**Magnetic polarity stratigraphy and geochronological constraints on
the sequence of thrusting in the Principal and Frontal cordilleras and
the Precordillera of the Argentine central Andes (33°S latitude)**

by
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To my teachers of all times

Frontispiece: Photograph looking westward to the Mariño Formation and in the distance the Cordón del Plata of the Frontal Cordillera, Mendoza Province, Argentina. The Mariño Formation is the oldest unit of the Neogene foreland basin deposits that accumulated in response to deformation and exhumation in the Principal and Frontal cordilleras. The photo is taken from a dry riverbed, Río Seco de Cacheuta; the road in the foreground leads south from the international route (connecting Chile and Argentina) at a point between Cacheuta and Potrerillos.

ABSTRACT

This research focuses on the reconstruction of the Cenozoic thrust history of the Andes Mountains at 33°S by analyzing the foreland basin strata that record the eastern migration of the Andean thrust front since the Neogene. About 4000 m of non-marine deposits crop out in the eastern flank of the Andes at the southernmost portion of the Mendoza Precordillera and accumulated in response to exhumation and deformation of the western Principal and Frontal cordilleras. The detrital clastic strata comprise five Miocene and Pliocene formational units that record fluvial, eolian and alluvial deposition. To link the sequence of deformational events in the western mountain belts with the sedimentary record, all units except the youngest one were dated using magnetic polarity stratigraphy calibrated with ^{40}Ar - ^{39}Ar isotopic dates of interbedded tephtras. Precise depositional ages in conjunction with facies, textural patterns, and sandstone petrography allowed determination of rates of sedimentation in the foreland, variations in depositional environment through time and timing of changes of detritus provenance. The multiple data set reveals details of tectonic activity, volcanism and deposition, which are summarized as follows.

The oldest unit, the Mariño Formation (~15.7-12.2 Ma), records two phases of thrusting in the Principal Cordillera. The earlier phase occurred prior to 15.7 Ma and is tied to deformation and exhumation of the Mesozoic andesitic volcanic complex cropping out in the western part of the Aconcagua fold-and-thrust belt (AFTB). The Purple conglomerates at the base of the unit could record the phase of deformation in the AFTB whose onset is known to have occurred at ~20 Ma. The initiation of the second phase of thrusting is interpreted to have occurred at ~15.5 Ma and coincides with the accumulation of the eolian middle member, interpreted as a major retrograding facies developed as a result of thrust-induced subsidence. The first appearance at ~14.7 Ma of fossiliferous limestone clasts in deposits at the base of the fluvial upper member documents the eastward migration of the deformation front and involvement of the Mesozoic marine sequences of the central part of the AFTB.

The overlying La Pilona Formation (~11.7-9.0 Ma) is thought to record the initiation of exhumation of the Frontal Cordillera just prior to ~10.9 Ma and an important phase of deformation, highlighted by a marked increase in subsidence rates, at ~9.7 Ma. Provenance and paleocurrent data are consistent with clast derivation from northwestern highland sources which provided abundant rhyolite clasts and low-grade metamorphic and sedimentary rocks, a petrologic assemblage that typifies this morphostructural province. Deposition of the Tobas Angostura unit (8.9-8.7 Ma) primarily reflects the influence of large volcanism-induced sediment load rather than tectonism. This unit is correlative with the largest and latest extrusive episodes occurring in the Aconcagua region, which produced large volumes of volcanic debris that overwhelmed the basin. The overlying Río de los Pozos Formation, at least as young as ~7 Ma, records the reestablishment of normal stream-flow processes where drainage of hill slopes with different lithologies from the Frontal and Principal cordilleras gradually regained precedence.

Exhumation of the Frontal Cordillera overlapped in time with thrusting in the eastern part of the AFTB, in which the latest phase of deformation is known to have occurred at ~8 Ma. Tectonic activity in the Frontal Cordillera shifted through time to the south along the eastern slope of the Cordón del Plata, as recorded by the proximal synorogenic deposits of the Mogotes Formation, which unconformably overlies the Río de los Pozos Formation. Deposition of this unit at the study site probably started at ~3 Ma and ceased prior to 0.4 Ma, according to temporal constraints provided by ash chronology on Quaternary deposits, and most probably before or at ~1 Ma. The thick Neogene succession of the Mendoza Precordillera was deformed into a series of north-trending anticlines and synclines in the Late Pliocene - Early Pleistocene?. The present location of the thrust front near the city of Mendoza indicates that deformation has progressed eastward in the last ~3 Ma. The eastern border of the Precordillera is still active as inferred from earthquake focal mechanisms, intense compressive deformation and evidence of neotectonic activity in the study area and farther east in the foothills near the city of Mendoza.

