction is interpreted to represent a secondary CRM, the mean direction carried by this component, records a similar amount of rotation to that recorded by the ChRM direction, suggesting it too is likely to represent a pre-tilt magnetisation. A slightly different situation is observed from the Q. Monardes sandstones, where, although the ChRM is clearly secondary in nature, the haematite only direction is more precisely constrained after tilt-correction, suggesting that it may be a pre-tilt magnetisation and pre-date the ChRM direction recovered. Indeed the haematite component isolated from the Q. Monardes samples is similar to the ChRM and haematite directions determined from the Lagunillas sandstones.
5.5 Caldera Jorquera

5.5.1 Palaeomagnetic Sampling Localities

Twelve sites were collected from the predominantly tuffaceous and ignimbritic horizons that comprise the intra-caldera deposits of Caldera Jorquera, along the Rio Jorquera valley (Figures 5.1-3 and 5.5-1, Table 5.5-1). The sampling locations represent a SW-NE transect through the caldera and were restricted by steep topography to be essentially along the main water course.

The pervasive flattening fabric observed within all of the ignimbrite/tuffaceous cooling packages was recorded as a potential marker for palaeo-horizontal, and/or bounding surfaces of lava flows where sampled. The flattening fabrics measured were generally characterised by distinctive fiammè type textures, with the flattened discs usually comprised of larger volcanic (often pumaceous) clasts or volcanic glass.

Thin section analysis of the ignimbrite units sampled indicates that none of the cooling units have undergone significant amounts of welding, with cuspate shards of volcanic glass observed not to have been significantly flattened/fused (see Figure 5.5-2 for explanation), which is in apparent contradiction to the field observation of flattening fabrics being pervasive. This suggests that the majority of the ignimbrites sampled were deposited either at reduced temperatures or were sourced from higher in the volcanic cloud, away from the high shear zone.

Due to the depositional nature of ignimbrite type deposits, these flattening fabrics may have been subjected to an amount of rheomorphism, whereby the preserved fabric does not actually truly represent the original flattening horizon. Fortunately, whilst rheomorphism can re-orientate volcanic particles after initial deposition, this
Figure 5.5-1 Geological map of Caldera Jorquera redrawn from the La Guardia mapsheet [Iriarte et al., 1999], with palaeomagnetic sampling localities indicated. 1-Paleozoic plutons, 2-La Ternera Formation, 3-Lautaro Formation, 4-Q. Vicunita strata, 5-Laguinillas Formation, 6-Q. Monardes Formation, 7(8)-Q. Seca lower (upper) member, 9-Jorquera intracaldera strata, 10-Sierra Los Leones strata, 11- Sierra La Dichosa strata, 12-rhyolitic domes, 13-Eocene magmatic arc, 14- Alluvium, 15- hydrothermal alteration.

<table>
<thead>
<tr>
<th>Site</th>
<th>Sampling Site</th>
<th>Lat (°S)</th>
<th>Long (°E)</th>
<th>Lithology</th>
<th>Tilt Correction</th>
<th>°C</th>
<th>ChRM Isolation</th>
<th>mT</th>
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<td>-</td>
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<td>670</td>
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Table 5.5-1 Site location, lithology and summary of sample demagnetisation behaviour from the Paleocene Caldera Jorquera.
Glass shards with cusps start to deform as welding increases, limbs of glass shards touch, and shards become increasingly flattened and stretched, creating typical flattened discs or Fiamme (referred to as eutaxitic texture—Fisher & Schmincke, 1984).

Figure 5.5-2 Diagram illustrating the origin/identification of welding in pyroclastic rocks. Cuspate glass shards represent side-wall remnants of gas bubbles in pumice fragments. Little welding is indicated by the retention of clearly cuspat e shards within a pyroclastic unit, indicating it was deposited away from the high shear zone at the base of a pyroclastic flow. As welding increases, limbs of glass shards touch and shards become increasingly flattened and stretched, creating typical flattened discs or Fiamme (referred to as eutaxitic texture—Fisher & Schmincke, 1984).

Figure 5.5-3 Diagram illustrating rheomorphic structures developed in pyroclastic units in Gran Canaria (From Fisher & Schmincke, 1984, redrawn after Schmincke & Swanson, 1967)
process occurs at temperatures in excess of 670°C (Neel temperature of haematite). Schminke & Swanson (1967) observed a range of structures in ignimbrites belonging to the Mogan Formation in Gran Canaria, which they interpreted to result from the process of rheomorphism affecting the stratified layers of a volcanic debris flow (Figure 5.5-3) and related these to the current flow direction. It was noted that the basal units of individual cooling units often exhibited a clearly imbricated fabric, while flattened clasts of pumice and volcanic glass were observed to be deposited horizontally above (w.r.t. the clast long axes) in the lower high strain zone. Higher in the flow, other features such as boudinaged clasts and ramp structures are also observed. The observation of such ramp structures really depends on the orientation of the rock-race in which the orientated clasts are viewed. As shown in Figure 5.5-3, ramp structures are only observed side-on or obliquely, with only horizontally oriented clasts viewed in the section perpendicular to the flow direction. No such ramp features were observed in any of the ignimbrite cooling units sampled from Caldera Jorquera, but this could reflect the orientation of the rock sections that were sampled. Away from such features however, the flattening fabric appears to be a reasonable estimate of a palaeohorizontal surface.

Another potential problem with using flattening fabrics as a proxy for bedding in ignimbrites is the likelihood that the flattening plane is also a function of the surface onto which the volcanic cloud was deposited. Towards the margins of Caldera Jorquera, ignimbrite deposits appear to steepen upwards towards the circular boundary fault (Figure 5.5-4). Although the deposits appear to be steeply inclined into the caldera, as indicated by the presence of a steep flattening fabrics suggesting the deposits have been substantially tilted post-deposition, the most likely explanation is that the tilt is the result of the ignimbrites being 'splashed'
Figure 5.5-4 Photograph illustrating the steep dip of intra-caldera ignimbrites towards the extreme NE margin of Caldera Jorquera. Cooling units in the foreground dip steeply towards the centre of the caldera, suggesting they were ‘splashed’ against the caldera margin. The top of the cooling units are therefore unlikely to represent reliable palaeohorizontal estimates. View to the west.
against the steep inner wall of the caldera, as the cooling units appear to flatten out towards the centre of the caldera.

With respect to the possibility that the measured flattening fabric may represent a false palaeohorizontal correction, AMS measurements were undertaken for all of the samples prior to demagnetisation to investigate if the preserved magnetic fabric could more reliably be used to correct for any tectonic tilting that has affected the intra-caldera deposits. However, the measurements of palaeohorizontal measures from the bounding surfaces of the lava flows sampled are considered to represent the most reliable estimate of the post deposition tectonic disturbance of the intra-caldera deposits, and therefore the tilt-corrected directions of the ignimbrite units will be compared to those of the lava flows sampled.

Another ambiguity that was encountered whilst sampling the ignimbrite units was in the determination of discrete cooling packages. In order to adequately average the effect of palaeo-secular variation of the ancient magnetic field, a wide range of separate cooling units needed to be sampled. Many of the larger ignimbrite deposits were laterally traceable over several kilometres within the caldera and care was taken to try to sample as many separate horizons as was possible. However two groups of sites (AT3-04, 51, 52, & 53, and AT3-48, 64 & 66) may represent only two individual units (Figure 5.5-5), and this possibility is considered further in the analysis by comparing the directions recorded.

5.5.2 Magnetic Mineralogy

Thermomagnetic experiments indicate that the majority of samples display a marked drop in bulk susceptibility at ~580°C, indicating magnetite was the dominant magnetic mineral (Figure 5.5-6a). In addition the heating curve (Figure 5.5-6a) is clearly irreversible, with a distinct increase observed during heating to
Figure 5.5-5  Photograph illustrating the relationship between three cooling units sampled from with Caldera Jorquera (only one was sampled at this locality-Site AT3-48 (unit 1 in photos)). These cooling units were tentatively traced to some extent throughout the caldera, with attempts not to sample the same unit in multiple locations, but may have been sampled more than once. The upper surface of each cooling unit appears to well defined, with successive units apparently onlapping on to the next, suggesting they belong to separate explosive events. The possibility remains that they may represent a single cooling unit however. A) View to the NW (truck for scale). B) View of the same units from the south, towards the NE.
Figure 5.5-6 Magnetic mineralogy experiments on representative pyroclastic & lava samples from Caldera Jorquera. A) Magnetite dominated ignimbrite, B) Pyrrhotite dominated ignimbrite, C) Altered andesitic lava.
150-350°C. This would suggest that either maghemite or pyrrhotite, are also present and likely to have been thermochemically altered during heating. Several samples showed the loss of susceptibility over a much wider range of temperatures, possibly indicating that magnetite is not the dominant magnetic mineral within these samples (Figure 5.5-6b). High temperature ‘tails’, where the observed susceptibility is not reduced to ~0 until somewhere between 650-700°C, indicates the presence of small amounts of (titano-) hematite is present. This may be in the form of either a primary constituent of the initial lava flow, or a secondary high temperature oxidation product originally derived from the alteration of iron-rich minerals.

IRM acquisition and backfield coercivity experiments indicate that magnetite and/or another mineral dominate the magnetic mineral assemblage of most samples, the majority of which are fully saturated in applied fields up to 300mT (Figure 5.5-6a), with reverse fields of between 45-60mT required to remove the initial IRM. Those samples displaying a drop in bulk susceptibility over broader temperature ranges also display slightly higher coercivity behaviour, with samples not completely saturated in the maximum applied fields. Whilst this could be due to increased amounts of haematite, results from the thermomagnetic experiments indicate that this likely to be due to the presence of pyrrhotite, which saturates in fields of in excess of 3T [e.g. Lowrie & Heller, 1982], but has a Curie temperature of 325°C [e.g. Hunt et al., 1995-Figure 5.5-6b].

The presence of pyrrhotite (in all pyroclastic samples) is confirmed by the thermal demagnetisation of a three-component IRM which indicates that whilst a very small hard (800mT) component is removed by treatment to 670°C, a pronounced component of intermediate coercivity is unblocked at ~300°C suggesting pyrrhotite is indeed present in some quantity (Figure 5.5-6a & b).
The two (altered) lava flows (sites AT3-02 and AT3-65) sampled from within Caldera Jorquera exhibit simple magnetite behaviours, with a small pyrrhotite component also possibly identified during thermal demagnetisation of a composite IRM. Examples of this behaviour are shown in Figure 5.5-6c, but are not discussed further due to the limited number of lavas sampled.

5.5.4 Demagnetisation Behaviour

_Ignimbrites_

The ignimbritic samples collected from Caldera Jorquera, generally demonstrate a reasonably straightforward demagnetisation behaviour, with the removal of the characteristic remanence accompanied by very little 'noise' on application of AF fields up to 100 mT, or up to temperatures of between 580-670°C. Prior to the isolation of the ChRM, the initial demagnetisation step, using either AF or Thermal demagnetisation, generally removes a small component of VRM, followed by the removal of a more persistent low temperature component that is thermally unblocked by treatment to 300°C (light grey shading on intensity plots-Figures 5.5-7), but is not removed during AF demagnetisation. The orientation of this direction is often subtly different to that of the overall ChRM and is considered likely to reflect a component of magnetisation carried by pyrrhotite, due to its thermal unblocking spectrum and apparent resistance to AF treatment.

In the majority of samples the ChRM direction is often recognised between temperatures of 300-640°C and is completely destroyed by treatment 670°C (Table 5.5-1, Figures 5.5-7a & b-dark grey shading on intensity plot). The typical demagnetisation behaviour is essentially univectorial and is of predominantly normal polarity (although reverse polarity directions are also observed), with both magnetite and hematite carrying the same direction of magnetisation (Figure 5.5-
Figure 5.5-7 Representative Zijderveld plots for typical Thermal and AF demagnetisation behaviour of samples collected from intracaldera pyroclastic units and lava flows from the Paleocene Caldera Jorquera, in the La Guardia Area. A & B) A-type behaviour characterised by a univectorial and origin bound ChRM direction, carried by magnetite and haematite. C & D) B-type behaviour characterised by dual polarity remanence that demagnetises along a great circle path with a normal polarity (ChRM) remanence carried by magnetite and high temperature, reverse polarity remanence carried by haematite. E-F) Univectorial remanence carried by (altered) andesitic lava flows, carried predominantly by magnetite only.
7a & b). This is termed the A-Type remanence and is probably carried SD or ti-
poor magnetite, as indicated by the reasonably hard AF demagnetisation curve
(Figure 5.5-7a) and narrow temperature range over which the majority of
remanence is actually unblocked (520-560°C, mid grey shading-Figure 5.5-7b).

Samples from site AT364, collected from an ignimbrite unit from the northeast
quadrant of the caldera, (Figures 5.5-7c & d), appear to record a dual polarity
remanence as demagnetisation paths of these samples follow great circles from
an initially normal to reverse polarity direction with successive demagnetisation
steps. The reverse polarity component becomes most evident after treatment to
540°C and above, which removes the dominant normal polarity direction, similar to
the range noted for A-type behaviour. This suggests that the reverse polarity
component is dominantly carried by hematite. A substantial component of
magnetisation is carried by pyrrhotite, which is unblocked by treatment to 300°C
(Figure 5.5.7d). AF demagnetisation is far less effective at resolving this second B-
type demagnetisation behaviour, although some evidence of it is apparent (Figure
5.5.7c).

Lavas

The two andesitic lava flows were shown to differ only slightly from the A-type
behaviour identified for the ignimbrites, although magnetite is the sole carrier of
remanence and remanence was often demagnetised on application of AF fields of
as little as ~40 mT, above which the magnetisation becomes unstable (Figure 5.5-
7e). Again thermal magnetisation indicates that a small component of remanence
is removed by treatment to 300°C indicating that pyrrhotite contributes towards the
initial NRM (Figure 5.5-7f). Thermal demagnetisation reinforces that the ChRM is
carried exclusively by magnetite, with the sample being fully demagnetised by
treatment to 580°C (Figure 5.5-7f). The ChRM is unblocked evenly when treated to
temperatures between of ~300-580°C, which may indicate that magnetite of either a range of grain sizes or ti-compositions carry remanence. It is notable however that the site average direction of magnetisation is similar to that calculated for the rest of the sampling localities, indicating that the differing magnetic mineralogy has no effect on the actual direction recorded.

5.5.3 Anisotropy of Magnetic Susceptibility

The anisotropy of magnetic susceptibility was measured for each of the sampling localities within Caldera Jorquera, in order to establish whether or not a primary (palaeohorizontal) fabric could be identified, as a check on the mesoscopic flattening fabrics measured in the field for palaeohorizontal control.

All of the samples measured exhibit less than 5% AMS (Table 5.5-2), which suggests that all of the samples can be considered to have recorded the geomagnetic field with high fidelity. The majority of samples at the majority of sites (~60%) are characterised by oblate (flattened) susceptibility ellipsoids (Figure 5.5-8a), with many of the prolate (cigar shaped) magnetic ellipsoids interpreted as individual outliers within the overall site pattern. Whilst an oblate (flattened) fabric would be expected, the observation of prolate ellipsoids may result from the fact that individual 1" samples may be too small relative to the size of some of the fiamme and lithics to have adequately measured the overall pattern of flattening observed at sites as a whole in the field.

The mean principal susceptibility axes and the measured mesoscopic flattening fabric for each site were compared (Table 5.5-2, Figure 5.5-8b). If the microscopic AMS fabric reflects the mesoscopic flattening fabric measured in the field, the pole to the (mesoscopic) flattening plane would be expected to coincide with the $k_3$ (or $k_{\min}$) axis determined from AMS measurements. From plots of the site mean
Table 5.5-2  Site mean AMS parameters calculated for the Paleocene Caldera Jorquera.

<table>
<thead>
<tr>
<th>Site</th>
<th>Mesoscopic Fabric</th>
<th>Magnetic Foliation (k1-k2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT3-02</td>
<td>226/22</td>
<td>330/17</td>
</tr>
<tr>
<td>AT3-03</td>
<td>173/15</td>
<td>079/12</td>
</tr>
<tr>
<td>AT3-04</td>
<td>005/13</td>
<td>-</td>
</tr>
<tr>
<td>AT3-48</td>
<td>197/20</td>
<td>184/28</td>
</tr>
<tr>
<td>AT3-51</td>
<td>192/20</td>
<td>194/51</td>
</tr>
<tr>
<td>AT3-52</td>
<td>200/30</td>
<td>201/83</td>
</tr>
<tr>
<td>AT3-53</td>
<td>180/20</td>
<td>-</td>
</tr>
<tr>
<td>AT3-63</td>
<td>085/22</td>
<td>059/25</td>
</tr>
<tr>
<td>AT3-64</td>
<td>189/32</td>
<td>-</td>
</tr>
<tr>
<td>AT3-65</td>
<td>204/30</td>
<td>169/48</td>
</tr>
<tr>
<td>AT3-66</td>
<td>168/26</td>
<td>017/61</td>
</tr>
</tbody>
</table>

Table 5.5-3  Orientation of flattening fabric measured in the field and magnetic foliation determined from AMS studies of the intracaldera pyroclastic/volcanic units from the Paleocene Caldera Jorquera.

<table>
<thead>
<tr>
<th>Site</th>
<th>n/N</th>
<th>Dec. (°)</th>
<th>Inc. (°)</th>
<th>Tilt-corrected Dec. (°)</th>
<th>Inc. (°)</th>
<th>k</th>
<th>a95 (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT3-02</td>
<td>5/5</td>
<td>032.0</td>
<td>-63.9</td>
<td>075.1</td>
<td>-60.8</td>
<td>71.7</td>
<td>9.1</td>
</tr>
<tr>
<td>AT3-03</td>
<td>6/6</td>
<td>051.0</td>
<td>-47.8</td>
<td>057.3</td>
<td>-34.6</td>
<td>642.7</td>
<td>2.6</td>
</tr>
<tr>
<td>AT3-04</td>
<td>6/6</td>
<td>024.0</td>
<td>-58.4</td>
<td>002.0</td>
<td>-60.3</td>
<td>113.0</td>
<td>6.3</td>
</tr>
<tr>
<td>AT3-48</td>
<td>8/8</td>
<td>022.6</td>
<td>-61.1</td>
<td>052.6</td>
<td>-53.8</td>
<td>222.3</td>
<td>3.7</td>
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<tr>
<td>AT3-51</td>
<td>7/7</td>
<td>031.9</td>
<td>-52.0</td>
<td>050.9</td>
<td>-42.0</td>
<td>683.5</td>
<td>2.3</td>
</tr>
<tr>
<td>AT3-52</td>
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<td>025.9</td>
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<td>056.6</td>
<td>-41.1</td>
<td>523.3</td>
<td>2.9</td>
</tr>
<tr>
<td>AT3-53</td>
<td>4/4</td>
<td>016.5</td>
<td>-55.0</td>
<td>038.2</td>
<td>-45.6</td>
<td>45.9</td>
<td>13.7</td>
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<td>AT3-63</td>
<td>6/6</td>
<td>032.9</td>
<td>-54.6</td>
<td>021.0</td>
<td>-35.7</td>
<td>162.0</td>
<td>5.3</td>
</tr>
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<td>AT3-64</td>
<td>5/5</td>
<td>180.9</td>
<td>41.1</td>
<td>208.2</td>
<td>37.9</td>
<td>126.1</td>
<td>8.5</td>
</tr>
<tr>
<td>AT3-65</td>
<td>6/6</td>
<td>318.5</td>
<td>-47.2</td>
<td>353.1</td>
<td>-70.8</td>
<td>87.4</td>
<td>7.2</td>
</tr>
<tr>
<td>AT3-66</td>
<td>6/6</td>
<td>350.3</td>
<td>-64.7</td>
<td>031.9</td>
<td>-53.6</td>
<td>466.4</td>
<td>3.1</td>
</tr>
<tr>
<td>Mean</td>
<td>10/11</td>
<td>017.3</td>
<td>-57.0</td>
<td>-</td>
<td>-</td>
<td>24.3</td>
<td>9.5</td>
</tr>
</tbody>
</table>

Table 5.5-4  Site mean characteristic directions and overall mean direction from the Paleocene Caldera Jorquera (n.b. tilt corrected using mesoscopic flattening fabric).
Figure 5.5-8  AMS data measured from the intracaldera strata within Caldera Jorquera. A) Plot of anisotropy parameters indicating that the majority of sites exhibit oblate (flattening) micro-fabrics although within each site samples are observed to display both oblate and prolate fabrics (colour coded by site for comparison (not numbered)). B) Stereonet plots of individual specimen and site mean principle susceptibility axes ($k_1$ (kmax)-squares (Purple), $k_2$ (kint)-triangles (Blue), $k_3$ (kmin)-circles (green)). Mesoscopic flattening fabrics measured in the field are displayed as planes (with pole indicated). C) Orientation of magnetic foliation. D) Orientation of magnetic lineation.
principal susceptibility axes (Figure 5.5.8b) it is clear that for the majority (7/11) sites, this relationship appears to be true, although the magnetic foliation (plane containing the $k_1$ ($k_{\text{max}}$) and $k_2$ ($k_{\text{int}}$)-Figures 5.5-8a & b) is not always observed to be coincident with the measured plane of flattening, suggesting that some disparity exists between the orientation of the macroscopic and microscopic fabrics. For reference the orientation of both fabrics measured at/for each site, are listed in Table 5.5-3.

The magnetic lineation (the $k_1$ direction in the $k_1$-$k_2$ plane) is generally sub-parallel to the magnetic dip (azimuth of magnetic foliation), but can switch to be sub-parallel to magnetic strike (compare figures 5.5-8 c and d). Given the limited availability of samples and the complexity of behaviours often observed in AMS of individual flows there is little to be gained from a further in depth discussion. It is concluded that the field observed flattening and AMS fabrics are in general parallel but the field observations are probably more reliable given that they represent much larger volumes of material. Both measured (flattening) fabrics will be used to tilt-correct the palaeomagnetic data from Caldera Jorquera in order to determine which, if any of the fabrics represents the most reliable proxy to a palaeohorizontal surface.

5.5.5 Discussion of Palaeomagnetic Data

Of the twelve sites sampled from Caldera Jorquera, eleven produced meaningful and well-constrained mean ChRM directions (Table 5.5-4), site AT301 being too weakly magnetised to be measured. With the exception of AT3-65, the site mean directions are well grouped, with the aberrant direction perhaps recording a non-dipole field direction or resulting from the sampling of a displaced block (Figure 5.5-9a). Application of a tilt-correction based on the mesoscopic flattening fabric is observed to cause the individual site mean directions to disperse slightly (Figure
Figure 5.5-9  Stereonet projections of site mean ChRM directions and overall mean direction from Caldera Jorquera A) In situ. B) Tilt-corrected i) using the mesoscopic flattening fabric measured in the field and ii) using the magnetic foliation determined through AMS studies.
5.5-9b) and produces a negative fold test. This would suggest that either the volcanic deposits have been remagnetised post-tilting of the flattening fabric, or that the flattening fabrics measured in the field were never originally horizontal reference surfaces.

The same analysis was repeated using the magnetic foliation and this produces even greater dispersion of directions (Figure 5.5-9b-directions shown graphically only). As the application of a tilt correction based on both the flattening fabric and the magnetic foliation causes dispersion of the site mean directions, two possible scenarios are presented. Firstly the ChRM is an overprint postdating the fabric or secondly the measured fabrics do not represent good approximations to palaeo-horizontal. The case for a remagnetisation is weak as the data appear to show a reasonable spread of directions, normal and reverse polarities, and are recorded by magnetite and haematite where these mineral components occur. Instead it is suggested that the flattening fabric does not represent a true palaeo-horizontal but as was suggested at the start may well include an element of palaeo-topography further complicated by the internal fabric development of ignimbrites.

Given initial concerns about the potential to sample the same cooling units at differing locations sites AT3-48, 64 & 66 are clearly different. The other potential grouping was sites AT3-04, 51, 52, & 53 which although generally similar are not sufficiently close to be considered a single cooling unit.

Caldera Jorquera is preserved within a small graben formed by two oppositely verging, subvertical reverse faults (Figure 5.1-3). Movement on these faults has been interpreted to be predominantly vertical [Iriarte et al., 1999] and appears to have had little effect on the overall orientation of the ignimbritic and volcanic strata. The ignimbrite and lava flows of Caldera Jorquera are therefore interpreted to
carry a primary remanence, with an overall mean direction of $D=017.3^\circ$, $I=-57.0^\circ$
with an associated $a_{95}$ error of $9.5^\circ$ (Table 5.5-4).
5.6 El Gato Pluton

5.6.1 Setting and sampling

The mid. Eocene (c.42 Ma-Iriarte et al., 1999) pluton El Gato has an irregular form with a 3:1 or 4:1 length to width ratio with a pronounced NE-SW elongation (Figures 5.1-3 & 5.6-1). North and west of the main sampling area, the pluton intrudes the Jurassic Lautaro Fm. limestones, where its roof is sub-parallel to bedding (Figure 5.6-2) and the late Cretaceous Quebrada Seca Formation in the east. The overall form and nature of the pluton suggests it was likely emplaced as a large sheet-like intrusion.

Fifteen sites were collected, from along Quebrada Carrizalillo and the northeast trending Quebrada San Miguelillo, (Figure 5.6-1, Table 5.6-1). Two sites (AT3-44 & 47) were sampled from a marginal monzodiorite facies and a third (AT3-45) from this facies' boundary with the main granodioritic facies (dotted line on Figure 5.6-1-Iriarte et al., 1999). The contact was not easily recognisable in the field, although the monzodiorite facies did appear to be slightly finer grained. Two further sites were sampled from a small granodioritic body intruding the western margin of Caldera Jorquera, ~10km south of the main sampling area and 2-3km from the eastern margin of the pluton. This intrusion is associated with the same phase of Eocene magmatism as the main El Gato pluton, and is one of many such small-scale intrusions observed in the La Guardia area [Iriarte et al., 1999]. Although finer grained in hand specimen, this intrusion appears to be of a similar granodioritic facies to the main El Gato pluton.

5.6.2 Magnetic Mineralogy

Rock magnetic experiments indicate that the magnetic mineralogy is remarkably consistent throughout the pluton and the smaller satellite intrusion into Caldera
Figure 5.6-1 Geological map of Pluton El Gato redrawn from the La Guardia mapsheet [Iriarte et al., 1999], with palaeomagnetic sampling localities indicated. 1- La Ternera Formation, 2-Lautaro Formation, 3-Q. Vicuinita strata, 4-Lagunillas Formation, 5-Q. Monardes Formation, 6-Q. Seca lower member, 7-Sierra Los Leones strata, 8-Q. Seca upper member, 9-Pluton El Gato/related intrusions, 10-Alluvium.

Figure 5.6-2 Photograph illustrating the near conformable contact between the sill-like Pluton El Gato and the marine sediments of the Jurassic Lautaro Formation, which comprises the roof of the pluton to the west. View to the NW.

Table 5.6-1 Site location, lithology and summary of sample demagnetisation behaviour from the Eocene Pluton El Gato.
Jorquera. The rock magnetic behaviour of the Pluton El Gato samples is summarised using the results from a single site, AT3-44 (Figure 5.6-3).

Powdered specimens display almost reversible thermomagnetic heating curves, with the Curie point observed to be between 560-600°C, which although slightly higher than would be expected, is consistent with (titano-) magnetite being present as the dominant magnetic mineral (Figure 5.6-3). A small inflexion is evident at around 350-400°C during heating, but is absent from all of the cooling curves, indicating that the minor were thermochemical alteration during heating and a slight lowering of the susceptibility being induced by the heating process.

To the north of the main sampling area within Pluton El Gato, widespread epigenetic mineralisation (i.e. formed after the host-rock) is observed to be associated with shallow epithermal (50-200°C) to mesothermal (200-300°C) Au-Ag mineralisation. This mineralisation manifests as elliptical brecciated deposits or 'breccia chimneys' [Diaz, 2000] that are observed to be particularly concentrated at the eastern border of the pluton and are exploited in a series of small mine workings. A classic consequence of such low temperature alteration (T<300°C) is the oxidation of magnetite to yield maghemite, chemically equivalent to haematite, but retaining the spinel structure of magnetite. Due to this similarity in structure, maghemite displays a very similar coercivity spectrum to that of magnetite, but is metastable and hence changes crystal system to that of haematite (hexagonal), between 300-500°C. The magnetic coercivity of haematite is known to be much greater than that of maghemite (i.e. haematite only reaches saturation magnetisation in much higher magnetic fields than is needed for maghemite), and therefore the transition of maghemite to haematite during heating, would explain the lower susceptibility cooling curve observed for all of the Pluton El Gato samples (Figure 5.6-3).
Figure 5.6-3  Magnetic mineralogy experiments representative of all samples analysed from Pluton El Gato and satellite intrusion.
IRM acquisition curves indicate that all of the samples saturate in fields of ~300 mT, with backfield coercivities ranging from 15-30 mT (Figure 5.6-3). This is consistent with magnetite being the dominant or sole magnetic fraction present. Thermal demagnetisation of a 3-component composite IRM [Lowrie, 1988] clearly indicates that the lowest coercivity (50 mT) component dominates, in addition to a smaller contribution from the medium coercivity (300 mT) component (Figure 5.6-3). Both of these components are evenly destroyed up to 570°C, although much of the IRM is removed by treatment to 400°C, but without a pronounced drop in intensity of any component indicative of the presence of another magnetic mineral.

In conclusion, rock magnetic experiments indicate that although titanomagnetite (probably of a range of Ti-compositions) is likely to be the dominant magnetic mineral, the pluton has undergone some low temperature alteration that has resulted in the oxidation of magnetite to produce maghemite.

5.6.3 Demagnetisation Behaviour

A-type demagnetisation behaviour

The majority of samples from all of the sites collected exhibit apparently very simple single to slightly more complex dual-component behaviour, which is observed to be relatively invariable across the pluton, with similar behaviour also noted from the small body intruding Caldera Jorquera. In general, all of the samples demagnetised reasonably cleanly (without excessive noise) using both AF and thermal demagnetisation techniques, allowing the determination of generally stable and fully resolvable magnetisation vectors interpreted using PCA (explained in Chapter Three).

By far the most common demagnetisation behaviour, whether in response to AF or thermal methods, was removal of a well-defined and origin-bound component of
ChRM ± a low (to intermediate) temperature/coercivity component (Figures 5.6-4a-d). Although typically of normal polarity, sites dominated by reverse polarity remanences are also observed to display identical behaviour, whereby the ChRM observed to be evenly demagnetised by alternating fields of ~10-100mT (Figures 5.6-4a & b), or is unblocked between temperatures of 400-580°C (although predominantly between 560-580°C-Figures 5.6-4c & d). This is denoted as the 'A-type' demagnetisation behaviour for Pluton El Gato.

The low-intermediate temperature components removed between 100-400°C, are interpreted to reflect the removal of a component of magnetisation carried by maghemite (e.g. sample AT3-42-02B, Figure 5.6-4c). This sometimes coincides with the present day field direction, particularly within samples carrying a normal polarity remanence (Figure 5.6-4c), but in the case of samples carrying a reverse polarity ChRM direction, the thermal unblocking spectrum of the intermediate component clearly overlaps with ChRM, producing an anomalous direction that cannot be rationalised as a record of the ancient Geomagnetic field (Figure 5.6-4d).

**B-type demagnetisation behaviour**

Many samples treated using AF demagnetisation display apparently univectoral directions, that are demagnetised past the origin (Figures 5.6-4e-g). There is little evidence of these vectors curving back towards the origin as a reverse polarity direction, as might be expected if two magnetisations are present with overlapping coercivity spectra. Stereographic representation of the demagnetisation data however, indicates the presence of an unresolved component of magnetisation, which is defined by great circle paths from the initial component, towards an opposite polarity component of NRM. This demagnetisation behaviour is termed the 'B-type' behaviour for Pluton El Gato, and is observed in samples displaying
Figure 5.6-4 Representative Zijderveld plots for typical Thermal and AF demagnetisation behaviour of samples collected from Pluton El Gato and a satellite intrusion, in the La Guardia Area. A-D) A-type behaviour characterised by a univectorial and origin bound ChRM direction, carried predominantly by magnetite. E-G) B-type behaviour characterised by dual polarity remanence that demagnetises along a great circle path with the ChRM remanence carried by magnetite and high temperature, reverse polarity remanence carried by haematite. H) Anomalous B-type behaviour dominated by a maghemite, displays an unexpected linear component of magnetisation is identified using PCA.
both normal & reverse polarity ChRM (magnetite carried) directions (Figures 5.6-4e-g).

The nature of this 'unresolved' component of magnetisation is unclear. The fact that it is not resolved through AF demagnetisation suggests that a high coercivity carrier mineral such as haematite may carry it, but no evidence of this was established during rock magnetic studies of the samples and Thermal demagnetisation does not indicate that the ChRM extends above temperatures of 580-600°C (Table 5.6-1). Thermal demagnetisation does attest to the removal of what is interpreted as a maghemite component (between 100-400°C) that is often observed to be slightly different to that carried by magnetite (Figure 5.6-4c).

Magnetite and maghemite display near identical coercivity spectrums [e.g. Hunt et al., 1995], should the two minerals carry slightly differing directions of magnetisation therefore, these components will be destroyed simultaneously by AF treatment, resulting in the observation of a composite direction that may not be directed towards the origin (Figures 5.6-4e-g). Thermal demagnetisation indicates that magnetite dominates the overall remanence, suggesting that any composite direction determined through AF demagnetisation, will be biased substantially towards the magnetite component. AF demagnetisation produces identical ChRM directions as observed during Thermal demagnetisation suggesting this is indeed the case.

Although constrained great circle paths were used to define potentially unresolved (maghemite) directions (Chapter Two), the ChRM is interpreted to represent that direction carried by magnetite, as defined by the linear segment on the Zijderveld plots shown in Figure 5.6-4. The only exception to this concerns samples from Site AT3-35, where demagnetisation produces a (normal-reverse) GC path, but which displays anomalously orientated linear 'ChRM' directions (Figure 5.6-4h).
Maghemite appears to be the dominant mineral in these samples, contributing ~75% of the initial NRM intensity. AF demagnetisation indicates the predominant carrier of magnetisation is of extremely low coercivity, suggesting that larger (MD) grains of maghemite developed at this locality. The increased dominance (and larger grain size) of maghemite suggests that this locality may have undergone more significant low temperature alteration than observed to affect the remainder of the pluton. It is noted that this site is situated on the boundary between what is mapped as the granodioritic groundmass of the main body of the pluton and the monzodioritic groundmass at the eastern fringe of the pluton [Irriarte et al., 1999].

5.6.4 Measurements of the anisotropy of magnetic susceptibility (AMS)

The anisotropy of magnetic susceptibility (AMS) was measured for the majority of samples from Pluton El Gato. Overall the data indicates that the majority of samples exhibit quite pronounced anisotropy of their magnetic fabric, with generally between 5-15% AMS observed (Table 5.6-2), with the most extreme AMS (15-20%), observed in samples from sites AT335 (which displays the greatest concentrations of maghemite) and AT338. Within the majority of sites, the shape parameter T is observed to be between 0 and 1, indicating an oblate (flattened) fabric dominates within Pluton El Gato, with only sites AT333 & AT344 exhibiting overall prolate (cigar-shaped) fabrics (Figure 5.6-5a).

The individual sample principle susceptibility axes are generally well clustered within each site and therefore the overall mean susceptibility axes are well defined, with minimal error (Figure 5.6-5b). The site average k₁ (maximum), and k₂ (intermediate) principle susceptibility axes are generally well constrained (with exception of sites AT3-40 & AT3-43, where the maximum and intermediate axes form a great-circle girdle), and indicate that the main body of Pluton El Gato is characterised by a steeply dipping oblate micro-fabric, that faces east-southeast.
### Table 5.6-2 Site mean AMS parameters calculated for the Eocene Pluton El Gato.

<table>
<thead>
<tr>
<th>Site</th>
<th>n/N</th>
<th>Dec. (°)</th>
<th>Inc. (°)</th>
<th>Dec. (°)</th>
<th>Inc. (°)</th>
<th>k</th>
<th>α95 (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT3-33</td>
<td>7/7</td>
<td>6.7</td>
<td>-56.8</td>
<td></td>
<td></td>
<td>255.0</td>
<td>3.8</td>
</tr>
<tr>
<td>AT3-34</td>
<td>7/7</td>
<td>7.5</td>
<td>-64.7</td>
<td></td>
<td></td>
<td>428.9</td>
<td>2.9</td>
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<tr>
<td>AT3-35</td>
<td>4/7</td>
<td>180.6</td>
<td>56.4</td>
<td></td>
<td></td>
<td>68.8</td>
<td>21.5</td>
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<tr>
<td>AT3-36</td>
<td>7/7</td>
<td>199.8</td>
<td>59.3</td>
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<td>28.3</td>
<td>11.5</td>
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<td></td>
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<td>6.6</td>
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<td>-61.4</td>
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<td>AT3-42</td>
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<td>9.0</td>
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<td>9.5</td>
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<tr>
<td>Mean</td>
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<td></td>
<td></td>
<td>52.9</td>
<td>5.3</td>
</tr>
<tr>
<td></td>
<td>17/17</td>
<td>4.4</td>
<td>-57.2</td>
<td></td>
<td></td>
<td>55.5</td>
<td>4.8</td>
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### Table 5.6-3 Site mean characteristic directions and overall mean direction from the Eocene Pluton El Gato and satellite intrusion.

<table>
<thead>
<tr>
<th>Maximum-k, Intermediate-k, Minimum-k</th>
<th>Corrected Anisotropy</th>
<th>Shape Parameter</th>
</tr>
</thead>
<tbody>
<tr>
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<td>Inc. (°)</td>
<td>Dec. (°)</td>
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<tr>
<td>AT333 206.1 10.1 63.4 77.4 297.4 7.5 1.062 -0.651</td>
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<td></td>
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<td>AT334 184.4 32.2 6.3 57.8 275 0.8 1.038 0.472</td>
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<td></td>
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<td>AT335 184.3 15.2 25.1 73.8 275.8 5.5 1.165 0.456</td>
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<td></td>
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<td>AT336 14.2 4.3 258.2 80.2 104.9 8.8 1.127 0.482</td>
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<td>AT337 185.4 21.3 56.1 58.3 284.6 22.2 1.089 0.122</td>
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<td></td>
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<tr>
<td>AT338 14.1 7.4 141.8 78.1 282.8 9.3 1.165 0.279</td>
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<tr>
<td>AT339 206 4.1 86.3 81.8 296.5 7.1 1.109 0.617</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AT340 183.7 41.7 38.2 42.8 299 16.1 1.144 0.889</td>
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<tr>
<td>AT341 191.9 41.3 47.2 42.9 299 18.5 1.083 0.278</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AT342 187.3 45.2 48.1 36.9 300.8 21.6 1.101 0.191</td>
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<td></td>
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<tr>
<td>AT344 Too few samples to calculate mean</td>
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<td></td>
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<td>AT345 156.3 45.9 49.0 22.6 293.6 35.5 1.072 0.092</td>
<td></td>
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<td>AT346 184.3 15.2 25.1 73.8 275.8 5.5 1.165 0.456</td>
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<td>AT347 206 4.1 86.3 81.8 296.5 7.1 1.109 0.617</td>
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<td>AT348 183.7 41.7 38.2 42.8 299 16.1 1.144 0.889</td>
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<td>AT349 191.9 41.3 47.2 42.9 299 18.5 1.083 0.278</td>
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<tr>
<td>AT350 187.3 45.2 48.1 36.9 300.8 21.6 1.101 0.191</td>
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</tr>
</tbody>
</table>

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337
Figure 5.6-5 AMS data measured from the main body of Pluton El Gato. A) Plot of anisotropy parameters indicating that the majority of sites exhibit oblate (flattening) micro-fabrics (colour coded by site for comparison-yellow sites along Q, San Miguello (not numbered)). B) Stereonet plots of individual specimen and site mean principle susceptibility axes ($k_1$ (kmax)-squares (Purple), $k_2$ (kint)-triangles (Blue), $k_3$ (kmin)-circles (green)). Mesoscopic flattening fabrics measured in the field are displayed as planes (with pole indicated). C) Orientation of magnetic foliation. D) Orientation of magnetic lineation.
Figure 5.6-5b). This generally oblate fabric is observed to shallow substantially towards the western margin of the pluton.