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ORIGINAL CONTRIBUTION

Before the beginning of this research project at Carleton University, in March of 1991, I spent one week of reconnaissance in the field collecting oriented rock samples for testing suitability for paleomagnetic procedures. I processed the collected samples at the Paleomagnetism Laboratory of Buenos Aires University. The project itself required two field seasons totalling approximately 6 months of work. The first took place between September and December of 1992 and the second from April to June of 1994. Since most of the geological contacts were already mapped by previous fieldworkers, my field work involved some mapping to corroborate or supplement already existing mapping in the area, making regional observations, constructing sedimentological profiles, describing the Tertiary units and collecting numerous rock samples for three purposes: (a) paleomagnetic processing, (b) ^{40}Ar - ^{39}Ar dating and (c) provenance analysis.

To establish the magnetic polarity stratigraphy, I performed time-consuming cleaning experiments, mainly stepwise thermal demagnetization procedures, at the Paleomagnetism Laboratory of the Geological Survey of Canada. I performed all stages of mineral separation at Carleton University on the samples collected for radiometric dating. This involved examination of thin sections, crushing, grinding, and preparation of mineral separates, which required handpicking procedures in the last stages. The chemistry and mass spectrometry were performed by technicians at the Geological Survey of Canada in Ottawa and at the MIT Laboratory. I examined about 45 thin sections of sandstone samples for provenance study, 34 of which I point-counted (in most cases a total of 400 points per slide) using standard point-count techniques. In the preparation of thin sections, I was assisted by a technician in the procedure for staining feldspars. The thin sections, which required impregnation techniques, were done by a technician at Carleton University. X-ray mineral determinations were, when required, provided by a technician at Carleton University.

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CHAPTER 1

GENERAL INTRODUCTION

1. FOREWORD

(a) Purpose of investigation

In the late 1950's, Jorge Polanski was commissioned by the Dirección Nacional de Geología y Minería to initiate extensive mapping in the Mendoza Province of west-central Argentina. His work has provided an overview of the geology and structure of the Andes between about 33° and 34°S latitude (Polanski, 1958, 1963, 1964 a and b, 1972). In the framework of geosynclinal theory, he envisaged the present-day configuration of the Andes as the result of pulses of uplift and deformation of tectonically active belts where the locus of deformation has migrated from west to east since Neogene times. Although some of his explanations are no longer valid in light of plate tectonic theory, he provided the basis for understanding the Cenozoic tectonic evolution of the Andes at this latitude.

Spurred by Polanski's ideas but within the basic framework provided by plate tectonics, Ramos (1988) proposed a modern model for the Neogene evolution of the Andes. He also linked Cenozoic synorogenic deposits that crop out in the eastern flank of the Andes Mountains with eastward migration of the orogenic thrust front since late Tertiary time. This hypothesis, however, needed to be tested since at that time there were no geochronological

data on which to base a chronology of thrusting.

In 1991, Ramos encouraged this writer to investigate the timing of thrusting of the Andes Mountains at about 33°S latitude by analyzing the foreland basin strata located in the southernmost extension of the Mendoza Precordillera belt. An indirect method of dating thrust fault activity was required since the lack of cross-cutting relationships and the scarcity of paleontologic remains would otherwise preclude analysis. Because previous studies utilizing the techniques of magnetic stratigraphy in other areas of Argentina (e.g., Johnsson et al., 1984; Johnson et al., 1986; Jordan et al., 1988, 1990; Reynolds et al., 1990, among others) have proved to be successful in deciphering the times of motion on thrust faults from the sedimentary record, a similar approach might be attempted with the Mendoza strata.