The magnetic lineation (parallel to $k_1$) is shown to generally trend southwards across the pluton, generally parallel to the strike of the magnetic foliation, but is observed to steepen to the west, with the most westerly site (AT347) trending towards the east (Figure 5.6-5c). The origin of the magnetic lineation is likely to result from the direction of magma transport, and is likely to have been determined by the preferred orientation of the silicate framework of the groundmass as the pluton cooled [e.g. Hargraves et al., 1999].

The minimum susceptibility ($k_3$) axes, which represents the pole to the magnetic foliation, is consistently well constrained at site level. For almost all of the sites, this $k_3$ axis is shallowly inclined towards the west/northwest, although it becomes more steeply inclined towards the western margin of the pluton. The only exception to this is observed at site AT336, where the minimum ($k_3$) axes are inclined shallowly to the east (opposite direction). Consequently the magnetic foliation, which lies normal to the $k_3$ susceptibility axis (and parallel to the common plane of $k_1$ and $k_2$), dips consistently to the east across the pluton, shallowing from near vertical to the east, to a more moderate dip of 44° at the western margin of the pluton (Figure 5.6-5d).

The magnetic foliation observed throughout Pluton El Gato most likely represents a primary fabric relating to the emplacement of the pluton, with the steep, sub-vertical foliation observed at the eastern margin of the pluton, shallowing to the west indicating that space was likely created for the pluton through either roof uplift (top to the west) and/or floor subsidence [e.g. Grocott & Taylor, 2002], similar to the mechanism proposed for the emplacement of Pluton Las Campañas by Truelove et al., (2005-Figure 4.5-2). Diaz (2000), states that the pluton was
emplaced through a process of 'telescoping' (as described by several authors for many copper porphyry-type Eocene-Oligocene aged deposits in northern Chile—Comejo et al., 1999, Zentilli et al., 1995, Perello et al., 1996, Comejo & Mpodozis, 1996), associated with ongoing rapid uplift and erosion at the time (42-40 Ma).

**Degree of AMS**

The generally accepted rule of thumb states that magnetic fabric anisotropies of $\leq 5\%$ will not significantly deflect any magnetisation direction acquired in a magnetic field and therefore the true magnetic direction of an ancient magnetic palaeo-pole will be recorded. Magnetic directions measured from rock samples with anisotropies in excess of 5% are considered to be suspicious as the measured magnetisations might not represent the actual direction of the magnetic field under which they were acquired, with the possibility that the magnetising field was deflected along a preferred direction as a consequence of the inherent rock magnetic fabric. In the case of the samples measured from Pluton El Gato, the high degree of AMS indicates that an intense magnetic fabric is observed.

The magnetic fabric of a rock results from the alignment of elongate or platy magnetic (usually ferromagnetic) grains, hence why AMS fabrics have been widely used as a proxy to mesoscopic fabrics in the study of highly deformed rocks for example. The degree of anisotropy of the samples from Pluton El Gato were measured using the low-field technique as described in Chapter Two (Palaeomagnetic Methods), in which AMS is often dominated by low coercivity (large?) multi-domain (tinano-) magnetite fractions, as well as by paramagnetic mafic minerals such as biotite, rather than by the (pseudo-) single domain (titano-) magnetite fraction which is more often than not seen to be dominantly responsible for carrying magnetisation within the rock. The consequence of this is that rocks
may exhibit fairly extreme degrees of anisotropy, but this may not affect the inherent capability to accurately record a magnetic field.

From the rock magnetic experiments carried out on samples from Pluton El Gato, it is clear that the dominant magnetic mineral present in all of the samples is (titano-) magnetite. Although the majority of samples exhibit quite extreme AMS indicating they may not have accurately recorded the ancient magnetic field at the time magnetisation was acquired, it is important to note that in magnetite dominated samples, AMS is dominated by multi-domain grains while single-domain (and pseudo-single-domain grains) are responsible for recording the applied field at time of acquisition of magnetisation [Stephanson et al, 1986]. If this is the case then the measured magnetisation directions of the samples from Pluton El Gato could be considered to represent accurate records of the ancient magnetic field.

5.6.5 Discussion of Palaeomagnetic Data

Low coercivity or unblocking temperature components (generally removed by 5-10 mT and 100-150°C) were easily removed and in the main found to be scattered and hence are believed to be due to short-term VRM acquisition post collection. The demagnetisation level at which the characteristic remanent magnetisation (ChRM) was recovered over all of the sites was reasonably consistent using both thermal and AF demagnetisation techniques (Table 5.6-1), with the majority of sites either stable up to, or completely demagnetised by 580-600°C or 80-100 mT, consistent with low-ti or single domain magnetite being the primary magnetic mineral.

The apparently pluton-wide, simplistic nature of the demagnetisation behaviour meant that a stable site ChRM was determined for each of the 17 sampling
localities and these mean directions (Table 5.6-3, Figure 5.6-6a). Within the majority of the sites there is a clear normal-polarity bias, with only 3 of the 17 recording reverse polarity ChRMIs demonstrating stable vector end-points. The three reverse polarity directions pass a reversal test [McFadden and MacElhinney, 1990] with a classification of C, indicating that the all of the site mean directions are drawn from populations with means 180° apart. This suggests that the isolated remanence is ancient in origin. Of the 17 sites sampled, 16 produced meaningful ChRM directions with a95 errors of less than 12° (Table 5.6-3), with the remaining site (AT335) producing a reasonable (reverse) direction but with a much higher a95 error of 21.5°. As discussed in the previous section, the demagnetisation behaviour of site AT335 is anomalous when compared with that of the rest of the sites sampled from Pluton El Gato, with the ChRM determined from GC analysis only as no stable vector endpoints were observed. This anomalous behaviour is possibly associated with localised higher temperature mesothermal alteration of magnetite to maghemite.

Díaz (2000), documented that the entire pluton has undergone a degree of low temperature epithermal alteration associated with the Au-Ag mineralisation of the pluton, and this may be the cause of the maghemite evident in all the samples, determined from the magnetic mineralogy experiments discussed at the start of this section. If the whole pluton has indeed undergone some alteration, the site mean ChRM's determined through palaeomagnetic analysis may represent secondary magnetisations and not the primary magnetisation associated with the emplacement and cooling of the Pluton. However, Díaz (2000), also concluded that the mineralisation of the pluton occurred very soon after the emplacement of the pluton, so even if the site-mean ChRM's isolated are secondary in nature, the timing of this magnetisation will not be far removed from that of the original acquisition of TRM. Initially, the interpretation of the palaeomagnetic data from
Figure 5.6-6  Stereonet projections of A) site mean ChRM directions and overall mean direction from Pluton El Gato, B) high temperature directions constrained using great circle analysis and overall mean direction (ChRM direction indicated in yellow.)
Pluton El Gato will be based on the assumption that the isolated site mean ChRM's are at least a proxy to the primary TRM, with an age similar to that of the pluton itself.

The inclusion of the two site mean directions from the intrusion into Caldera Jorquera makes very little difference to the overall Pluton mean direction and therefore are included, even though they belong to a smaller isolated body, approximately 10km to the south of the main sampling area in Pluton El Gato (Table 5.6-3). A mean direction calculated using all of the high coercivity components constrained by great-circle analysis (36 in total), identified from all of the sites from Pluton El Gato (Figure 5.6-6b). This produced a very similar (but ant-parallel) direction to that calculated from the site mean directions. In the remaining sites, where origin bound ('A-type') magnetisation vectors dominate, two options present themselves. Either these A-type directions represent Primary TRM's, with the rocks not affected by epithermal mineralisation evident elsewhere in the pluton (this could be explained by the fact that the mineralisation is seen to manifest as small irregular elliptical zones of brecciation up to 70m in diameter [Diaz, 2000], or that they represent a complete remagnetisation of the rocks, resulting in a secondary CRM, entirely overprinting the primary TRM. In either case, this 'type-A' magnetisation represents the ChRM of these sites.

As both the antiparallel magnetite and haematite carried components of magnetisation are interpreted to effectively be equivalent (at least with regard to the amount of rotation they record), the more widely observed ChRM direction carried by magnetite will be used to calculate the overall Pluton mean direction and is considered to represent a Primary TRM. Overall therefore, Pluton El Gato records a mean direction of $D=004.4^\circ$, $I=-57.2^\circ$, with an $\alpha_{95}$ error of 4.8°.
5.7 Summary

Although an almost contiguous range of strata spanning 200-40 Ma has been sampled in the La Guardia area, palaeomagnetic data indicates that Secondary magnetisations are widespread. The overall magnetisation direction recorded by these units will therefore reflect the amount of crustal rotation that has occurred post-dating remagnetisation. In order to properly determine not only the magnitude, but also the timing of crustal rotation, it is important to constrain the age of remagnetisation as precisely as possible and at least to the nearest 10 Ma, which is the temporal resolution of the palaeomagnetic reference poles used in this study [Besse & Courtillot, 2002, corrected 2003].

The relative age of the mean ChRM direction isolated from each of the units sampled in the La Guardia area has been discussed in the preceding sections and field stability tests have been used to determine a Primary/Secondary origin for each remanence. Based on this categorisation the overall pattern of observed magnetisations in the La Guardia area can be divided into three 'packages'.

5.7.1 Primary Triassic-Jurassic magnetisations

Although there are some differences noted between the age of magnetisation isolated at the different areas sampled in the La Ternera Formation, field stability tests suggest that the oldest Mesozoic strata of the latest Triassic La Ternera and Jurassic Lagunillas Formations sampled along the Rio Jorquera Valley preserve pre-deformation magnetisations (Chapters 5.3 & 5.4). Positive tilt-tests and failed conglomerate tests achieved from the La Ternera Formation in the Banderitas area and from the Lagunillas Formation (using data from both this study and that of Riley et al., 1993), suggest that these Formations have not been subjected to widespread (post-tilt) remagnetisation.
The fact that palaeomagnetic data from two separate studies of the same strata produce such similar results, combined with the observation of mixed normal and reverse polarity directions, strongly suggests that a stable, Primary magnetisation-acquired at the time extrusion/deposition-has been identified for both Formations. The La Ternera remanence is therefore considered to be a 200 Ma remanence, with the Lagunillas remanence slightly younger at 170 Ma. Only in the Fundo Santa Rosa-Manflas area is a post-tilt magnetisation of the La Ternera Formation suspected, interpreted to have produced the 'A-type' remanence (haematite dominated) that characterises the majority of the lava flows sampled. The age of this remagnetisation event is therefore <200Ma.

5.7.2 Secondary Cretaceous magnetisations

Palaeomagnetic data from both this study and that of Riley et al., (1993), demonstrate that the generally Cretaceous aged strata of the Quebrada Monardes, Cerillos and Quebrada Seca Formations appear to retain magnetisations that post-date their observed deformation (Chapters 5.2 & 5.4). Although the application of a tilt correction on palaeomagnetic data from the latest Jurassic-earliest Cretaceous Quebrada Monardes sandstones causes only a small dispersion of site mean directions (Figure 5.4-9), the complex demagnetisation behaviour dominated by reverse polarity directions, contrasts with the single-dual component NRM observed during demagnetisation of sandstones from the Lagunillas Formation, which essentially conformably underlie the Quebrada Monardes Formation (Chapter 5.4). Considering the similarity in lithology, if the Quebrada Monardes did retain a similar Primary magnetic remanence as observed for the Lagunillas Formation, one would expect to see greater conformity in the demagnetisation behaviour of samples from the two Formations. The overall magnetisation of the Quebrada Monardes Formation is therefore considered to be
a Secondary magnetisation that post-dates the deposition and subsequent deformation of the sandstones (i.e. remagnetisation occurred post 140 Ma). although a high temperature component isolated from a small number of the Quebrada Monardes red beds may reflect a pre-tilt magnetisation similar to that observed from the Lagunillas samples (Section 5.4).

The mid. Cretaceous Cerrillos Formation in the Elisa de Bordos area [Riley et al., 1993] is characterised by normal polarity in-situ directions that fail a tilt-test (Figure 5.2-4-Chapter 5.2)). The quite pronounced clustering of the in-situ directions is indicative of a very rapidly acquired Secondary magnetisation that may not fully average out the effects of PSV. This may explain the quite extreme declination of 046.8°, which is ~15° greater than that observed for the oldest strata sampled in the La Guardia sampling area from the late Triassic La Ternera Formation. Although the effects of PSV may not be properly accounted for, the overall direction is considered to be a genuine record of a remagnetisation event that must have occurred post 95 Ma.

The latest Cretaceous Quebrada Seca Formation (reinterpreted from Riley et al., 1993, using Iriarte et al., 1999) also appears to record a similarly secondary magnetisation, with the dominantly reverse polarity in-situ directions failing a tilt-test (Figure 5.2.6). The in-situ mean direction of the Quebrada Seca is almost identical, within error, to that observed for the Quebrada Monardes Formation to the East and both Formations are dominated by site mean directions of reverse polarity, which is not observed in any other (overall) sampling unit in the La Guardia Area. This would suggest that both the Quebrada Monardes and Quebrada Seca Formations were remagnetised at a similar time, post 65 Ma.

In summary, all of the Cretaceous aged Formations sampled in the La Guardia area, appear to have been remagnetised either at or post 60 Ma. Apparently two
remagnetisation events are identified, corresponding to a normal polarity remagnetisation evident in the western part of the field area, affecting the Cerrillos Formation and Sierra la Dichosa Lavas and La Higuera Strata, with a reverse polarity remagnetisation recorded to the east by the Quebrada Monardes and Quebrada Seca Formations.

5.7.3 Primary Paleocene and younger magnetisations

The youngest material sampled in the La Guardia area is associated with the products of the active Paleocene-Eocene magmatic and volcanic arc and that dominates the post-Cretaceous stratigraphy. Palaeomagnetic data from three different sampling units suggest no evidence that anything other than a Primary remanence is recorded (Chapters 5.6 & 5.7).

The various intra-caldera deposits sampled within the boundary of Caldera Jorquera displayed reasonably tightly clustered and predominantly normal polarity in-situ site mean directions. Palaeohorizontal control was measured from the often very pronounced flattening fabrics, visible within the majority of the ignimbritic and tuffaceous cooling units that were sampled. AMS measurements of the microscopic fabric of each sample were also made, in order to use the magnetic foliation as a second proxy to a palaeohorizontal surface.

The application of both measured fabrics produced negative tilt-tests, which initially suggests that a secondary remagnetisation is recorded by the collapse deposits of Caldera Jorquera. The use of flattening fabrics for palaeohorizontal control is suspect however where the cooling unit is interpreted to have initially have been deposited on a slope, such as towards the edges of the Caldera. In such instances, the flattening fabric will relate to the attitude of the palaeo-surface and not to the palaeo-horizontal. As a result, the only indication of the gross
deformation to have affected the Caldera deposits as a whole, are the limited post-collapse lacustrine limestones [Iriarte et al., 1999], which are sub-horizontal, suggesting very little deformation. For this reason the Caldera Jorquera deposits are considered to carry a Primary (~60 Ma) remanence.

As discussed at the start of this chapter, Caldera Jorquera is a single small caldera, associated with the construction of a much larger volcanic structure to the West referred to Megacaldera Carrizalillo. During the collapse of Megacaldera Carrizalillo, the Cabeza de Vaca pluton was emplaced along the western annular bounding fault, intruding the infill of the Hornitos half-graben. Although the Cabeza de Vaca pluton is shown to record predominantly normal polarity directions, as observed for Caldera Jorquera [Taylor et al., 2007], three antiparallel reverse polarity directions are also observed, suggesting that the overall magnetisation is also likely to be a Primary (~60 Ma) remanence.

The youngest sampling unit is the Eocene El Gato pluton (c.42 Ma), which records a mixture of normal and reverse polarity directions that pass a reversal test, in almost equal numbers. This suggests that an ancient magnetisation has been isolated, with no reason to suggest that it is anything other than a Primary (c.42Ma) remanence. The overall declination is much lower than observed for any of the other sampling units, suggesting that the observed remagnetisation of the Cretaceous sampling units certainly occurred prior to 42 Ma.

5.7.4 Magnetisation and Structural History of the La Guardia Area

The overall pattern of magnetisation observed in the La Guardia area, appears to suggest that the late Jurassic to late Cretaceous strata was reset by a 60Ma remagnetisation event. In contrast the older strata retain what is interpreted to represent a Primary (pre-tilt) remanence, whilst the Paleocene to Eocene strata
also retain a Primary remanence, but one that was acquired after the observed
deformation in the La Guardia area.

The main phase of deformation recorded in the La Guardia area appears to be
associated with the initial period of Andean uplift during a phase of Eocene
sinistral transpressive deformation, c. 42-36 Ma [Mendoza et al., 1994; Tomlinson
et al., 1993]. This phase of deformation is generally referred to as the Incaic
orogenic event and one of the main features associated with this in the La Guardia
area is the La Ternera fault system.
Chapter Six

Magnetostratigraphic sampling of the Barremian-Aptian

Pabellón Member of the Chañarcillo Group

6.1 Introduction

The opportunity arose to sample the Pabellón formation as part of a team interested in the Barremian-Aptian transition, which is marked by significant climatic and environmental change characterised by a global positive carbon isotope excursion (Figure 6.1-1a). The predominantly calcareous lower Cretaceous marine sequence of the Atacama region of northern Chile offered the opportunity to study marine rocks of Barremian-Aptian age and a magnetostratigraphic investigation in support of a high-resolution oxygen and carbon isotope sampling across the Barremian-Aptian interval was undertaken during late January, 2003. While it was recognised from the outset that carbonates would not necessarily provide a good magnetic recording medium it was felt worthwhile in the overall context of the region to undertake the study.

6.1.1 Background to the isotope excursion

Recent studies of Tethyan successions have demonstrated that this positive early Aptian event was preceded by a brief, but large magnitude, negative carbon isotope excursion and this has been linked to large-scale release of methane from gas hydrate [Bralower et al., 1999]. An alternative mechanism that could account for this early Aptian negative excursion is increased volcanic activity and enhanced CO₂ emissions associated with a pulse in crustal production attributed to a mantle 'superplume' which resulted in the emplacement of the Ontong Java
Figure 6.1-1  A) Global carbon isotope curve for the Barremian-Aptian period (Bralower, 1999), with data from the Pabellon Formation shown on the left. Note the positive excursion in both curves at the Barremian-Aptian transition. B) Mesozoic marine basins in the Andean margin of South America. C) Geological map of the area to the south of Copiapo illustrating the stratigraphical position of the Pabellon Formation (from Price et al., 2006).
Plateau. This massive and nearly instantaneous outpouring of basalt would have led to an increase in hydrothermal activity and an increase in CO₂ outgassing.

6.1.2 Magnetostratigraphy

The Barremian-Aptian boundary (Figure 6.1-1a) coincides with very short M0 reverse chron, which represents the last period of reverse polarity before the onset of the Cretaceous Long Normal Polarity Superchron, which lasted some 35Ma with no stable periods of polarity reversal. The top of the Pabellón Formation has been given a minimum Aptian age based upon ammonite biostratigraphy [Perez et al., 1990], with the underlying Totoralillo Formation is assigned a Late Hauterivian to Barremian age [Marschik & Fontboté, 2001].

The intention was to compile a polarity record through the Pabellón Formation in an attempt to isolate the M0r reverse chron. Due to the late Barremian to Aptian biostratigraphic age suggested for the Pabellón Formation, should a primary magnetisation be isolated from the sampling area, any reverse polarity direction or directions was likely to be a record of the M0r reverse polarity chron and hence an independent check on the stable isotope correlation.

6.1.3 Stratigraphic setting

The Pabellón Formation represents the uppermost member of the Chañarcillo Group, which in total comprises some 1700-2000m of shallow marine carbonate sediments [Arévalo, 1994, 1995], with near continuous exposure in the Copiapó valley and adjacent areas (27-28°S). The Chañarcillo Group is interpreted to be part of an Early Cretaceous backarc basin, the Atacama Basin, which is believed to have been connected to the Argentine Neuquen Basin during the mid-late Cretaceous, with similar faunal content recorded in both basins (Figure 6.1-1b). As noted in chapter 5 the marginal basin facies interfinger with the terrestrial deposits.
of the volcanic arc rocks of the Bandurrias group to west, these sediments are therefore interpreted to represent the western shoreline of the backarc basin in the Copiapó area [Marschik & Fontboté, 2001].

Although a regionally unconformable relationship is suggested for the contact between the marine Pabellón Formation and the overlying purely continental, volcanic/volcaniclastic Cerrillos Formation by Marschik & Fontebote (2001), this is was not observed in the Quebrada El Molle Area, where the Pabellón Formation passes conformably upward into the Cerrillos Formation, with no obvious erosional unconformity observed.