Consequently, this research focuses on the reconstruction of the Cenozoic thrust history at 33°S by studying the synorogenic strata which are the 'tape recorders' of the thrust belt deformation. A brief overview of the geology and tectonics of the study area is provided in this chapter.

As the interpretation of the timing of thrusting requires a precise chronology of the foreland basin strata, the Tertiary nonmarine deposits have been dated using magnetic polarity stratigraphy calibrated with ^{40}Ar - ^{39}Ar isotopic dates. Chapter 2 reports on the age determinations obtained by ^{40}Ar - ^{39}Ar isotopic dating of interbedded tuff beds and includes a brief overview of this method. Chapter 3 describes the paleomagnetic study carried out on the Neogene deposits and the age assignment for the individual magnetic polarity sections based on the correlation with the standard reversal chronology.

Chapter 4 focuses on the physical description of the Tertiary nonmarine deposits.

These exposures consist of five formational units; all but the youngest formation have been dated using magnetic polarity stratigraphy. The chapter includes a facies description and interpretation of the environments of deposition for each of the study units; the provenance study on these rocks, which permits recognition of the source areas exposed by thrusting; and the determinations of the accumulation rates for each of the study sections. The chapter concludes with the integration of a multiple data set where sedimentation rates, facies variations and timing of changes of detritus provenance in the foreland are linked to episodes of thrust activity in the hinterland.

Chapter 5 presents a tectonic synthesis of the kinematic history of thrusting at 33°S, its duration, and the relationship between sedimentation and tectonic activity during Neogene times.

(b) Geographic location

The study area is located along the eastern flank of the Andes at the southernmost portion of the Precordillera where the thrust-and-fold belt plunges beneath the Tertiary strata at 33°S. Most of the study was carried out south of the Cacheuta peak, where Neogene strata are exposed in a south-plunging anticline known as La Pilona. These exposures were also studied in the eastern flank of the Tupungato anticline, located approximately 25 km to the south (fig. 1.1). The international route that connects Argentina and Chile and, locally, the towns of Cacheuta and Potrerillos, south of the Mendoza River, provides access to the region. Dirt roads allow access to the study sections including the outcrops of the Tupungato

Fig. 1-1. Map and location of the study area and detail of La Pilona and Tupungato anticlines.

area to the south. The towns of San José and Tupungato lie in the southern part of the study region.

(c) Previous work

Geological studies have been ongoing in the area since the 1930's as part of an effort by the Argentine national oil company (Y.P.F) to evaluate hydrocarbon potential. An important contribution was made by Truempy and Lhez (1937), who provided the basis for establishing the stratigraphy of the Tertiary strata in the Cacheuta and Tupungato region. Unpublished reports produced by Y.P.F. shortly thereafter (e.g., Dessanti, 1943; Tognon and Pesciulezi, 1947; Tognon and Pérez, 1949; Rolleri, 1950; Yrigoyen et al., 1954; De La Motta, 1957, among others) were used for consultation in the central offices of the company. New surveys were produced by Y.P.F. in the 1980's where the northwestern portion of the study area was mapped at 1:50000 (Kokogian et al., 1987). Geological maps at the scale of 1:200000 (Rolleri, 1950; Polanski, 1964a, 1972) and 1:500000 (Introcaso and Ramos, 1989) provided the regional setting for this research. Maps and information produced by Amos and Rolleri (1965), Caminos (1965, 1979), Rolleri and Fernández Garrasino (1979), Ramos and Yrigoyen (1987), Ramos (1988, 1990, 1993, 1996 a and b), Ramos et al. (1996a), Ramos and Nullo (1993), Yrigoyen (1979, 1993 a and b), and Cortés (1993) were also used for consultation. Subsurface data depicting the structure of the region are published in simplified form (e.g., Sarewitz, 1988; Kozlowski et al., 1993), but most of the reflection profiles remain proprietary.

2. GEOLOGICAL SETTING

The study area is situated above a dip transition in the orthogonally subducting Nazca plate (Isacks and Barazangi, 1977). To the north, the underthrusting plate is nearly horizontal, and to the south, it assumes a dip of approximately 30° ; this transition is reflected in the structure and morphological expression of the thrust belt (Jordan et al., 1983).

Andean tectonics above the flat subducting segment (28° - 33° S latitude, figs. 1.2) depicts a steady topographic rise from the Chilean coast in the west to the peaks of the Andes in the east; a narrow belt of eastward migrating, apparently thin-skinned Neogene to Quaternary shortening in the eastern slope of the Andes; a Plio-Pleistocene uplift of the crystalline basement of the Pampeanas Range (Sierras Pampeanas); and no significant Quaternary volcanism (Jordan et al. 1983). In contrast, south of $33^\circ 15'$, the most conspicuous feature is the development of Quaternary arc and extensive back-arc volcanism, evidence of the correlation between the dip of the subducted slab and the tectonic segmentation of the upper plate (Ramos and Nullo, 1993).

(a) Major geological provinces

The central Andes (28° - 33° S) has been subdivided into a series of north-trending morphotectonic units or geological provinces based on structural styles, geologic evolution, and morphological expression. These units are displayed in figure 1.2. The regional geology

Fig. 1.2. Major morphostructural units of central Andes between 28° and 35° S latitude (after Ramos, 1988). RV= retroarc volcanic province. Dotted lines indicate the approximate limit of the flat-slab subduction segment.

of the Principal and Frontal cordilleras and the Precordillera, of particular interest to this study, is shown in figure 1.3.

Coastal Cordillera

Along the present continental margin, Devonian to Lower Jurassic strata are exposed and are interpreted to be part of an exotic terrane that accreted during the Early Jurassic (Mpodozis and Forsythe, 1983; Hervé, 1988). The Pichidanguí Terrane, which includes metamorphic rocks, sedimentary sequences and volcanic rocks, was emplaced by margin-parallel strike-slip motion in the Late Triassic - Early Jurassic (Forsythe et al., 1986). Strike-slip motion along the Chilean coast was accompanied by the development of Triassic rift basins in Argentina (Ramos and Kay, 1991). The basement is covered by a thick, autochthonous Mesozoic volcano-sedimentary sequence consisting mainly of Jurassic rhyolitic and pyroclastic rocks interbedded with limestones and of Lower Cretaceous andesitic lava flows. These rocks are intruded by Jurassic and Cretaceous batholiths (Mpodozis and Ramos, 1990). During the Late Cretaceous, compression seems to have resulted in the beginning of eastward thrusting of the Cretaceous volcanic pile (Mpodozis and Ramos, 1990).

Principal Cordillera

The Principal Cordillera, also called the Main Andes, was the locus of the Andean orogeny during latest Mesozoic to Cenozoic times. It is composed primarily of a thick sequence of Jurassic and Cretaceous marine deposits which unconformably overlie marine Carboniferous strata and the Choiyoi volcanics of the Frontal Cordillera (Yrigoyen, 1979).

Fig. 1.3. Map of the regional geology of the Principal and Frontal cordilleras and the Precordillera (modified from Introcaso and Ramos, 1989).

These marine deposits display thick-skinned deformation in the northern portion of the belt (Mpodozis and Forsythe, 1983), while to the south thin-skinned structures are recognized, such as the Aconcagua fold-and-thrust belt (Ramos, 1988). Along the main continental drainage divide these sediments interfinger with a thick pile of Upper Jurassic - Lower Cretaceous volcanic and pyroclastic rocks. These volcanic sequences occur along a western inner arc developed between the Coastal and Principal cordilleras and an eastern arc located along the present international border (Ramos, 1990). A series of volcanic arcs shifted from the Coastal Cordillera in the Jurassic to the Principal Cordillera in the late Tertiary (Ramos, 1988). A thick sequence of conglomerate and sandstone interpreted as Tertiary alluvial-fan deposits (González Bonorino, 1950) is interbedded with these volcanics.