6.2 Palaeomagnetic Sampling

6.2.1 Sampling localities

The type section of the Pabellón formation crops out in Quebrada el Molle (70°20'E, 27°51'S), situated ~60km due south of Copiapó (Figures 6.1-1c, 6.2-1). The section was logged during the palaeomagnetic and stable isotope sampling. Palaeomagnetic samples were collected using the normal equipment and core recovery procedure, as described previously. Over 180 core samples were collected from five sub-sections (PBA-PBE of Figure 6.2-1 and Table 6.2-1).

The majority of samples were collected from the lower 200m of the formation (section PBA–PBC) with the upper part sampled at a lower resolution, (PBD and PBE) due to the limited time available and the fact that the established biostratigraphy indicated the Barremian-Aptian transition was likely to be toward the base of the section. It is clear from the log in Figure 6.2-1a, that there are two large gaps in the sampling of the upper 300m of the formation. Approximately 90m of the stratigraphic log comprises two andesitic-basaltic, sill-like intrusions/extrusive lavas known as the 'Ocoitas' and it is below these intrusions
Figure 6.2-1  A) Stratigraphic log for the Pabellón Formation, compiled during stable isotope and palaeomagnetic sample collection. B) Satellite image of the Pabellón Formation (screenshot from Google Earth™) indicating the position of individual sampling profiles and extent of the field photograph shown in C) which indicates the position of the most prominent units shown in the log.
Table 6.2-1 Summary of division of palaeomagnetic data from the Pabellón Formation profile and generalised demagnetisation behaviour observed within each division.

<table>
<thead>
<tr>
<th>Site</th>
<th>Sampling Site</th>
<th>Lithology</th>
<th>Tilt Correction</th>
<th>°C</th>
<th>ChRM Isolation</th>
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<td>Long (°)</td>
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<td></td>
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<td>150</td>
<td>500-580</td>
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<tr>
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<td>289.66</td>
<td>Various</td>
<td>150</td>
<td>580</td>
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<td>580</td>
</tr>
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<td>400-580</td>
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<tr>
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<td>Limestone-various</td>
<td>150</td>
<td>500-580</td>
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<td>C</td>
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<td>289.66</td>
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<td>289.66</td>
<td>Limestone-various</td>
<td>150</td>
<td>500-580</td>
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</tbody>
</table>

Table 6.2-2 Palaeomagnetic data from the sub-divided Pabellón Formation profile. Samples collected between 337-355m (shaded grey) are sourced exclusively from the Ocoitas intrusive sill which postdates the deposition of the Pablélon Formation and therefore are not used to calculate the overall mean direction of the Pabellón Formation.
that the majority of samples were collected. Sampling section PBD incorporated the upper Ocoitas horizon (Figures 6.2-1a & b), along with the beds immediately above and below in order to compare the preserved magnetisation in both rock types.

The uppermost section, PBE, was sampled from material within the upper 200m of the formation (Figure 6.2-1a). This part of the Pabellón Formation becomes increasingly more marginal upwards, with an obvious increase in the amount terrigenous clastic material. This sampling section also coincided with the only in-situ ammonite find during the fieldwork, (Figure 6.2-1a), with only a handful of finds described previously for this Formation [e.g. Perez et al., 1990]. The ammonite was found in the Parahoplites gr. Nutfieldensis (late Aptian) zone and is tentatively identified as this species. The specimen was presented to Ernesto Perez of the SERNAGEOMIN for identification. This part of the formation was sampled so as to tie the palaeomagnetic data with the scant macro fauna found.

6.2.2 Magnetic Mineralogy

Several different lithologies are recognised throughout the Quebrada El Molle section, from dark coloured cherty limestones, to sandy cross-bedded units towards to the top of the formation (Figure 6.2-1a). Samples that were broadly representative of each major lithology were subjected to the normal suite of rock-magnetic experiments in order to determine the nature of magnetic minerals present and hence interpret the demagnetisation data more effectively.

The rock magnetic experiments indicate that the magnetic mineralogy of the limestones of the Pabellón Formation can be split into two groups. The first, and more common type of behaviour has samples dominated by a low to medium coercivity carrier of magnetisation (Figure 6.2-2). These samples are characterised
Figure 6.2.2  Typical examples of (titano-) magnetite dominated rock magnetic experiments on samples from the Barremian-Aptian Pabellon Formation.
by rapid acquisition of IRM up to 100 mT, after which the acquisition of magnetisation is only acquired gradually, although none of the samples are observed to fully saturate in the maximum field applied of 800 mT. Such behaviour suggests that titano-magnetite, perhaps of a larger grain size, or possibly but less likely maghemite (which displays a similar coercivity spectrum to that of magnetite), is likely to dominate the magnetic mineral assemblage. The slower acquisition of magnetisation within higher strength fields indicates the presence of another, high coercivity mineral, such as hematite or goethite for example.

The corresponding susceptibility (k) versus temperature (T) plots show that the majority of samples tested underwent substantial thermochemical changes on heating, such that the bulk susceptibility on cooling (open symbols) greatly exceeds that during heating. This indicates that a weakly magnetic (or non-magnetic) mineral is converted to a much higher susceptibility phase. Recycling the experiment clearly indicates this effect was due to the generation of magnetite, with susceptibility decreasing between 400-580°C. One exception to this behaviour was sample PBA66 (not illustrated), where another mineral in addition to magnetite was also observed to be produced during heating, evident when the sample is cooled to ~300°C and below. The Curie point of this second mineral is possibly consistent with the production of pyrrhotite, but the specimen was very weakly magnetic and the results of the experiment consequently very noisy.

Three-component demagnetisation experiments (Figure 6.2.2) are consistent with titano-magnetite or maghemite dominating the magnetisation with the highest coercivity component imparted along the Z-axis contributing only very little to the overall magnetisation. Thermal demagnetisation of the multi-component remanence shows that a very low temperature/low coercivity component is removed after treatment to 100-200°C, with subsequent heating steps steadily
destroying the IRM up to 580°C, often with a noticeable drop in intensity observed at ~400-420°C. These observations suggest that a mixture of both titanomagnetite and maghemite was probably present within the samples.

Figure 6.2-3 illustrates some examples of rock magnetic experiments that indicate the presence of varying amounts of a much higher coercivity mineral, that dominates the overall magnetic mineral assemblage, but in addition to magnetite (+/- maghemite). All of the samples displaying this behaviour initially acquire an IRM rapidly in fields up to 100mT, as observed for those samples illustrated in Figure 6.2-2. Although the presence of hematite would explain the continued acquisition of magnetisation in the largest magnetic fields (Figure 6.2-3), thermal demagnetisation of a three component IRM clearly shows the high coercivity component of magnetisation along the Z-axis is almost completely destroyed by heating to 100°C. This high coercivity but low temperature behaviour is indicative of Goethite, which possesses a Curie temperature of ~120°C, but saturates in magnetic fields in excess of 1T [e.g. Lowrie, 1990].

The only slight exception to this behaviour is demonstrated by sample PBC04 (Figure 6.2-3b), where the hard component of magnetisation along the Z-axis is observed to persist after the removal of goethite by treatment to 100°C. Whilst the hard component is not demagnetised directly with the soft and intermediate components, all three components are destroyed at ~580°C. This may indicate that titanomagnetite of a range of compositions (and probably grain sizes) is present within this sample, with a significant component of ti-poor magnetite interpreted to be responsible for carrying the more pronounced hard component.

Natural dehydration of goethite (or laboratory heating to 300°–400°C) produces haematite and is an important process in formation of red sediments [Butler, 1992]. Whilst this probably explains the presence of small amounts of hematite,
Figure 6.2-3 Typical examples of goethite dominated rock magnetic experiments on samples from the Barremian-Aptian Pabellon Formation.
the recognition that magnetite is produced through heating of powdered samples, may reflect the alteration of goethite (Figures 6.2-2 & 3), which can produce magnetite as an intermediate product during dehydration to haematite [e.g. Özdemir & Dunlop, 2000].

In summary, rock-magnetic experiments on representative samples from the Pabellon limestones indicate that whilst several minerals comprise the bulk magnetic mineralogy of the samples, the most prevalent minerals are titanomagnetite (as a range of grain-sizes), which are presumably detrital in origin and goethite, identified in ~30% of the samples subjected to rock magnetic analysis.

6.2.3 Analysis of Palaeomagnetic Data

The low intensity of the NRM magnetisation typically <1 mA/m (and often <<1 mA/m) meant that the specimens could not be reliably measured using the facilities available in the Palaeomagnetic Laboratory at the University of Plymouth. Samples were therefore measured using an automated 2G SQUID cryogenic magnetometer, housed in the University of Oxford Palaeomagnetic Laboratory.

Initially a series of pilot specimens, chosen as representative of the range of lithologies were subjected to progressive AF demagnetisation using the in-line 3-axis static AF demagnetiser, incorporated within the SQUID magnetometer. Specimens were treated using 17 demagnetisation steps up to a maximum field of 100 mT. For the vast majority of samples in the pilot study, the AF demagnetisation technique proved to be exceptionally efficient in the progressive destruction of remanent magnetisation (Figure 6.2-4a), and due to this effectiveness and the apparent simplicity of the remanent magnetisation isolated from the pilot samples, many of the remaining samples were demagnetised using nine AF steps up to 100 mT (Figure 6.2-4b).
Figure 6.2-4  Typical examples of the demagnetisation behaviour (Thermal & AF) of samples from the Barremian-Aptian Pabellón Formation.  

A) 'A-type' (univectorial, origin bound).  B) 'B-type' (poorly constrained GC demagnetisation trend, but linear segments interpreted as ChRM-equivalent to 'A-type' direction).
The most common demagnetisation behaviour (designated A-type) consists of a single origin-bound vector of remanent magnetisation, although often subsequent to the removal of a low coercivity component. This A-type remanence is evenly and almost completely eliminated in fields up to 100 mT. The direction of remanent magnetisation is very stable and maintained over a wide coercivity spectrum (Figure 6.2-4a & b) consistent with titano-magnetite (or maghemite) as the main carrier of remanence, as suggested from rock magnetic experiments.

This A-type remanence was also recognised through thermal demagnetisation of samples, such that they demagnetised evenly up to 500-580°C at which point the remanence was completely destroyed (Figures 6.2-4c & d). During heating to >580°C, large increases in measured intensity were accompanied by randomised directions, that many of the samples underwent thermochemical alteration during the final heating steps, resulting in the production of another magnetic mineral. The pronounced increase in susceptibility after treatment to temperatures >580°C probably reflects the secondary production of magnetite, through the dehydration of goethite [Özdemir & Dunlop, 2000].

Although the majority of samples show the A type characteristics noted above, some ~40% of the samples produced apparently linear magnetisation vectors that either overshot, or otherwise were not directed toward the origin, with this behaviour noted during both AF and Thermal demagnetisation (Figure 6.2-5). Although the measured vector of magnetisation initially clusters about a point, as the remanence is destroyed by subsequent demagnetisation steps, so the vector of magnetisation is observed to track along a (often poorly defined) great-circle path, suggesting the presence of an unresolved component of magnetisation (Figure 6.2-5).
Figure 6.2-5  Examples of goethite and magnetite dominated remanences from the Barremian-Aptian Pabellon Formation. A) Initial AF demagnetisation destroys magnetite carried component of remanence-ChRM direction represented by (normal polarity) linear segment-but high coercivity component remains. Subsequent Thermal demagnetisation destroys remanence by treatment to 100°C, suggesting goethite. B) Initial Thermal demagnetisation removes Goethite component to leave univectorial ('A-type') remanence.

Figure 6.2-6  Solitary reverse polarity remanence determined from samples from the Pabellon Formation that is not carried by Goethite.
Whilst AF demagnetisation suggests that a high coercivity component is unresolved (Figure 6.2-5a), samples treated thermally appear to be fully demagnetised on treatment to 580°C, with the majority of remanent magnetisation unblocked at 100°C. This suggests that the ChRM direction is effectively an A-type remanence, (interpreted using PCA to fit linear segments) carried by magnetite and later overprinted by a substantial goethite contribution.

AF demagnetisation of sample PBB44 shows that a high coercivity component is left intact after AF treatment to 100 mT (Figure 6.2-6a), although a normal polarity component of magnetisation is clearly removed, accompanied by an increase in measured intensity. Once this component is removed, a hard, reverse polarity component remains, with the measured vector of magnetisation clustering on both the Zijderveld and stereonet plots. Subsequent thermal demagnetisation of this high coercivity component shows that it is destroyed after treatment to 100°C, with only random directions observed above this (Figure 6.2-6a). Sample PBB51 was subjected to thermal demagnetisation only and clearly displays the unblocking of a substantial proportion of the initial NRM at 100°C (Figure 6.2-6b). After this component is removed an effectively A-type remanence is demagnetised, carried by (titano-) magnetite.

Assuming goethite is formed as a secondary magnetic mineral, any remanence carried by goethite must by definition represent a secondary CRM. It is therefore proposed that the linear, normal polarity components of magnetisation, identified through demagnetisation of the B-type remanences, observed to overshoot or otherwise diverge away from the origin on Zijderveld plots of the data, actually represent A-type remanences carried by titano-magnetite/maghemite, over which a CRM carried by goethite is overprinted at a later date. The observation of goethite dominated remanences corresponds to the actual appearance of the
samples, being extremely reddened in comparison to those dominated by
titanomagnetite dominated magnetisations. The remanence carried by goethite is
therefore considered to be a secondary CRM and therefore any reverse polarity
directions carried by goethite cannot be considered valid for magnetostratigraphic
purposes.

6.2.4 Magnetostratigraphic Profile

The combined log for the Pabellón Formation includes bulk susceptibility, initial
NRM intensity, NRM declination and inclination and ChRM declination and
inclination (Figure 6.2-7) all plotted in the diagram. All directions plotted are in
geographic coordinates.

The bulk susceptibility varies between 2.38-227.00 x 10^-6 SI units, with a mean
susceptibility of 67.35 x 10^-6 SI units with the majority of high bulk susceptibilities
concentrated around 50m from the base of the section and again between 400-
450m, towards the top of the formation. Initial NRM intensity varies between
0.103–9.794 mA/m, with a mean of 1.20 mA/m, and there is, as one might expect,
a degree of correlation between bulk susceptibility and NRM intensity. The highest
values of both bulk susceptibility and initial NRM intensity, appear to be associated
with those beds of the Pabellón Formation that are more sandy in composition,
perhaps reflecting an increase in the amount of detrital magnetic particles,
although the highest single NRM recorded-9.794 mA/m (sample PBB51-Figure
6.3-6b)-appears to be associated with the presence of significant amounts of
goethite.

As expected for a unit deposited entirely within the Cretaceous Long Normal
Period, only a single component of normal polarity is present (with the exception of
a single sample-PBB47-Figure 6.3-7). However the directions presented to date
Figure 6.2-7  Magnetostratigraphic log for the Pabellon Formation sampled through Quebrada El Molle.
are all in situ, as tilt correction (including of sample PBB47) produces an impossibly low inclination. This coupled with the extremely uniform nature of the ChRM directions of the Pabellón samples and the fact that the only clearly defined reversed sample is interpreted as having been magnetised after the sequence was tilted/deformed, indicates that the isolated ChRM magnetisation is an overprint and hence is of no value in making magnetostratigraphic comparison.

6.2.5 Discussion of palaeomagnetic data from the Pabellón Formation

For directional analysis purposes, the palaeomagnetic data were split into groups of similar numbers of samples (in comparison to the standard sampling of six or more core samples per site) based on stratigraphic position (Table 6.2-2 and Figure 6.2-8). This produces 19 well-constrained site mean directions that are clearly tightly clustered in the upper (normal polarity) hemisphere. The overall mean is tightly defined in both in situ and tilt corrected coordinates, but does however fail a tilt test, with or without exclusion of the most extreme directions. Further evidence for this being a secondary magnetisation is the lack of detection of the M0r anomaly (alternatively one would need to disregard the biostrat evidence and suggest the section is as a whole younger). Furthermore, the tilt corrected declination, if primary, would inevitably mean that the crustal rotation in this sector would be markedly different from results elsewhere in the Copiapó area. There is therefore, compelling evidence to believe this to be a secondary magnetisation, meaning that the Pabellón Formation records an overall direction of D=029.4°, I=-44.0°, with an α95 error of 3.1° (k=121.4).

The earliest potential remagnetisation event(s) likely to have reset the magnetisation is intrusion of plutons within the eastern part of the Coastal Batholith. Ages as young as 110 Ma [Arévalo et al., 2006] are known from the immediate area and in general plutonic activity continued until c. 90 Ma in the
Figure 6.2-8  'Site' mean directions and overall mean direction calculated for the Pabellon Formation.
extreme east of the batholith (see chapter 5). A younger possible event could be linked to the intrusion of the Ocoitas as sills (although doubt remains as whether or not they are actually lava flows), which have been K-Ar dated at 77 Ma [Arévalo, 1994; 1995]. Clearly the palaeomagnetic data (Table 6.2-2) show no noticeable difference between the sills and the sediments hence it is likely they were magnetised at the same time. Other speculative remagnetisation events would include the expulsion of fluids during the deformation of the Chañarcillo fold and thrust belt or more distant events associated with the younger but more easterly Megacaldera Carrizalillo.

Currently remagnetisation by the intrusion of the Ocoitas sills is favoured as the most likely trigger for this event. Firstly the Ocoitas outcrops over a substantial distance, >40km strike length [Arévalo, 1995, 1995] and therefore would represent a significant heatflow into the upper crust. It would however mean that the sills would have been intruded post tilting, implying the Chañarcillo fold and thrust belt pre-dates this event which is the accepted stratigraphic relationship [e.g. Grocott & Taylor, 2002].

As has been discussed previously, reference directions calculated using the APW paths of Besse & Courtillot (2002, 2003), show very little change between the field direction of the mid-Cretaceous and the present day and therefore it is impossible to use palaeomagnetic evidence alone to distinguish the timing of remagnetisation. What is clear however from the isolated directions is that the remagnetisation must have been acquired prior to the regionally consistent, and very large ~35°, net clockwise rotation (discussed in Chapter Eight). This implies that whatever mechanism is responsible for the overall remagnetisation of the Pabellón Formation, the recorded clockwise rotation must be younger than the 77 Ma published age for the Ocoitas Intrusions.
7.1 Tres Cruces Study Area

7.1-1 Introduction

All sampled units in the Tres Cruces area were interpreted as retaining primary magnetisations (see Chapter Four). The magnitudes of crustal rotation have therefore been calculated with reference to the expected field direction corresponding to the age of each sampling unit (Table 7.1-1, Figure 7.1-1a).

Palaeomagnetic sampling in the Tres Cruces area was concentrated along ~30km transect in Quebrada de Los Choros, spanning some 60My of stratigraphy. In addition plutons Tilgo and Las Campañás, which lie close to the transect, are discussed in relation to the gross pattern of rotation identified along Quebrada de Los Choros. The discussion also covers the recalculated palaeomagnetic data from the Condoriaco area some ~50km to the south of Quebrada de Los Choros.

7.1-2 Quebrada de Los Choros

All of the four sampling units collected (plutons Los Colorados and Corredores, the Arqueros Formation and the Quebrada de Los Choros dyke swarm) with ages between 130-70Ma, yield remarkably similar magnitudes of crustal rotation (Table 7.1-1, Figure 7.1-1a). The area is therefore interpreted to record a homogenous CW rotation of ~10° (±2°(standard deviation)), which must be post 70Ma in age. Even at the individual site level, the observed rotation is essentially invariant with respect to lithology, site location along the transect, or the polarity of the isolated remanence (Figure 7.1-1b). The fact that only a very small number of sites were
<table>
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<th>Sampling Unit</th>
<th>Latitude (°S)</th>
<th>Longitude (°E)</th>
<th>Mean Direction Dec. (°)</th>
<th>Inc. (°)</th>
<th>Reference Pole</th>
<th>Rotation ± error (°)</th>
<th>Flattening ± error (°)</th>
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<td>11.2</td>
<td>-50.5</td>
<td>60 Ma</td>
<td>18.6 ± 12.2</td>
<td>-1.9 ± 7.6</td>
</tr>
</tbody>
</table>

Table 7.1-1 Overall mean directions, age of magnetisation and magnitude of crustal rotation (+ve clockwise) recorded by sampling units in the Tres Cruces and Condoriaco field areas. Condoriaco data recalculated from Palmer et al., (1980a).
Figure 7.1-1  A- Magnitude of crustal rotation recorded by sampling units from the Tres Cruces (numbered circles (located to the north)) and Condoriaco (numbered squares (located to the south)) areas [recalculated after Palmer et al., 1980a]. Blue arrows- primary magnetisations, orange arrows-secondary magnetisations. Circles-1- Arqueros Fm, 2-QLC dyke swarm, 3-Pluton Tilgo, 4-Pluton Los Colorados, 5- Pluton Corredores, 6-Pluton Las Campañas. Squares-1-Arqueros/Q. Marquesa Fm., 2-QLT strata/Vinita Fm., 3-Los Elquinos Fm. B- Site by site pattern of crustal rotations recorded along Quebrada de Los Choros. Magnitude of crustal rotation [calculated after Beck, 1980] indicated by angle from north with arc segment representing the calculated error [calculated after Demerest, 1983].
rejected due to poorly constrained or clearly anomalous directions, also suggests
that no errors have been introduced during the process of sample collection or
data acquisition/interpretation that could produce fictitious rotations. The observed
uniformity also demonstrates the accuracy/internal consistency of the reference
poles used to calculate the rotations [Besse & Courtillot, 2002, 2003].

One potential concern surrounding the homogeneity of the observed pattern of
crustal rotation is that the entire sampling area may have undergone complete
remagnetisation. This is considered as unlikely because of a) the variety of
lithologies, b) lack of evidence of any significant alteration or metamorphism that
might be expected to accompany a pronounced (hydro-) thermal or burial event. In
addition to the lack of alteration, a reasonable amount of dispersion is observed
between site mean directions (Figure 7.1-1b), which is consistent with the record
of PSV.

Some doubt remains concerning the precise age of the magnetisation isolated
from pluton Los Colorados as several reverse polarity directions were recovered
contradicting the radiometric age, which implies intrusion during the Cretaceous
Long Normal period. The remanence isolated from the Arqueros Formation host
rocks is interpreted to pre-date the deformation of the area and therefore is
considered to record a primary remanence. It is considered highly unlikely that
pluton Los Colorados could have undergone a significant remagnetisation event,
without affecting the adjacent host rocks. In addition, the presence of mixed
polarity directions from the latest Cretaceous-earliest Paleocene pluton
Corredores also suggests that an ancient (primary) TRM was acquired by the
70Ma pluton, suggesting that a remagnetisation after this time is unlikely. The
remanence isolated from Pluton Los Colorados is therefore also considered to be
Primary, with the reverse directions considered to represent an abberation of the localised field at the time of intrusion.

7.1-3 Pluton Tilgo

Whilst homogenous CW crustal rotations of ~10° are identified from the Quebrada de Los Choros transect, pluton Tilgo, situated ~50km to the west, records a substantially greater CW rotation of ~25° (Table 7.1-1, Figure 7.1-1a). The increased rotation is possibly a function of the small number of sites used to calculate the overall mean direction, which may not fully average the effects of PSV. Conversely, there is no reason to suggest that the overall mean direction is not a valid record of the ancient magnetic field.

7.1-4 Pluton Las Campañas

This pluton located ~50km N of Quebrada de Los Choros, displays an identical range of magnetic behaviour to the equivalent pluton Corredores. Both plutons belong to the latest Cretaceous-earliest Paleocene magmatic arc and should therefore record essentially the same palaeo-field direction (70Ma). Pluton Las Campañas records ~30° of CW rotation (Table 7.1-1, Figure 7.1-1a), substantially greater than the ~10° recorded by the entire Quebrada de Los Choros locality (including Pluton Corredores) and approximately equal to the magnitude of crustal rotation observed in the Vallenar area (c.28°30'S-e.g. Gipson, unpublished data) further to the north. This suggests that plutons Las Campañas and Corredores are situated in different structural blocks and that a ~20° decrease in the magnitude of crustal rotation is accommodated by a major structure located between 29°00' and 29°30'S (~50Km-Figure 7.1-1a).
Results from the stratigraphically equivalent Arqueros and Quebrada Marquesa Formations (130Ma) were combined and clearly record a primary magnetisation. Likewise the combined data from the La Totora Strata and Vinita Formation (100Ma) are also inferred to record a pre-tilt magnetisation. Although the older strata appear to record substantially more rotation, possibly suggesting that an early Cretaceous component of rotation is identified in the Condoriaco area, the disparity is within the level of error associated with the calculated rotations (Table 7.1-1) and is therefore considered to be insignificant. The Elquinos Formation fails a fold test (attributed to a 60Ma remagnetisation) but records an identical amount of CW rotation as the Arqueros/Quebrada Marquesa Formations (Table 7.1-1, Figure 7.1-1a). This infers that as observed along Quebrada de Los Choros, the small magnitude crustal rotations isolated from the Condoriaco area are locally consistent and post-date 70-60Ma.

The average rotation of ~16° (±4°) in the Condoriaco area slightly exceeds that observed along Quebrada de Los Choros, although the amount of difference is considered to be small and within error. As observed along the Quebrada de Los Choros transect, the uniformity of the rotation pattern recorded by the (time-equivalent) units sampled from the Condoriaco area again indicates that rotation post dates 60Ma (the inferred timing of the remagnetisation of the Elquinos Formation). Sample sites span a distance of ~40km across strike, similar in dimension to the Quebrada de Los Choros area. Together the site to site uniformity of rotation at these scales strongly suggests that localised (small-scale) structures, exert little or no control over the observed pattern of crustal rotation at these latitudes. The two areas are therefore interpreted to record a small magnitude, but regionally consistent crustal rotation.
7.1-6 Summary

Palaeomagnetic data from the Tres Cruces and Condoriaco study areas may be summarised as follows:

1. Statistically significant clockwise crustal rotations are identified between 29°30'-30°00'S of 10-16°. These rotations are markedly lower in magnitude than crustal rotations identified further north in the Coastal Cordillera/Precordillera region of northern Chile (as discussed in Chapter Three).

2. Crustal rotations observed in the Quebrada de Los Choros (Condóriaco) areas, occurred entirely post 70Ma (60Ma). This suggests that there is no early Cretaceous component of rotation.

3. The uniformity of rotation suggests that localised tectonics, operating at a scale <30-40km, are not involved in accommodating crustal rotation in either area.

7.2 Crustal rotation in a wider context: the Coastal Cordillera-Precordillera region (26-30°S)

7.2-1 Background

The Tres Cruces and Condoriaco study areas span the Coastal Cordillera-Precordillera boundary (Figure 5.1-2). The geology in these regions (west of the Domeyko Fault System) is, in essence, relatively invariant throughout northern Chile. The present day forearc region is an amalgamation of Early Jurassic to Eocene magmatic/volcanic arc (and their volcano-sedimentary equivalents), with successively younger arcs lying immediately east of the previous incarnation. Consequently there is a marked parallelism to the geology as a whole with each arc situated within a particular setting that is characteristic along the length of the
Dating of each arc tends to show a relatively short-lived event corresponding to emplacement over timescales of \( \leq 10 \text{My} \) with little real overlap. Therefore the age and location of a particular magmatic arc is generally well constrained and consistent over long distances. This means that each arc should be capable of preserving variations in the magnitude of crustal rotation along the strike of the margin.

The along strike continuity in geology presents an ideal opportunity to observe gradients or discontinuities within the overall pattern of crustal rotations. In contrast, the general trend of geology younging to the east makes the investigation of cross-strike (E-W) variation in crustal rotation slightly more problematic, as it becomes more complex to separate possible temporal and spatial components of the gross rotation pattern.

While considering the use of magmatic arcs for paleomagnetic purposes it should be noted that a major source of contention concerning the use of plutons surrounds the difficulty in establishing a reliable palaeohorizontal reference surface. In the absence of a suitable proxy (such as widespread magmatic layering with evidence of horizontality in the first place), it is difficult to test whether or not an igneous body has undergone significant post-emplacement post-magnetisation tilting. In general it is often assumed that such bodies carry a primary TRM (unless a remagnetisation can be proved), and anomalous directions caused by regional tilting could remain undetected. Tilting of plutonic bodies is considered to be one of the main problems in the long running debate involving the interpretation of palaeomagnetic data from the west coast of North America with respect to terranes being far travelled [e.g. Dickinson & Butler, 1998; Butler et al., 2001].
7.2-2 Description of the crustal rotation pattern between 26-30°S

Crustal Rotation recorded by magmatic arcs

As noted in Chapter Four, and above, the most striking aspect of the rotation pattern in the Tres Cruces area is the discontinuity in magnitude of rotations recorded by plutons Las Campañas (32.3°) and Corredores (10.2°) (Table 7.1-1, Figure 7.1-1). Several possible explanations exist;

a) Pluton Las Campañas records an anomalous direction (Corredores being in close agreement with other units from the immediate area),
b) Pluton Las Campañas is situated in a different structural block to Los Corredores and suffered a significantly larger rotation, or
c) Pluton Las Campañas has suffered very localised rotation.

To investigate this apparent discontinuity, data from the wider region, 26-30°S are considered.

Palaeomagnetic data from the Vallenar area (~28°S), situated immediately north of pluton Las Campañas, indicate an average of 30 ± 6° (one standard deviation) of clockwise rotation (Table 7.2-1). (This excludes one result from the La Higuera dyke swarm which intrudes pluton Camarones in the area of Vallenar and clearly records ~10° less CW rotation than the host pluton—possibly suggesting a later time of intrusion than previously interpreted). These data are therefore consistent with that from pluton Las Campañas. The two datasets together imply that there is a marked discontinuity in rotation magnitude between pluton Las Campañas and the main Tres Cruces sampling area. As with the Condoriaco-Quenbrada de Los Choros area there is a string uniformity of rotation between 28 and 29°S, reinforcing the view that localised structures exert very little control on the observed rotation pattern.
<table>
<thead>
<tr>
<th>Sampling Unit</th>
<th>Latitude (°S)</th>
<th>Longitude (°E)</th>
<th>Mean Direction Dec. (°)</th>
<th>Mean Direction Inc. (°)</th>
<th>Reference Pole a95 (°)</th>
<th>Rotation ± error (°)</th>
<th>Flattening ± error (°)</th>
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<td><strong>Early Cretaceous Arc</strong></td>
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<td>Remolino dykes&lt;sup&gt;(1)&lt;/sup&gt;</td>
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<td>289.70</td>
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<td>43.1 ± 12.3</td>
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<td>289.30</td>
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<td>5.6</td>
<td>90 Ma</td>
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<td>100 Ma</td>
<td>7.5 ± 12.8</td>
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<td><strong>Latest Cretaceous-earliest Paleocene Arc</strong></td>
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<td>Copiapina Granitic Pluton + aureole&lt;sup&gt;(3)&lt;/sup&gt;</td>
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<td>9.2</td>
<td>70 Ma</td>
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<td>Pluton Cabeza de Vaca&lt;sup&gt;(3)&lt;/sup&gt;</td>
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<td>289.90</td>
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<td>7.2</td>
<td>60 Ma</td>
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<td>-49.6</td>
<td>11.1</td>
<td>70 Ma</td>
<td>10.2 ± 14.1</td>
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Table 7.2-1  Overall mean directions, age of magnetisation and magnitude of crustal rotation (+ve clockwise) recorded by sampling units from the early Cretaceous, late Cretaceous and latest Cretaceous-earliest Paleocene magmatic arcs, sampled between 26-30°S. Data from this study and 1-Randall et al., (1996), 2-Gipson (unpublished PhD), 3-Taylor et al., (2007).
In order to further investigate the rotation hiatus between 28 and 29° palaeomagnetic data from the Early, mid and latest Cretaceous-earliest Paleocene magmatic arcs are compared (Figure 7.2-1, Table 7.2-1) as sufficient data is recorded from these three suites of rocks to make informed along strike comparisons.

Although plutons from the older magmatic Early (130-120Ma) and mid (~90Ma) Cretaceous arcs are not as densely sampled as those from the latest Cretaceous-earliest Paleocene arc (70-60Ma), the rapid decrease in rotation exhibited between plutons Las Campañás and Corredores is still observed. Both of the older arcs display a clear decrease in the magnitude of clockwise rotation between 28-30°S (Figure 7.2-1 c.f. plutons Camarones and Los Colorados, and Freirina and Tilgo). This confirms the observed ~20° decrease in rotation between plutons Las Campañás and Corredores and implies this discrepancy is confined to a relatively narrow belt of no more than ~50km (Figure 7.2-1) and that the blocks to north and south of this are of a substantial size. These blocks are hereafter referred to informally as the Tres Cruces and Vallenar blocks.

In comparison to the Tres Cruces and Vallenar areas clockwise, rotations of ~45° in magnitude, are observed north of ~28°S (Figure 7.2-1, Table 7.2-1). Again these rotations appear to be remarkably uniform with the latest Cretaceous pluton Copiapina (67Ma) [Taylor et al., 2007] recording an identical result to the early Cretaceous Las Tazas and Remolino dyke swarms [Randall et al., 1996]. Interestingly this not only suggests that crustal rotation occurred post-70Ma (as is the case to the south), but also that there is no apparent easterly decrease in rotation, at least at a scale which is detectable across the Coastal Cordillera and westernmost Precordillera. Furthermore given Copiapina’s location it confirms that the presence/absence of the Central Valley, present at the latitude of Copiapina
Figure 7.2-1 Map illustrating the magnitude of crustal rotations recorded by successive magmatic arcs in the present day forearc of northern Chile between 26-30°S. Data presented in Table 7.2-1. Arrows represent the amount of crustal rotation recorded by an individual sampling unit (i.e. a single pluton, dyke swarm or locality within a particular formation), each of which are identified on the corresponding stereonets. The age of magnetisation is colour coded as shown to the left of the map.
but absent south of Copiapó (Figure 1.3-2), makes no difference to the results observed. The dataset indicates that the transition between this northern domain, the Chañaral-Copiapó block, of rotations and the Vallenar block may be as equally sharp as that observed at 29°S, (Figure 7.2-1, Table 7.2-1).

Comparison of crustal rotations recorded by non-magmatic rocks

Although the plutons of the various magmatic arcs appear to record apparent discontinuities in the rotation pattern between 26-30°S, it is important to assess whether palaeomagnetic data from the plutons themselves are reliable and that the plutons themselves have not undergone significant post-magnetisation tilting.

In the Tres Cruces area plutons Los Colorados and Corredores are interpreted to record primary TRM's, although unfortunately neither display recognisable palaeohorizontal features. However the plutons record near identical amounts of crustal rotation as the tilt-corrected strata from the Arqueros Formation, implying tilt of this unit pre-dates their emplacement and that no significant post emplacement tilting has taken place or has affected the entire area uniformly. Comparison of the ‘non-batholith’ data with that from the ‘batholith’ between 26-30°S (Tables 7.2-1 & 7.2-2, Figures 7.2.1. and 7.2-2) shows there is little significant variation between the two datasets. This clearly implies that this part of the forearc has had little differential tilting post the latest Cretaceous-Paleocene arc episode of granite emplacement and/or that any such tilt is widespread, relatively uniform and of very small magnitude. Any slight differences in the data patterns of these two maps probably reflects a combination of the following-variable numbers of sampling sites in some units (perhaps not averaging out PSV completely leading to a certain level of palaeomagnetic noise), minor variations in rotation magnitude because of local structures enhancing or retarding the gross
<table>
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<th>Sampling Unit</th>
<th>Latitude (°S)</th>
<th>Longitude (°E)</th>
<th>Mean Direction Dec. (°)</th>
<th>Mean Direction Inc. (°)</th>
<th>Referencepole a95 (°) B &amp; C, 2002, 2003</th>
<th>Rotation ± error (°)</th>
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<td>289.04</td>
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</tbody>
</table>

**Table 7.2-2** Overall mean directions, age of magnetisation and magnitude of crustal rotation (+ve clockwise) recorded by remaining sampling units between 26-30°S. Data from this study and 1-Randall et al., (1996), 2-Taylor et al., (1996), 3-Gipson, (unpublished PhD), 4-Palmer et al., (1980a), 5-Taylor et al., (2007), 6-Riley et al., (1993).
Figure 7.2-2  Map illustrating the magnitude of crustal rotations recorded by successive magmatic arcs in the present day forearc of northern Chile between 26-30°S. Data presented in Table 7.2-1. Arrows represent the amount of crustal rotation recorded by an individual sampling unit (i.e. a single pluton, dyke swarm or locality within a particular formation), each of which are identified on the corresponding stereonets. The age of magnetisation is colour coded as shown to the right of the map.
rotation, or slight miscalculation of the rotation either because of age error in remagnetised units or the reference poles/directions themselves.

Overall the two datasets indicate little significant variation in the magnitude of crustal rotation across the margin, with only slight variation observed in the magnitude of rotations within the three blocks identified. This would appear to indicate that no gross errors exist with regard to the age of magnetisation assigned for each sampling unit, as comparison to an inappropriate reference pole should produce appreciable errors in the calculated magnitude of rotation.

7.2-3 The Pabellón Formation

The overall density of the rotation pattern established between 26-30°S has reached a sufficient level of detail as to allow the prediction of expected rotation at these latitudes. As discussed in Chapter Six, the Pabellón Formation was sampled southwest of Copiapó, at ~27.9°S in Quebrada El Molle, to the west of the Chañarcillo Fold and Thrust Belt (or Coastal Cordillera-Precordillera boundary), for magnetostratigraphic purposes. It was however found to carry a post tilt remagnetisation. The remagnetised Cerrillos Formation, sampled at four localities some 20-40km to the north and east, record rotations of 40-55° east of the Chañarcillo Fold and Thrust Belt (Figure 7.2-2, Table 7.2-2) [Taylor et al., 2007]. It is therefore predictable that the Pabellón Formation should record a similar magnitude of clockwise rotation. The results of Chapter seven were interpreted to be a remagnetised direction recording a clockwise rotation of 36.6° (± 4.3°-Table 7.2-2). Although of slightly smaller magnitude than recorded by the Cerrillos Formation to the north, it is slightly larger than observed from equivalent strata sampled from Quebrada del Cama ~100km to the south (Figure 7.2-2, Table 7.2-2).
7.2-4 Continuum v. discrete deformation

The distribution of rotations, as described above, does not appear to be consistent with a continuum-type description of deformation as discussed in Chapter Three. Instead it favours a discrete type mechanism for deformation with a small number of large, homogenously rotated blocks or domains (Figures 7.2-1 & 2). Three such blocks are identified; the Chañaral-Copiapó block with the largest rotations of 45-55° north of 28°S; the Vallenar block between 28-29°S with rotations of ~30°; and the Tres Cruces-Condoríaco block south of 29°S with rotations of 10-15°(Figures 7.2-1 & 2, Tables 7.2-1 & 2). The pattern is clearly one of discrete changes with relatively large blocks rather than one of gradational change. This does not however preclude distributed deformation over relatively substantial distances at the block boundaries themselves—at best these are limited to a distance of some 30-50km in width. These boundary zones clearly require further detailed investigation.

7.2-4 Accommodation of Crustal Rotation between 26-30°S

Clockwise crustal rotations in the Coastal Cordillera-Precordillera region of northern Chile show marked and stepped decreases in magnitude from north to south. The same pattern (west of the La Ternera Fault System) is observed irrespective of age (>60Ma), lithology or relative distance (inboard) from the present-day coastline/trench (Figures 7.2-1 & 2). This indicates that the observed rotation pattern is;

- a robust feature of the gross regional deformation to have affected the CC-PC of northern Chile, at least between 26-30°S;
- it is post 62 Ma in age;
• it is not a function of the margin parallel Atacama Fault System or the Chañarcillo Fold and Thrust belt which are too old to have affected the youngest Paleocene plutons at 62 Ma;
• it does not appear to involve differential shear at the observable scales although diffuse deformation at block boundaries is not precluded;
• three identifiable divisions are noted—the Tres Cruces-Condoriaco, Vallenar and Copiapó-Chañaral blocks, with respective average (CW) rotations of \(12.4^\circ \pm 4.3^\circ, 30.8^\circ \pm 5.6^\circ, 45.1^\circ \pm 7.8^\circ\) (all \(\pm\) one standard deviation).

With respect to the eastern limit of rotation, the pattern delineated in Figures 7.2-1 & 2 does not show evidence of a west-east gradient or discontinuity in the magnitude of rotation, either of which could mark such a boundary. This would imply that the eastern boundary of the observed rotation pattern lies further towards the High Cordillera region and will be addressed further with regard to palaeomagnetic data sampled from the La Guardia area in a later section.