Frontal Cordillera

The Frontal Cordillera comprises mainly andesitic to silicic magmatic rocks of the Choiyoi Group formed in the late Paleozoic to early Mesozoic. Andesite breccias and rhyolitic tuffs intruded by gabbro dykes and granitic stocks are widespread in the eastern slope of the Cordón del Plata (Camino, 1979). These rocks are interpreted to be the result of an Andean type of subduction orogenesis followed by a generalized extension (Ramos, 1990). Exposures of the oldest units are represented by a Precambrian to lower Paleozoic crystalline basement formed of low- to medium-grade metamorphic rocks (Camino, 1979). Upon this basement rests a thick pile of two Upper Carboniferous units representing littoral and deep sea strata, the former cropping out along the crest of the Cordón del Plata and the latter on its eastern slope. Limited exposures of Tertiary synorogenic deposits are found along a narrow north-

south trending zone in the easternmost part of the belt. During the Andean orogeny the Frontal Cordillera apparently behaved as a rigid block, disrupted by high-angle reverse faults but probably not recording significant internal shortening (Mpodozis and Ramos, 1990). The eastern slope of the Cordón del Plata is characterized by an imbricate structure (Las Carreras fault system) developed in the latest Pliocene to early Pleistocene? (Camino, 1979).

Precordillera

The Precordillera, which defines the eastern flank of the Andes, is a predominantly east-verging fold-and-thrust belt with a thin-skinned structure that has been developed on early to late Paleozoic sedimentary rocks (Baldi and Chebli, 1969). The western edge of the Precordillera coincides with a longitudinal depression known as the Calingasta-Uspallata valley. The present morphology of this tectonic trough is controlled by the old Paleozoic continental margin (Ramos, 1990). South of 33°S, the thrust belt plunges beneath Tertiary cover rocks, and the Precordillera dies out as a physiographic unit. The Paleozoic stratigraphy is represented by a Lower Cambrian to Ordovician carbonate platform covered by lower Paleozoic clastic marine facies in the central and eastern Precordillera and oceanic facies in the western Precordillera (Baldi et al., 1982). These conditions persisted until the Late Devonian when the collision and accretion of an allochthonous terrane changed the paleogeography (Ramos et al. 1984). Mesozoic deposits overlie late Paleozoic strata and are mainly preserved in basin structures along the western and eastern margins. Tertiary synorogenic deposits with volcanic intercalations are also found along both margins of the belt (Leveratto, 1968; Rolleri and Fernández Garrasino, 1979). These strata allowed

reconstruction of the chronology of thrusting, which began at about 20 Ma in the western Precordillera and has continued to the present in the eastern part of the belt (Jordan et al., 1993 a and b).

Sierras Pampeanas

This geological province is characterized by crystalline basement rocks of Precambrian - early Paleozoic age that form mountain blocks that have been uplifted on north-northwest-trending reverse faults (Gordillo and Lencinas, 1979; Caminos, 1979). Vertical throws of more than 6 km separate the mountain blocks from the adjacent broad basins. These faults place the metamorphic and igneous rocks over upper Cenozoic strata, with associated drag folds (Gordillo and Lencinas, 1979). Although poorly constrained, the main deformation is thought to be Pliocene and Pleistocene in age. Continental red beds of late Paleozoic age and Tertiary synorogenic deposits in alluvial and fluvial facies cover this basement. The resulting structure is thought to be related to Tertiary Andean compression (González Bonorino, 1950) in association with a period of shallow subduction (Jordan et al. 1983). The style of deformation of the Pampeanas Ranges has been compared with the Laramide region of the United States (Jordan and Allmendinger, 1986).

(b) Andean tectonics

After South America became completely separated from Africa, less than 100 million years ago, the tectonic regime changed from being extensional, with large manifestations of

volcanism, to compressive, thus initiating the deformation and the build-up of the Andes (Ramos, 1989). This change in the tectonic regime gave rise to the beginning of the Andean Orogenic Cycle, with the development of a magmatic arc (Coastal Cordillera), a trans-arc extensional basin and a sedimentary platform, marking the transition to the foreland (Introcaso et al., 1992). It is not until the early Tertiary that there is an important eastward migration of the orogenic thrust front recorded by the Paleogene synorogenic deposits outcropping in the southwest of the Mendoza province (Ramos, 1993). However, the development of a thick synorogenic wedge with the typical geometry of a foreland basin is well documented only for the Miocene and later; several depocenters resulting from the tectonically induced subsidence are recognized (Ramos, 1993).