In summary, the overall pattern appears to define rotational domains that appear reasonably free of localised rotations. Evidence suggests that the transition between these domains is associated with a relatively rapid change in the magnitude of rotation and deformation is concentrated within, albeit relatively, narrow zones of 30-50km at maximum.

**NW structures**

As has been argued above the rotation pattern is one of very large blocks that undergo relatively uniform rotation, with little internal (palaeomagnetically detectable) differential deformation and which are separated by reasonably discrete and narrow zones of deformation.
Although a single, large-scale fault is not found between plutons Las Campañas and Corredores, structural reconnaissance within the Tres Cruces area, carried out by Dr Mark Anderson (University of Plymouth), suggests that the local deformation pattern is dominated by sinistral strike-slip faults, orientated ~NW-SE (Figure 7.2-3a). If this albeit 'diffuse' deformation is taken at face value, a NW-SE corridor of sinistral strike-slip deformation may be inferred to take up the relative rotation between the two plutons. Whilst this (kinematically) fits the observed rotation pattern, there is no direct evidence to suggest that these faults were active post 60Ma, when crustal rotation is interpreted to have occurred. Further structural investigation of the Tres Cruces area is therefore required in order to establish whether or not such a zone of diffuse deformation is temporally and kinematically consistent with the observed rotation pattern. In the absence of such verification however, the deformation zone is tentatively proposed to represent the boundary between the two blocks and is referred to as the 'Tres Cruces Lineament'.

The proliferation of dominantly NW orientated faults is continued in the High Cordillera of the Vallenar region, where Rivera & Falcon (2000) recognise a major lineation striking 295°. Referred to as the 'Fortuna Lineament' it coincides with/defines the southern limit of the Yerbas Buenas Volcanic-Plutonic Complex (Figure 7.2-4), which was emplaced during the Paleocene-Eocene. This suggests that this lineament (fault) must have been active during or after this period, and was probably associated with left-lateral offset to accommodate the opening of the caldera field. The strike and sense of displacement of this lineament is almost

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1 A slightly obliquely trending set of normal faults is also observed in the Tres Cruces area (Figure 8.2-3b), but palaeostress analysis indicates that the two sets of faults are not directly related. The normal faults are therefore not discussed further.
Figure 7.2-3  Frequency diagrams illustrating the dominant NW-SE strike of lineated faults measured in the Tres Cruces area. A-NW striking sinistral strike-slip faults (red) and subordinate dextral fault set (blue) and B-NW striking normal faults (orange) and generally orthogonal reverse faults (green). Data collected by Dr Mark Anderson (University of Plymouth) from many localities throughout the Tres Cruces study area, to the east of the Atacama Fault zone (see inset map).
Figure 7.2-4 Diagram illustrating the extent of the caldera field that developed in the Copiapo region during the late Cretaceous. The southernmost boundary is interpreted by Rivera & Falcon (2000) to coincide with the Fortuna Lineament.
identical to that of the diffuse deformation observed in the Tres Cruces area therefore, lending credence to the importance of this fault corridor in the local deformation pattern. The fact that the Fortuna Lineament is believed to have been active during the Paleocene-Eocene would imply that these NW-lineaments are potentially at least temporally equivalent to the period of crustal rotation.

The location of the Fortuna Lineament as defined by Rivera & Falcon (2000), fits remarkably well between the Vallenar and Copiapó-Chañaral crustal blocks defined in the previous section. The fact that a there is a marked difference in the magnitude of crustal rotation to the north and south of this lineament, suggests that the Fortuna Lineament may have, effectively formed a block boundary fault between to rotating blocks, possibly accommodating a greater amount of shortening to the north. Whilst it is conceded that the locations of the Fortuna Lineament and proposed Tres Cruces Lineament may be purely coincidence, the apparently rapid change in the observed magnitude of crustal rotation affecting the area in close proximity to these lineaments, combined with the lack of recognition of any other potential ‘block bounding structures’, would strongly suggest that these NW-striking lineaments were associated with the accommodation of the clockwise crustal rotations observed in the present day forearc between 27-30°S.

The observation that NW orientated structures control (or are at least involved with) crustal rotation in Northern Chile is not a new idea. NW orientated sinistral strike-slip faults have previously been proposed by several authors, if not necessarily at such a large scale. Forsythe & Chisholm (1994) proposed that sinistral NW faults accommodate crustal rotations within a strike-slip duplex bounded to the west by the subduction trench and to the east by the Atacama Fault system (Chapter 3), whilst Randall et al., (1996) and Taylor et al., (1998) suggest that the eastern boundary of the duplex was represented by the Central
Valley Shear Zone (Chañarcillo Fold & Thrust Belt). Although both of these structures are now shown not to control the observed rotation pattern, the apparent importance of NW structures remains (e.g. Grocott & Taylor, 2002; Taylor et al., 2007).

The Importance of NW structures at a crustal scale

NW structures have been widely recognised to form a significant, if often-diffuse component of both the regional and margin-wide deformation pattern in northern Chile and NW Argentina and are shown to be extremely closely related to many of the world class copper porphyry (and other) deposits situated in the Andean Cordillera of Chile and Argentina [e.g. Richards et al., 2001; Chemicoff et al., 2002]. The strike of the Fortuna and Tres Cruces lineaments is similar to that of a suite of NW trending, albeit arcuate, trans-continental lineaments, suggesting that these lineaments may reflect the surface expression of the reactivation of a pre-existing crustal anisotropy (Figure 7.2-5-e.g. Salfity, 1985; Bassi, 1988; Chemicoff et al., 2002; Jacques, 2003a, b).

The origin of these NW faults and their role in the development/evolution of the Andean margin as a whole has been the subject of much recent debate. In a seminal paper, Salfity (1985) established that a number of regional transverse lineaments were very important features during the palaeogeographic evolution of the late Precambrian-Cenozoic basins of NW Argentina and inferred that these lineaments stretch to the Chilean coast to the west (Figure 7.2-5).

In his synthesis of the overall tectonostratigraphy of the sub-Andean basins and Andean Cordillera, Jacques (2003a & b) concludes that the differential amount of subsidence and uplift recognised between the Sub-Andean provinces has been taken up across broad trans-continental accommodation zones. These broadly
Figure 7.2-5  Distribution of trans-crustal NW faults within the present day forearc of Northern Chile [Figure 1 from Chemicoff et al., 2002, based on the original mapping by Salifty (1985). NW faults are observed to have acted as conduits during metallogenesis, with a number of economically important deposits identified at intersection between these NW faults and the major margin parallel fault systems such as the La Ternera Fault Zone.
defined lineaments probably reflect the repeated reactivation of pre-existing fault zones in the underlying basement and are hypothesised to have constrained the large-scale reorganisation of crustal blocks during break-up of the supercontinent Gondwana. Should these lineaments represent fundamental structures that have controlled the development of the Andean Cordillera, it is not unreasonable to extend their control to the observed pattern of crustal rotations.

7.2-5 Preferred Rotation Model (26-30°S)

As discussed previously, the NW structure identified in the Tres Cruces study area and the similarly orientated Fortuna Lineament to the north, exactly coincide with the pronounced discontinuities in the observed pattern of CW crustal rotations between 26-30°S (Figure 7.2-6). Although the rotated domains are interpreted as large blocks with rigid interiors, the transition zones between these blocks are most likely not sharply defined features, but rather broad zones of deformation <50Km in width. The precise nature of these transition zones could be readily tested by a concentrated palaeomagnetic sampling campaign, combined with a detailed structural investigation across one of the boundary zones identified.

The observation that identical rotations are recorded, irrespective of sampling lithology, age or location means that the linear magmatic arcs in particular are very suitable targets to investigate the pattern of crustal rotation across the block boundaries. If the block boundaries are truly discrete in nature, one would expect this to be reflected in the palaeomagnetic data, whereas continuum type deformation (at the block boundaries) may be expected to produce a smooth gradient in the observed magnitude of rotation.
Figure 7.2-6 Preferred rotation model proposed to explain the large domains of homogenous clockwise rotations. The overall model is analogous to the slat-rotation models discussed in Chapter Three, except that it operates at a crustal scale, rather than in a localised shear zone setting. The location of the faults and shaded areas are for illustrative purposes only, reinforcing the fact that the lineaments are recognised as very broad zones of diffuse deformation.
7.3 La Guardia Study Area

7.3-1 Introduction

As discussed in Chapter Six, well-defined magnetisation directions have been determined from all sampled units in the La Guardia area (Figure 5.1-2) and where appropriate combined with data from Riley et al., (1993). The magnitude of crustal rotation recorded by each sampling unit has been calculated following the method discussed in Chapter Two (Table 7.3-1). In comparison to the simple primary magnetisations observed from the Tres Cruces study a rather more complex magnetisation history has been determined for the La Guardia area. This history includes both primary (pre-) and secondary (post-tilt) magnetisations (Table 7.3-1).

While there is little evidence to suggest that the margin parallel Atacama fault system and Chañarcillo Fold and Thrust Belt are directly involved in controlling the magnitude of crustal rotations in the Coastal Cordillera-Precordillera region, the La Ternera Fault System (LTFS) of the Precordillera is clearly implicated in the deformation and at the very least appears to have modified the observed pattern of crustal rotation in the La Guardia area.

7.3-2 Primary Mesozoic Magnetisations

La Ternera Formation

As the oldest (non-plutonic) strata in the La Guardia study area, the La Ternera Formation should carry a record of the gross rotation to have affected the Chilean Precordillera at this latitude during the past 200Ma, assuming that a primary remanence is retained. Although strata have been sampled at three distinct localities, the results indicate that it is more appropriate to discuss the overall palaeomagnetic dataset in terms of two geographically and structurally distinctive
<table>
<thead>
<tr>
<th>Sampling Unit</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Mean Direction</th>
<th>Reference Pole</th>
<th>Rotation</th>
<th>Flattening</th>
</tr>
</thead>
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<tr>
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<td></td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>Rio Aguas Blancas Area (*1)</td>
<td>27.67</td>
<td>290.59</td>
<td>30.2</td>
<td>-51.0</td>
<td>8.2</td>
<td>200 Ma</td>
</tr>
<tr>
<td>Banderitas Area (2*)</td>
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<td>290.40</td>
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<td>-52.4</td>
<td>7.4</td>
<td>200 Ma</td>
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<td>Lagunillas Formation (2*)</td>
<td>28.02</td>
<td>290.04</td>
<td>156.8</td>
<td>53.1</td>
<td>9.8</td>
<td>160 Ma</td>
</tr>
<tr>
<td>Quebrada Monardes Formation</td>
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<td>290.47</td>
<td>197.7</td>
<td>53.5</td>
<td>8.1</td>
<td>60 Ma</td>
</tr>
<tr>
<td>Quebrada Seca Formation (1*)</td>
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<td>290.38</td>
<td>196.3</td>
<td>49.3</td>
<td>12.9</td>
<td>60 Ma</td>
</tr>
<tr>
<td>Caldera Jorquera</td>
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<td>17.3</td>
<td>-57.0</td>
<td>9.5</td>
<td>60 Ma</td>
</tr>
<tr>
<td>Pluton El Gato</td>
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<td>4.4</td>
<td>-57.2</td>
<td>4.8</td>
<td>40 Ma</td>
</tr>
</tbody>
</table>

Table 7.3-1 Overall mean direction and magnitude of crustal rotation recorded by sampling units from the La Guardia study area. The shaded direction represents the very localised crustal rotation recorded by the remagnetised La Ternera strata sampled from the Fundo Santa Rosa and Manflas Plantation areas.
domains [Iriarte et al., 1993]. The two areas sampled to the north appear to be affected by different generations of faulting which allow some insight into the stability of the magnetic remanence with respect to deformation. As the Fundo Santa Rosa-Manflas area is interpreted to have undergone a remagnetisation, the magnitude of crustal rotation recorded in this area will be discussed in a following section with the other remagnetisations identified in the La Guardia area.

The northern domain is located in the northeast quadrant of the La Guardia area and includes data from the Banderitas and Rio Aguas Blancas localities [this study and Riley et al., 1993]. The dominant structural features in this area are a series NE-SW striking, steep, reverse faults that also display a significant component of sinistral (strike slip) displacement. These faults are strands of the wider La Ternera (Domeyko-West Fissure) Fault System, and may be older structures reactivated during the compressive, Eocene-Oligocene, Incaic orogeny. As such these structures divide the La Guardia area into a series of horst and graben like features and the Banderitas and Rio Aguas Blancas areas are separated by such a structure.

Although situated in different structural blocks separated by strands of the LTFS, the Banderitas and Rio Aguas Blancas localities record near identical pre-tilt magnetisation directions (Table 7.3-1). Both areas are therefore interpreted to record overall magnetisations that were acquired contemporaneously and are interpreted not to have undergone differential crustal rotation with respect to each other.

When individual site mean directions are plotted on the geological map of the area, local anomalies in the overall pattern are observed in both areas, with directions observed to undergo a number of swings in declination suggesting an element of localised differential shear deformation is recorded (Figure 7.3-1). In
Site by site rotation pattern recorded by the Triassic La Ternera Formation in the La Guardia study area. Blue (orange) arrows indicate a primary (secondary) magnetisation. Note that all three sampling localities appear to record some evidence of shearing. Fundo Santa Rosa-Manflas area records a secondary remagnetisation although the tilt-corrected rotations (Blue arrows) are shown for comparison (see text for details). Geology is as shown in Figure 5.1-2. As there is no recent mapping to the immediate south and east of the La Guardia mapsheet, sampling localities are projected on to a section of LandSAT 7 ETM+ imagery, displayed using Red (Band 7), Blue (Band 4) and Green (Band 2) channels. Generally speaking the La Ternera Formation appears as dark blue to purple in colour, with the underlying Lagunillas and Quebrada Monardes sandstones appearing dark green in colour.
the Banderitas sampling area for example, significant anti-clockwise deflections are recorded by sites AT3-54 & 55 (i.e. these sites record less CW rotated directions), that are in close proximity to a mapped fault belonging to the LTFS. A similar pattern is observed at sites AT3-59 & 60 although no fault is mapped, but it is entirely likely to have been the result of localised deformation. The discontinuous nature of the site mean directions from the Banderitas and Rio Aguas Blancas areas (c.f. Figure 7.1-1) is therefore ascribed to the effect of localised structures, related to displacement along the LTFS.

Both the Banderitas and Rio Aguas Blancas areas record ~40° of (gross) CW rotation (Table 7.3-1), which is approximately equal to the magnitude of rotation recorded by the remagnetised Cerrillos Formation in the Coastal Cordillera-Precordillera boundary region (Section 7.2). Although a localised component of shear is evident from the distribution of individual site mean directions, the similarity in the gross magnitude of crustal rotation both within and to the west of the LTFS would suggest that:

1. Despite the presence of numerous strands of the LTFS they appear to exert little or no control over the gross rotation pattern in this area;
2. deformation along the LTFS post-dates the majority of crustal rotation;
3. the LTFS cannot therefore define the eastern 'master-fault' bounding the overall rotation pattern observed in the Coastal Cordillera-Precordillera boundary zone;
4. the large domain of homogenous crustal rotation identified in the previous section as the Copiapo-Chanaral block (Section 7.2), must therefore extend to the east of the LTFS.
Lagunillas Formation

As observed for the La Ternera Formation in the north eastern part of the La Guardia area, the continental sandstones sampled from the Lagunillas Formation also record pre-tilt magnetisations in the form of a primary DRM and a high temperature component carried by haematite, recording an anti-parallel direction, interpreted as a CRM acquired soon after deposition. The overall compound ChRM direction records 36.7° of clockwise rotation (Table 7.3-1), similar to that recorded by the La Ternera Formation. The two formations are therefore interpreted to record identical rotation histories.

7.3-2 Secondary Cretaceous-Paleocene magnetisations

Fundo Santa Rosa-Manflas area

The southern structural domain sampled from the La Ternera Formation, situated in the southwest quadrant of the La Guardia study area includes data from the Fundo Santa Rosa and Manflas localities. Although a western boundary is not established, the homoclinal dip of the sampled strata suggests that these areas are situated within a single structural block, delimited to the east by the N-S striking Manflas and Tres Chanares Fault (Figures 5.1-1 & 2).

In contrast to the NE striking reverse faults observed in the northern sampling domain, here the faults are of normal displacement (e.g. the Quebrada Las Vacas fault) and display normal displacements in addition to an overall sinistral strike-slip component [Iriarte et al., 1999]. Whilst the age of these faults is not clear they are truncated by the Eocene aged Pluton El Gato [Iriarte et al., 1999], suggesting they pre-date its intrusion (i.e. pre 42Ma). While it is likely that these faults have suffered some inversion their normal displacement remains obvious.
While palaeomagnetic data of the La Ternera Formation in the northern sampling areas record a pre-tilt remanence, the overall magnetisation observed from this more southerly area, is complicated by the observation of an intense haematite overprint, that post-dates deformation. Individual site mean directions from the southern sampling areas are plotted in both in situ and tilt corrected coordinates for comparison (Figure 7.3-1). The tilt corrected inclinations are clearly far too steep for this latitude to be considered reasonable measurements of the ancient geomagnetic field and the magnetisation is therefore believed to be an in situ remagnetisation. The fanning of the in situ declinations (orange in Figure 7.3-1) are best interpreted as displaying the effects of distributed shear, albeit recording an overall anticlockwise crustal rotation of 16.0° (Table 7.3-1) (in stark contrast to the overall clockwise rotations observed throughout this part of Chile).

Randall et al., (2001) published data from a small sampling area that recorded an anticlockwise rotation, from the Quebrada Monardes Formation (Mina Vieja locality), collected from very close to the Sierra Castillo-Agua Amarga Fault system close to the small town of Porterillos. The anticlockwise rotation, in an area otherwise dominated by clockwise rotations, was interpreted to result from localised deformation as a consequence of differential shearing/local pinning on a thrust sheet within the fault system, which forms part of the LTFS (Eocene-Oligocene). A similar interpretation is consistent with the results from the Fundo Santa Rosa-Manflas localities.

The identification of an anticlockwise rotations in the Fundo Santa Rosa-Manflas area, the Mina Vieja locality of Randall et al., (2001) and the suggestion of similar albeit very localised rotations in the Banderitas and Rio Aguas Blancas areas may point to such rotations being a consequence of proximity to strands of the LTFS.
Quebrada Monardes Formation

Palaeomagnetic data from these continental sandstones fail a tilt-test indicating the ChRM is a post-tilt remagnetisation. The age of this event is difficult to ascertain, but is likely to be no older than the deformation best observed to the west in the Chañarcillo and Paipote areas, resulting from sinistral transpression during the late Cretaceous (77Ma-Arevalo, 1994; 66Ma-Matthews et al., 2001). Iriarte et al., (1999) interpret the easterly vergence of folds affecting Early to Late Mesozoic strata (Figure 5.1-5) to represent an extension of the Chañarcillo Fold & Thrust Belt. This therefore suggests that remagnetisation of the Quebrada Monardes sandstones occurred post-66Ma. Assuming a 60Ma remagnetisation, the sandstones record 24.9° of clockwise rotation when compared to the expected direction at this time (Table 7.3-1). Whilst the choice of a younger (possibly as young as 35 Ma) reference pole has some effect concerning the expected inclination at the sampling locality, there is negligible effect on the expected declination and hence the overall magnitude of rotation recorded.

The rotation recorded by the Quebrada Monardes Formation is substantially lower than that recorded by the Lagunillas sandstones and La Ternera Formation in the north-easteren part of the La Guardia study area. The close proximity of the sampling localities (Figure 5.1-2), suggests that approximately 20° of the clockwise rotation recorded by the older sampling units, must have been accommodated prior to the remagnetisation of the Quebrada Monardes Formation.

The differing primary/secondary nature of the ChRM's calculated from the Lagunillas/Quebrada Monardes Formations respectively, is unexpected given their lithological similarities. As discussed in Section 5.1, Iriarte et al., (1999), suggest that the two red continental sequences are separated by a period of Jurassic volcanism that produced a sequence of lavas traceable across much of the La
Guardia field area. As a speculative suggestion it is possible that this lava sequence could have acted as an aquitard to the Lagunillas sandstones and prevented the ingress of any orogenic or magmatic/volcanic fluids and perhaps even enhanced fluid flow through the highly permeable Quebrada Monardes Formation hence leading to the observed differences in magnetisation history.

**Quebrada Seca Formation**

This formation clearly records a post-tilt magnetisation, as observed for the Quebrada Monardes sandstones on which it rests conformably. The mixed volcanic/sedimentary strata records a clockwise rotation equal in magnitude to the Quebrada Monardes Formation (23.5° (compared to the 60Ma reference pole of Besse & Courtillot, 2002, 2003)-Table 7.3-1), suggesting the two Formations were remagnetised at the same time.

7.3-4 Primary Paleocene-Eocene Magnetisations

**Caldera Jorquera**

The intracaldera pyroclastic and lava units sampled from Caldera Jorquera record an in situ (primary) magnetisation inferring 24.5° of clockwise rotation (Table 7.3-1). The caldera is one of a number of discrete volcanic centres, nested within and on the periphery of the much larger and older Megacaldera Carrizalillo, that were active between 62-55Ma (Iriarte et al., 1999). Caldera Jorquera therefore provides an effective marker with which to evaluate the temporal accumulation of crustal rotation in the Precordillera of the Copiapo region. The similarity in the magnitude of rotation recorded by the primary magnetisation recorded by Caldera Jorquera and the remagnetised Quebrada Monardes and Quebrada Seca Formations, suggests that the timing of their remagnetisation was close to that of caldera construction.
Like Caldera Jorquera, the mid-Eocene intrusion Pluton El Gato records a primary magnetisation, with no evidence that it has been subjected to significant tilting as the overall inclination recovered from the pluton agrees with the expected inclination at the time of emplacement (c.42Ma average of four K-Ar dates). For this reason the pluton represents a second temporal marker with which to evaluate the timing of crustal rotation in the La Guardia area. The pluton records 7.8° (± 7.8°) of clockwise rotation, significantly less than recorded by the older sampling units. This suggests that the majority of crustal rotation was completed prior to the intrusion of the pluton.

7.3-5 Cause of Remagnetisation and Timing of Rotation in the La Guardia Area

The intrusion of pluton El Gato, a sheet-like intrusion, provides a potential mechanism for the generation and expulsion of hot magmatic fluids into the immediately surrounding country rock and hence a potential source of the remagnetisation affecting the Quebrada Monardes and Quebrada Seca Formations. Although this is a plausible mechanism to produce a very localised remagnetisation event, the obvious difference in the magnitude of CW rotation recorded by pluton El Gato and the ~25° of CW rotation recorded by the Quebrada Monardes and Quebrada Seca Formations and Caldera Jorquera (Table 7.3-1), when compared to the 40Ma reference pole [Besse & Courtillot, 2002, 2003], indicates that remagnetisation occurred prior to the intrusion of pluton El Gato.

The period of caldera construction during the early Paleocene, responsible for the creation of the Megacaldera Carrizalillo and the numerous smaller calderas nested within it and on the rim provides an alternative and slightly longer-lived event that could have remagnetised the Cretaceous strata in the La Guardia area.
Palaeomagnetic data collected from Caldera Jorquera, situated to the east of the main mega-caldera structure, indicates a primary (60Ma) magnetisation is isolated, recording 24.5° of clockwise rotation (Table 7.3-1). When compared to the same reference direction, the Quebrada Monardes sandstones and Quebrada Seca Formation record 24.9° and 23.5° of CW rotation respectively, identical to that recorded by Caldera Jorquera (Table 7.3-1). This strongly suggests that the remagnetisation event is either a consequence of, or otherwise of similar age to, the late Paleocene phase of intracaldera formation/collapse, responsible for the creation of Caldera Jorquera [62-55Ma-Iriarte et al., 1999].

The timing of the main episode(s) of CW rotation in the Precordillera of northern Chile is further constrained by the magnitude of rotation determined for Pluton El Gato. The pluton has been dated as being early Eocene in age [~42 Ma-Iriarte et al., 1999], and records conspicuously less CW rotation than any of the strata into which it intrudes (Table 7.3-1, Figure 7.3-2). This indicates that the intrusion of Pluton El Gato either post-dates or came towards the end of the main phase of deformation responsible for producing the rotation pattern evident throughout the La Guardia and wider region. The difference in magnitude of clockwise rotation recorded by Caldera Jorquera and Pluton El Gato would imply that the La Guardia area underwent 15-20° of clockwise rotation during the period 55-42 Ma, subsequent to the construction of Caldera Jorquera, but prior to the emplacement of Pluton El Gato (Figure 7.3-2a). This implies that the bulk of CW rotation was completed prior to the onset of the Incaic Orogeny, c.38Ma.

7.3-6 Summary

Crustal rotations determined from the La Guardia field area provide considerable evidence to suggest that the bulk of crustal rotation in the Copiapo region of the Chilean Precordillera was completed subsequent to the remagnetisation of the
Figure 7.3-2  Inferred timing (A) and overall pattern (B) of crustal rotation recorded by the sampling units from the La Guardia area. Blue (orange) arrows represent primary (secondary) magnetisations. Rotations are colour coded with respect to the inferred age of magnetisation. Geology is as shown in Figure 5.1-2. As there is no recent mapping to the immediate east and south of the La Guardia mapsheet, sampling localities from the La Ternera, Lagunillas and Quebrada Monardes Formations are projected on to a section of LandSAT 7 ETM+ imagery, displayed using Red (Band 7), Blue (Band 4) and Green (Band 2) channels. The La Ternera Formation appears as dark blue to purple in colour, with the underlying Lagunillas and Quebrada Monardes sandstones appearing dark green in colour.
Quebrada Monardes and Quebrada Seca Formations (c.62-55Ma), but prior to the onset of Incaic deformation (responsible for deformation along the La Ternera Fault System-c.38Ma), as demonstrated by the small magnitude of crustal rotation recorded by Pluton El Gato (intruded c.42Ma) (Figure 7.3-2). This necessarily implies therefore that crustal rotation in northern Chile is not a direct result of the main period of Andean mountain building/crustal thickening during the late Eocene-Oligocene and Miocene (Incaic and Quechuan Orogenies). In addition, it appears that a significant component of rotation (~20°) was accommodated prior to the remagnetisation of the Quebrada Monardes Formation in the La Guardia area, as suggested by the large clockwise rotations recorded by the oldest Mesozoic strata sampled.

The magnitude of clockwise rotation recorded by the La Ternera and Lagunillas Formations is approximately equal in magnitude to that recorded by the remagnetised Cerrillos Formation, situated directly to the west in the Coastal Cordillera-Precordillera region (Section 7.2-Taylor et al., 2007). The LTFS is therefore considered to have superimposed only very localised rotations upon a regional rotation pattern that extends to the east of the La Guardia area, as demonstrated by the anticlockwise rotation recorded by the remagnetised La Ternera Formation in the Fundo Santa Rosa-Manflas area.

7.4 Overall Crustal Rotation Pattern in Northern Chile

7.4-1 Timing and magnitude of crustal rotation

Palaeomagnetic data from the Coastal Cordillera-Precordillera boundary zone between 26-30°S indicate that regionally consistent, at least within the context of large scale blocks with width dimensions of the order of 100km or more, clockwise rotation occurred post 70Ma if not 62Ma. The timing of rotation in the La Guardia
study area is more precisely bracketed by data from Caldera Jorquera and Pluton El Gato, which infers that the Precordillera region east of Copiapó underwent 15-20° of clockwise crustal rotation between 55-40Ma. There was therefore a further 15-20° of regional clockwise crustal rotation between 70-55Ma to account for the largest palaeomagnetically determined rotations from the oldest strata, and to explain the larger palaeomagnetic rotations observed in the Coastal Cordillera-Precordillera area west of La Guardia. This suggests that the overall regional rotation affecting this area was probably accumulated over a prolonged period between 70 and 40Ma (or at a minimum 62-42Ma), resulting either from a gradual, continuous process, or a number of separate, short-lived rotation events that accumulated during this time window.

The clear implication of this is that regional rotation in northern Chile between 26 and 30°S occurred prior to the Incaic orogeny (c.38Ma). This necessarily implies that 'Andean' uplift and crustal thickening during the Oligocene and Miocene is not the primary cause of crustal rotation at these latitudes, as originally suggested by Isacks (1988) and others to explain the overall Central Andean Rotation Pattern.

The bulk of crustal rotation in the present day forearc of northern Chile must therefore be associated with an older deformational event or events.

The quantitative study of the effects of older deformational events in the Andes is severely hampered by the overprinting of younger, more recent orogenic processes, in part at least due to the large magnitude of the most recent phases of deformation (i.e. substantial crustal thickening and plateau uplift). Add to this the co-planar and co-linear nature of much of the deformation as a consequence of the, essentially, tectonic stability of the margin in terms of long term E-NE directed oblique subduction and it becomes very difficult to isolate individual deformational events in the broader framework. The maximum amount of shortening across the
Andean margin has been estimated as >320km [Schmitz, 1994], from the trench in the west to the easternmost effects of the fold and thrust belt in the eastern sub-Andes and this is interpreted to account for as much as 70-80% of the overall crustal volume [Arriagada et al., 2005]. The remaining 20-30% has been variously interpreted to be due to magmatic addition or tectonic underplating, but could also be explained by crustal shortening prior to the Neogene.

*The Coastal Cordillera-Precordillera Fault System & Chañarcillo Fold & Thrust Belt*

Comejo & Matthews (2001) and Matthews et al., (2001) undertook an extensive mapping and dating campaign within the (latest Cretaceous-earliest) Paleocene magmatic arc formed at the Coastal Cordillera-Precordillera Fault system (incorporating the Chañarcillo Fold and Thrust Belt), which forms the western margin of the Hornitos basin north of Copiapó. This showed that the earliest phase of plutonism was contemporaneous with a short-lived but regionally consistent contractional event between 66-64Ma, but possibly as early as 70Ma, which was identified in the Porterillos area (c.26° 30'S) lying to the north of the La Guardia area [Comejo et al., 1997].

A slightly older deformation event has also been described in the Copiapó valley, observed to affect the Early Cretaceous and older strata. This (supposedly older) deformation is associated with a period of sinistral transpression along the Coastal Cordillera-Precordillera Fault System [Arevalo & Grocott, 2000; Grocott & Taylor, 2002; Taylor et al., 1998] and was believed to have occurred between 93-78Ma [Arevalo & Grocott, 2000] (based on a single K-Ar date) and was therefore associated with the Peruvian Orogeny of the late Cretaceous [Coira et al., 1982].

The marked parallelism of the structures (not withstanding the comments about co-planar, co-linear deformation made above) but more importantly the age of the the units affected does not preclude that the two events may in fact be one and the
same and to have occurred at ~65Ma. This supposition would be consistent with the easterly vergent set of folds deforming the Mesozoic strata of the La Guardia area [Iriarte et al., 1999] as part of the same regional compressive event. This event would therefore mark the onset of deformation by rotation.

7.4-2 Extent of CW rotations in northern Chile

The northern Chilean forearc has been sampled in similar detail to the Copiapó area to the north and east of Antofagasta (22°45'-23°30’S), where several studies have documented the widespread occurrence of clockwise crustal rotations throughout the Coastal Cordillera, Central Valley and Precordillera/Cordillera de Domeyko (Figure 7.4-1) [e.g. Arriagada et al., 2000, 2003; Hartley et al., 1988, 1992; Somoza et al., 1999; Somoza & Tomlinson, 2002]. As observed between 26 & 30°S, there is no obvious gradient in the magnitude of rotation across the margin, confirming that the magnitude of rotation is not a function of distance from the trench and suggesting that the observed rotations result from discrete-type deformation.

Arriagada et al., (2006) suggest that the Copiapó and Antofagasta regions record clockwise crustal rotations that are similar in magnitude if not timing. As observed between 26 & 30°S, the magnitude of rotation recorded by Jurassic-Cretaceous aged units sampled in the Antofagasta region, are not systematically larger than those recorded by latest Cretaceous-Paleocene units, suggesting that no rotation occurred prior to the Paleocene, with an overall clockwise rotation of 30.8° (± 17.9°) recorded in the area [Arriagada et al., 2003]. As suggested in Section 8.2 however, early Mesozoic and late Mesozoic-Paleocene units are not generally found in the same block due to the eastward migration of the active
Figure 7.4-1  Crustal Rotations from the Antofagasta region [from Somoza et al., 2003]. Localised variations in the magnitude of rotation are attributed to dextral shear accommodated along the NE trending Antofagasta-Calama lineament.

Figure 7.4-2  Age of magnetisation plotted against the magnitude of crustal rotation recorded by sampling units from the present day forearc of Northern Chile between latitudes of 22-30°S. Mesozoic-green, Paleocene-blue, Neogene-red. Error bars calculated using the method of Demerest, (1982). Yellow circles indicate localised rotations clearly associated with displacement along the DFS that represent outliers in the overall dataset. Data is sourced from all published palaeomagnetic studies from the present day forearc region between 22-30°S-see Appendix B for references.
magmatic/volcanic arc. For this reason it is not possible to definitely rule against the possibility that some rotation occurred during the Cretaceous period.

Distribution of crustal rotations in the Antofagasta region

In the Antofagasta area, Arriagada et al., (2003) suggest that clockwise rotation occurred in response to differential arc normal shortening along the strike of the Andes. In contrast to the NW orientated relay zones identified to separate domains of differential clockwise rotation between 26-30°S, Arriagada et al., (2003) propose that the Antofagasta area is dominated by the NE trending Antofagasta-Calama Lineament (ACL). The ACL would have acted as a transfer zone accommodating significantly greater shortening to the north than to the south, as suggested by an apparent dextral offset of Mesozoic strata (Figure 7.4-1). Whilst this interpretation may be locally tenable, the complete lack of similarly orientated structures throughout the Copiapó area, combined with the observation that at least 25°of clockwise rotation is recorded in the area between Antofagasta and Copiapó [e.g. Arriagada et al., 2006; Forsythe & Chisholm, 1994], suggests that overall the area between 23-30°S has undergone a reasonably uniform clockwise rotation. Given the absence of such NE striking structures from the rest of the area it is clear that rotation throughout the region cannot be explained by this single feature itself.

The ACL is the most southerly of the NE trending lineaments identified by Bassi (1988), Salfity (1985) and Jacques (2003a & b) which to the north are thought to be significant controls on the the deformation pattern throughout the northern Central Andes. While NE structures cannot explain the overall pattern of clockwise rotations between 23-30°S, they may present a possible explanation for the large error associated with the average clockwise rotation (standard deviation=17.9°) determined for the Antofagasta area by Arriagada et al., (2003). The much larger standard deviation (in comparison to that for the mean rotation south of 25°S) may
reflect more localised, small block rotations, resulting from the interference of
(sinistral) NW orientated structures to the south of ~23°S and (dextral) NE
orientated structures to the north

Timing of crustal rotation in the Antofagasta region

Due to the nature of the outcrop pattern in northern Chile, very little Neogene aged
material suitable for palaeomagnetic sampling is present within the forearc, but
what is sampled records little if any crustal rotation. Both Arriagada et al., (2003)
and Somoza et al., (1999) sampled the Sifon ignimbrite, a c.8Ma pyroclastic
deposit that is laterally traceable for some considerable distance. Although the
ignimbrite records an anomalous direction, interpreted to be associated with an
excursion of the geomagnetic field [Arriagada et al., 2003], no significant rotation
between sites is recorded, inferring that no internal block rotations or deformation
have occurred in the Antofagasta region of northern Chile since the emplacement
of the Sifon ignimbrite. Arriagada et al., (2003) therefore proposed the onset of
rotation to have commenced in Eocene time, as part of the Incaic orogeny, and to
have ended sometime in the Miocene, markedly younger than the interpretation
herein based on the La Guardia area.

The overall similarity in the extent, magnitude and timing of rotations in both the
Copiapó and Antofagasta regions, suggests that the overall pattern of crustal
rotation along the Chilean margin (at least between 22-30°S) developed in
response to a single event during the Tertiary. This is concisely demonstrated
when the magnitude of rotation is plotted against the interpreted age of
magnetisation for all of the palaeomagnetic data from the present day forearc
region between 22-30°S (Figure 7.4-2). Crustal rotations recorded by sampling
units with magnetisations >50Ma, are generally observed to be between 20-50°
clockwise, while magnetisations <40Ma are generally statistically insignificant. A
certain level of 'noise' is interpreted to result from the effects of purely localised deformation, with a number of outlying data-points sampled from within the La Ternera Fault System (Figure 7.4-2).

In summary, the entire northern Chilean forearc region between 22-30° is considered to record a single episode or period of rotation, the magnitude of which is observed to be remarkably consistent throughout the Coastal Cordillera-Precordillera region. Most notably this rotation event pre-dates the Incaic Orogeny (c.38-39Ma), which infers that crustal rotation in northern Chile was not driven by the mountain uplift and/or substantial crustal thickening associated with this event. The data presented herein instead suggests that rotation accumulated in the period 60-45Ma as documented by changes in magnitude of rotation in the La Guardia area.

**Northern limit of clockwise crustal rotations**

The Antofagasta-Calama Lineament represents the northern limit of NW and sinistral faults and lineaments being dominant over NE and dextral structures with the converse being true to the north [Bassi, 1988; Jacques, 2003a & b; Salfity, 1985]. It therefore would be reasonable to expect there to be a clear discontinuity in the overall rotation pattern in the forearc region if NW structures control clockwise rotations between 23-30°S. This appears to be the case, with a clear discontinuity observed in the overall rotation pattern derived from palaeomagnetic studies throughout the Central Andes (Figure 7.4-3). To the north of ~22°S, palaeomagnetic data from Mesozoic-Neogene aged material in the forearc record little or no rotation [Somoza et al., in-prep], in contrast to the relatively uniform, large magnitude crustal rotations recorded between 23 & 29°S. It is interesting to note that this change coincides with the well-documented change in crustal structure corresponding with the transition from the low relief Altiplano...
Figure 7.4-3 Overall crustal rotation pattern throughout the Central Andes. Box indicates the overall study area from this study, circle indicates unrotated Mesozoic-Neogene sampling units in the present day forearc in the Antofagasta region to the north of 22°S. Green arrows represent magnetisations acquired during the Mesozoic, blue-Paleogene, red-Neogene. Magnitude of rotation indicated relative to north. Data from various studies—see appendix B for references.
plateau to the north and high relief Puna plateau to the south [e.g. Allmendinger &
Gubbels, 1996; Allmendinger et al., 1997; Whitman et al., 1996].

One interpretation of this marked topographic/crustal transition is that pre-Andean
crustal inhomogeneities in the foreland indirectly controlled the along-strike
variation in the magnitude and style of shortening in the Central Andes
[Allmendinger et al., 1997]. This led to a diachronous deformation history, with
uplift initiating in the Altiplano segment at ~25Ma, followed by deformation in the
Puna segment 5-10 million years later [Allmendinger et al., 1997]. While this
differential deformation was associated with Miocene-recent uplift it might be
reasonable to assume that this fundamental crustal scale anisotropy may have
controlled crustal rotation during a previous stage of deformation. Further
investigation in this area is required to investigate the regional rotation pattern.

7.4-3 The 'Central Andean Rotation Pattern'

The gross pattern of crustal rotation throughout the Central Andes has long been
regarded as reflecting a continuous change from anticlockwise rotations to the
north of Arica (c.19°S) to clockwise rotations to the south. Between 19-21°S, very
little palaeomagnetic data exists from the forearc region itself, but data from the
wider Andean Cordillera indicates that Mesozoic-Neogene rocks in the High Andes
and Eastern Cordillera do indeed appear to record a continuous rotation pattern
along the Andean margin (Figure 7.4-3). The magnitude of these rotations is
approximately equal to the 10-15° of anticlockwise (clockwise) rotation to the north
(south) of the Arica Deflection, as predicted by Isacks (1988) to result from
differential shortening along the Andean margin during Neogene uplift.

The present day forearc between 23-29°S however, represents an anomalous
'domain' of much larger magnitude clockwise rotations than predicted by Isacks'
(1988) model, that is not consistent with the otherwise continuous rotation pattern observed elsewhere throughout the Central Andes. This area of the forearc is therefore interpreted to represent a separate 'terrane' within the Central Andes, in the sense that it has undergone a differential translation with respect to stable South America. It is important to stress that this differential translation is not observed to have involved significant latitudinal displacement and therefore the term 'terrane' is used advisedly. With regard to the description of the overall distribution of crustal rotation within the Central Andes, it is therefore inappropriate to refer to a 'Central Andean Rotation Pattern' (CARP), which implies that crustal rotations form a continuous distribution throughout the Central Andes, when the actual distribution has been demonstrated to be more complex.

The accommodation of crustal rotation in the present day forearc region of the southern Central Andes, appears to be best described in terms of a 'domino-type' rotation mechanism, involving the in situ rotation of large rigid blocks. As an analogue this is similar to the situation modelled in Figure 3.2-7d, with large blocks whose width approximates the width of the deforming margin (now recognised as the present day forearc), rotating within a domain bounded by the subduction zone to the west and a margin parallel fault system to the east. Rotation was accommodated via strike-slip displacement along margin oblique (NW-orientated), block boundary faults, that are interpreted to represent pre-existing crustal anisotropies that manifest as broad zones of highly diffuse deformation.

As discussed in Chapter Three, Abels & Bischoff (1999) propose a very similar model, albeit based on substantially less palaeomagnetic data (Figure 3.5-8). They suggest that the transcrustal lineaments identified by Salfity (1985) provide the structural framework within which domains of homogenous crustal rotation are accommodated. The effects of the Domeyko Faults System in the La Guardia area
appear to suggest that its effects are extremely localised, whilst the oldest (La Ternera) strata situated to the east of the DFS record large magnitude clockwise rotations of similar magnitude to those observed immediately to the west in the Central Valley and Coastal Cordillera regions. The effects of the DFS are therefore interpreted to locally overprint the widespread regional rotation, suggesting that easterly bounding fault of the deforming (rotating) domain, lies inboard of the DFS and is probably covered by the thick volcanic products of the more recent magmatic arcs.

7.5 Potential Driving Mechanism for Crustal Rotation in the Central Andes

The preceding discussion has outlined that the majority of widespread (regional) clockwise rotations identified within the present day forearc of northern Chile, occurred post 70-60Ma, but prior to the intrusion of Pluton El Gato (c.42Ma) and certainly prior to the Oligocene-Miocene, as constrained by the non-rotation of sediments and ignimbrites of this age [Arriagada et al., 2003; Somoza et al., 1999; Somoza & Tomlinson, 2002]. This time period coincides with a period of pronounced oblique plate convergence between the Nazca and South American plates (Figures 7.5.1 & 2) [Pardo-Casas & Molnar, 1987; Pilger, 1983; Somoza, 1998]. It should be noted that ideally the angle of convergence should be calculated with reference to the orientation of the paleo-trench (Figure 7.5-2a), but this is impossible to estimate given the large magnitude of the crustal rotations identified, which may have reoriented the trench over time.

The greatest obliquity is observed during the Paleocene (between 65-45Ma) and coincides with a gradual increase in the observed convergence rate (Figure 7.5-
Figure 7.5-1  Positions of two points on the Nazca plate, plotted with respect to South America, illustrating Nazca (Farallon)-South America convergence [redrawn from Figure 3, Pardo-Casas & Molnar, 1987].

Figure 7.5-2  A) Convergence angle between two plates is defined relative to north. To define the angle of obliquity, one really requires the orientation of the margin which in this case is not possible given the large magnitudes of crustal rotation discussed. B) Plot of the angle of convergence with time. Note the highest obliquity of convergence to be between 59 & 42Ma [Pardo-Casas & Molnar, 1987]. Figure 12 of Taylor et al., (2007).
2b). This is interpreted to represent the most likely driving mechanism of crustal rotation and corresponds closely with the proposed timing of rotation in the Coastal Cordillera, Central Valley and Precordillera zones of northern Chile.

The increased obliquity of convergence might be expected to produce a situation where the arc-normal component of shortening in the forearc was accommodated through the regional rotation of large crustal blocks, defined to the north and south by the reactivation of pre-existing crustal anisotropies, rather than through the formation of a classic fold and thrust belt. It is noted however that the easterly extension of the Chañarcillo Fold & Thrust Belt identified by Iriarte et al., (1999) may represent the effects of arc-normal contraction. As discussed previously, the proposed block boundaries are expressed at the surface as NW orientated areas of diffuse deformation, displaying dominantly sinistral offsets.

Crustal rotation is considered to have begun at some time post-60Ma, probably coincident with the pronounced period of caldera formation in the Precordillera. The widespread remagnetisation of the late Cretaceous-early Paleocene strata in the Precordillera, is suggested to be related to fluid expulsion resulting from Paleocene-Eocene volcanism and therefore probably not as a uniform event affecting the entire area. This would explain the observation of both normal and reverse polarity remagnetisation directions throughout the Precordillera.

By ~40Ma convergence between the Nazca (Farallon) and South American plates is near orthogonal, corresponding with the onset of the Incaic Orogeny (Figure 7.5-2b). This period is marked by a change in upper-plate deformation from regional rotation accommodated by diffuse deformation along predominantly widely spaced NW faults, producing regionally consistent clockwise rotations, to the development of discrete fold and thrust belts, with transpression becoming partitioned and localised within the Domyeko Fault System driving localised small block rotations.
The effect of displacement along the La Ternera Fault System (part of the overall DFS) is therefore to locally overprint the regional rotation pattern, with both clockwise and anticlockwise rotations observed.

7.6-Origin of the Central Andean Rotation Pattern

In a seminal paper on Andean tectonics, Isacks (1988) linked along-strike variations in the magnitude of crustal shortening during the Neogene, affecting Bolivia and north western Argentina, to the wholesale 5-10° clockwise (10-15° anticlockwise) rotation of the Chilean (Peruvian) forearc. This rotation model (essentially reworking the oroclinal bending model of Carey (1955), to make it more applicable to the observed deformation pattern and geological history of the Andes-Chapter Three), whilst correctly predicting the sense of rotation to the north and south of the Arica deflection, obviously fails to explain the presence of an anomalous domain of large magnitude clockwise rotations in the present day forearc between 23-29°S, as well as substantially underestimating the magnitude of crustal rotation recorded elsewhere in the margin.

The "Isacks" model infers that a pre-existing bend was modified during a period of rapid convergence between the Nazca and South American plates. Whilst such a modification of the margin in itself may be expected, the suggested differential shortening leading to a uniform, orogen-wide clockwise rotation of the southern limb of the 'Bolivian Orocline' during Miocene to recent orogenesis is at odds with the much earlier (pre-40Ma) time period suggested by palaeomagnetic data from the Northern Chilean forearc. However, as discussed in Chapter Three, the general trend of Neogene crustal rotations throughout the Central Andes suggests that a small component of Neogene rotation may actually be recorded across the margin (Figure 7.6.1). The magnitude of this Neogene component is of the order...
Figure 7.6-1  A-General trend of Neogene rotations (red) throughout the Central Andes. Also shown are Mesozoic and Paleogene rotations from the High Andes and Eastern Cordillera. Generally these rotations form a consistent pattern of rotations, indicating uniform rotations of ~15° both to the north and south of the Arica Deflection as Predicted by Kley, (1999). B-Rotation plotted against latitude (colours as in A), with the remaining Mesozoic and Paleogene rotations from the Present Day forearc plotted as open circles illustrating the 'anomalous' rotational domain between 23-29°S. Larger rotations in the High Andes and Eastern Cordillera generally reflect localised variations. Various data sources-see Appendix A for references.
predicted by Isacks (1988), if not of slightly larger magnitude as predicted by Kley (1999) (~15° clockwise to the north and south of Arica), but is reasonably insignificant in comparison to the large magnitude (older) rotations recorded in the forearc of Northern Chile, between 23-29°S. Oroclinal bending through differential shortening is therefore not considered to have accommodated clockwise rotations between these latitudes.

Although Arriagada et al. (2003) suggest that the large variation in the observed magnitude of crustal rotation in the Antofagasta region results from localised deformation producing in situ block rotations, they do not completely discard the idea that localised deformation may in part modify a larger scale regional rotation pattern. Indeed the consistency of rotation observed between Antofagasta and Copiapó, combined with the consistently large magnitude clockwise rotations observed in the Coastal Cordillera-Precordillera region between 26-29°S observed during this study, would indicate rather that the Chilean forearc between 22-29°S has undergone a large homogenous rotation, far greater than experienced by any other region in the southern Central Andes. As such the forearc between these latitudes represents an anomalous block that appears to have rotated in response to a period of rapid oblique convergence between the Nazca (Farallon) and South American plates. For this reason it is felt that descriptions of crustal rotations accommodated through the displacement of small blocks are inappropriate when considering the gross rotation pattern, as it is difficult to envisage such mechanisms producing the regionally consistent pattern of clockwise rotations observed.

7.7 Conclusions

This study has addressed some of the issues relating to the spatial and temporal distribution of crustal rotation in the present day forearc region of northern Chile. Samples for palaeomagnetic analysis have been collected from a range of
lithologies from two distinct tectono-structural zones. The first, comprising an area within the Coastal Cordillera-Precordillera region, is dominated by the intrusion of the Jurassic-late Cretaceous-earliest Paleocene magmatic arcs and was chosen to investigate the spatial of large clockwise rotations in the Chilean segment of the Central Andes. A second region spanning the La Ternera Fault System situated within the Precordillera is dominated by the intrusion of the Eocene magmatic arc and was chosen to investigate the temporal accumulation of crustal rotation in the Chilean Precordillera, as well as to investigate the effects of the La Ternera Fault system.

7.7-1 Spatial extent of rotations

Several rotation models that have been suggested to explain the pattern of crustal rotations the Chilean segment of the Central Andean margin have been reassessed with regard to the temporal and spatial constraints presented in this study in addition to the pre-existing palaeomagnetic database. Whilst it is appealing to regard the gross pattern as resulting from a single process operating at the scale of the orogen, such large-scale mechanisms (oroclinal bending and differential shortening), fail to explain some of the more intricate details that have been revealed by the relatively dense sampling undertaken during this study. At the other extreme, explanations of the rotation pattern observed in relatively small areas of the Chilean forearc invoking the in situ rotation of small scale blocks (i.e. within strike-slip bounded shear zones), only describe the effects of localised deformation, superimposed on a homogenous regional component of rotation. As such, none of the major fault systems within the forearc region are observed to control the magnitude of crustal rotation.

This study favours the rotation of large-scale crustal blocks, with dimensions approaching the same order of magnitude of the deforming region itself, with the
magnitude of crustal rotation within these blocks observed to be strikingly homogenous. The eastern boundary of crustal rotation is not directly observed, and is interpreted lie beneath the products of the Miocene-recent volcanic arc in the high cordillera to the east. Block boundaries to the north and south are defined by rapid changes in the magnitude of crustal rotation over a short distance (<<block dimensions), suggesting that rotation was accommodated through discrete deformation, although a stepped decline in rotation observed between 26-30°S probably indicates that a small element of differential shortening is also involved. Any component of differential shortening is likely to have resulted from the pre-existing curvature of the overall margin.

Whilst palaeomagnetic data suggests that the boundaries between blocks are relatively sharp, structurally these boundaries appear to be defined by zones of very diffuse deformation, rather than a single discrete structure (c.f. Plutons Las Campañas and Corredores). Between 26-30°S the block boundaries are observed to coincide with the location of NW-orientated transcrustal lineaments that are interpreted to represent pre-existing and fundamental crustal anisotropies and that have been repeatedly reactivated to displace the cover rocks. Such structures have controlled basin formation in the sub-Andes and have interacted with the margin parallel fault systems in the forearc to produce zones of intense mineralisation. It is therefore considered appropriate to invoke such architecturally fundamental lineaments to explain the observed rotation pattern in Northern Chile.

7.7.2 Temporal accumulation of rotation

A significant component of this study has been to detail a more precise determination of the timing of crustal rotation in the northern Chilean forearc. A common assumption has persisted that crustal rotation is directly linked to orogenic processes and therefore that the observed pattern of crustal rotation
developed during the Miocene-Oligocene uplift of the Altiplano-Puna plateau. Crustal rotation in the La Guardia area appears to have been more or less completed before the onset of mountain building and therefore a new driving mechanism is required.

An earliest Paleocene period of rotation is inferred to account for a significant proportion of the total magnitude of observed rotation, corresponding to time of rapid and oblique convergence at the Andean margin. Crustal rotations are interpreted to have accommodated shortening in the present day forearc domain, prior to the onset of more orthogonal convergence associated with the main period of orogenesis.

7.8 Recommendations for Future Work

The reference poles of Besse & Courtillot (2002, 2003) are demonstrated to provide a robust, but above all consistent and well-constrained set of reference directions from latest Triassic age to the present-day. Whilst there is little difference noted between these and pre-existing South American reference poles (Chapter Four), the continuous nature and narrower time constraints employed by Besse and Courtillot (2002, 2003), represent a great improvement in calculating the overall CARP. As such, palaeomagnetic studies in South America are no longer hindered by a lack of quality reference directions and consequently palaeomagnetic data has been used to constrain large-scale tectonic processes at the scale of the entire Andean margin, resulting in the observation of several spatial and temporal variations in the overall CARP. Little further work concerning South American apparent polar wander during the past 200Ma (spanning the Andean Orogenic cycle) is therefore deemed necessary.
Crustal rotations in the Coastal Cordillera-Precordillera region of northern Chile have been demonstrated to define large 'domains' of homogenous (CW) rotation, the magnitude of which decreases southwards between at least 26-30°S. The boundaries between these blocks appear to coincide with either previously defined, or recently observed zones of diffuse deformation associated with (NW-trending) trans-crustal lineaments [as defined by Salfity, 1985]. These lineaments are interpreted to be associated with the pre-existing architecture of the Central Andes and have been variously shown to control basin formation [e.g. Jacques 2003a & b] and economically significant zones of mineralisation [e.g. Bassi, 1988; Chemicoff et al., 2002], and are therefore considered to have played an important role in during Andean orogeny.

The fact that there is little evidence of significant latitudinal movement of the Coastal Cordillera-Precordillera, suggests that the N-S pattern of crustal rotation observed at these latitudes is the result of the in situ (CW) rotation. Although individual 'boundary' structures are not observed directly in the field, crustal rotation is observed to change rapidly across these zones of deformation, suggesting that rotation was accommodated through discrete deformation. A test of this hypothesis would be to establish the pattern of rotation within/across one of these block boundary zones, such as that identified between plutons Las Campaññas and Corredores, in addition to a detailed structural investigation.

With regard to determining the spatial pattern of rotation in the Central Andes of northern Chile, ideally a stratigraphically continuous (and homogenously magnetised) horizon, resulting from a relatively short-lived event and whose age pre-dates rotation, would be sampled across a large area. Whilst such studies have been undertaken on widespread ignimbrite flows, such as the Sifon
ignimbrite [Arribagada et al., 2003; Somoza et al., 1999; Somoza & Tomlinson, 2002], these are young units that post-date rotation.

In the absence of a more ancient equivalent, the latitudinally continuous magmatic arcs have been shown to represent effective markers of crustal rotation, recording an identical rotation pattern to that observed from strata corrected to the palaeohorizontal. The latest Cretaceous-earliest Paleocene magmatic arc appears to be the last magmatic episode to pre-date rotation and illustrates the rotation pattern between 26-30°S particularly clearly. A concentrated program of sampling the latest Cretaceous-earliest Paleocene batholith along its entire length may therefore provide an effective method to assess the variation of crustal rotation along the entire margin (assuming of course that a primary TRM is recorded in all cases). Likewise, any gradient in the magnitude of rotation that may exist across the margin could be observed by systematic sampling of the different magmatic arcs from the Jurassic plutons in the Coastal Cordillera to the latest Cretaceous-earliest Paleocene aged plutons in the Precordillera. The age of magnetisation could further be tested along the margin by sampling the Eocene magmatic arc to investigate whether the timing of crustal rotation is the same in all areas. Generally speaking, palaeomagnetic data from the Copiapo and Antofagasta regions suggests that the timing of (regional) rotation is of Paleocene-Eocene in age, predating Incaic deformation.

A further area of interest that cannot be fully resolved concerns the eastern limit of rotation. Whilst young crustal rotations, associated with Oligocene deformation, are identified in the actively/more recently deformed Eastern Cordillera, these are small and much more localised in nature when compared to crustal rotations in the present day forearc. The Atacama Fault System and Chañarcillo Fold & Thrust Belt have been shown not to control rotation, although the deformation associated
with the latter may be temporally associated with rotation. Although the La Ternera Fault System (associated with Incaic deformation) is shown to at least overprint the regional rotation pattern, the lack of palaeomagnetic data to the east (largely due to the difficult nature of the terrain), means that it is difficult to establish the full easterly extent of crustal rotation. The eastern boundary is interpreted to lie to the east of the Precordillera region, and is probably covered by the voluminous volcanic products of the Miocene-recent arc. As such the eastern may not be exposed but this needs further investigation.
Appendix A

Palaeomagnetic Archive of Data from this Study

The raw palaeomagnetic data and magnetic mineralogy experiment files are stored on a CD archive disc held at the University of Plymouth Palaeomagnetic Laboratory. Where free to distribute, the software used for the analysis and interpretation of the data, are also included. To access this data please contact either Dr Graeme Taylor or Dr Antony Morris at the School of Earth, Ocean and Environmental Sciences at the University of Plymouth.
Appendix B

Central Andean Rotation Pattern (CARP) Database

All of the palaeomagnetic data used during the discussion of the 'Central Andean Rotation Pattern' of CARP, including that from this study and recalculated during this study, are reproduced in this Appendix as a series of summary tables. Listed are the sampling locality (comprising the Formation/Pluton name and location), overall mean direction, age of magnetisation suggested by the original author (or reinterpreted during this study where appropriate) and the magnitude of crustal rotation calculated using the 10Ma reference directions calculated by Besse & Courtillot, (2002, 2003). The original reference is quoted in all cases unless otherwise stated.

The data is primarily divided by the tectonomorphic location of a particular study, being separated into the Coastal Cordillera, Central Valley, Precordillera (Western Cordillera), Altiplano-Puna Plateau, Eastern Cordillera and Sub-Andean zones. The data is further divided into Mesozoic (green arrows), Paleogene (blue) and Neogene (red) age brackets, with individual rotations plotted on map diagrams (A) and line/scatter graphs (B) accompanying the data tables. Plotted on all of the graphs are the best-fitting polynomial lines to rotations along the length of the Andes for each of the age brackets, using an identical colour scheme as for the map diagrams.
Table A.1-1  
**Neogene-Coastal Cordillera**

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Figure A.1-1  Mesozoic crustal rotations in the Coastal Cordillera
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Figure B.1-1  Paleogene crustal rotations in the Central Valley

A.

B.
Figure B.1-2  Mesozoic crustal rotations in the Central Valley
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Figure C.1-2  Paleogene crustal rotations in the Precordillera
Figure C.1-3  Mesozoic crustal rotations in the Precordillera
### Table D.1.1

#### Neogene-Altiplano-Puna

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Figure D.1-1  Neogene & Paleogene crustal rotations in the Altiplano-Puna Plateau
Table E.1-1

**Neogene-Eastern Cordillera**

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Table E.1-2

**Paleogene-Eastern Cordillera**

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Table E.1-3

**Mesozoic-Eastern Cordillera**

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Figure E.1-1  Neogene, Paleogene and Mesozoic crustal rotations in the Eastern Cordillera

A.

B.
### Table F.1-1

**Neogene-Sub-Andes**

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<th>Locality</th>
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<th>Long. (°E)</th>
<th>Ref. (Ma)</th>
<th>Dec. (°)</th>
<th>Inc. (°)</th>
<th>α&lt;sub&gt;ref&lt;/sub&gt; Rotation (°)</th>
<th>Flattening (°)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>South of Arica Deflection (&gt;19°S)</strong></td>
<td></td>
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<tr>
<td>Pirgua Group</td>
<td>25.80</td>
<td>294.30</td>
<td>100</td>
<td>3.6</td>
<td>-45.4</td>
<td>6.4</td>
<td><strong>6.9 ± 9.5</strong></td>
<td>-0.4 ± 7.6</td>
</tr>
<tr>
<td>Aubrey et al. (1996)</td>
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Figure F.1-1
Neogene and Mesozoic crustal rotations in the Sub-Andes
Appendix C

Publications & Conference Contributions

Four papers have been peer-reviewed, accepted and published, three of which are directly related to actual data from this study. A fourth concerns the acquisition of resistivity data in support of a micro-palaeontological study. References for these papers are presented below, in addition to the details of conference abstracts submitted during the course of this study.

Publications


Conference Contributions


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