An eastward migration of the orogenic thrust front is recorded by these Neogene strata. A thick pile of continental deposits (Santa María Conglomerates of Schiller, 1912; fig. 1.3), interpreted as talus and alluvial-fan facies, rests unconformably upon the Mesozoic rocks in the Principal Cordillera. This well developed regional unconformity is assigned to the Incaic phase of the late Eocene - earliest Oligocene (Ramos, 1985). A minimum age of 8.6 Ma is reported for the uppermost section of the Santa María Conglomerates (Ramos et al., 1990). These sediments are interfingered with the lower section of the volcanics of the Complejo Volcánico Aconcagua, whose age ranges from 15.8 to 8.9 Ma in the Argentine Principal Cordillera (Ramos et al., 1991). In the adjacent Chilean sector the age of these volcanics is older (Farellones Formation), ranging from 20.4 to 16.6 Ma, which indicates that there was a migration of the magmatic arc towards the east during the Lower Miocene (Ramos et al., 1991).

Farther east of the Main Cordillera, in the Precordillera, the Incaic phase unconformity separates the nonmarine Eocene deposits of the Divisadero Largo Formation from younger Tertiary strata. The latter, which outcrop as remnants in the Frontal Cordillera (in the southern part of the Uspallata valley, fig. 1.3) and are well preserved in the Precordillera, represent distal alluvial facies partially synchronous with the Santa María Conglomerates and the volcanic rocks of the Complejo Volcánico Aconcagua. They contain several tuff layers that attest to the volcanic activity at that time (Ramos, 1988).

In the Cacheuta region, an unconformity dated at about 12 Ma (cf. Chapters 2 and 3) separates La Piona Formation, presently interpreted as Middle to Late Miocene in age, from older Tertiary beds. The development of this unconformity, assigned to the Quechua phase, has been correlated with an episode of deformation in the Principal Cordillera, responsible for the folding and thrusting of the Santa María Conglomerates and the volcanics of the Complejo Volcánico Aconcagua (cf., Ramos, 1988). The study of synorogenic strata in the Cacheuta region (this research) suggests that the onset of deformation of the Frontal Cordillera also occurred at that time, in agreement with Polanski (1964a).

It has been proposed that the eastern border of the Frontal Cordillera was uplifted during the Diaguitic phase (Polanski, 1964 a and b; Yrigoyen, 1979). The basal unconformity of the coarse conglomerates of the Mogotes Formation developed as a result of tectonic uplift in the flanking thrust belt. Although this phase of deformation is rather imprecisely dated and most probably comprised of several tectonic pulses, it has been inferred that it occurred between about 5 and 4 Ma (Yrigoyen, 1993b; Yrigoyen et al., 1995). The whole pile of Neogene strata in the Uspallata and Cacheuta region were folded and thrust in the latest

Pliocene and possibly in the early Pleistocene (Irigoyen et al., 1995). Evidence of neotectonic activity is found in the study area, in the Cacheuta region and farther east, near the city of Mendoza (Polanski, 1963; Rodriguez and Barton, 1993).

At 33°S latitude, the active thrust front is located in the eastern border of the Precordillera, near the city of Mendoza (Ramos, 1990). Major earthquakes partially destroyed the city in the last century. Evidence of neotectonic activity affecting the different Quaternary units is found along a 60 km long area flanking the Precordillera (Bastías et al., 1993). The thrust front of the Mendoza Precordillera consists of a series of east-vergent thrusts where Tertiary deposits are overridden by Paleozoic strata (Figueroa and Ferraris, 1989). Distinctively, the north-south trend of different thrusts is abruptly modified by an east-west deflection near the city of Mendoza. This deflection which corresponds to a lateral ramp, is interpreted to be a reactivated east-west transfer fault of the Triassic rift system (Ramos, 1990; Kozłowski et al., 1993).

(c) The study area

The oldest rocks in the La Pilona area are Ordovician low-grade metasedimentary strata cropping out in the Cacheuta area (Varela et al. 1993; see fig. 1.4a). They consist of a deformed pile of sandstones interbedded with mudstones. An igneous suite of monzodiorites and monzogabbros of Silurian age intruded these strata (Varela et al. 1993). Permian-Triassic volcanic and plutonic rocks of the Choiyoi Group exposed in Cacheuta peak are mainly andesites, rhyolites and felsic granitic rocks. These volcanics thicken to the west toward the

Fig. 1.4. Geology of the study area showing the locations of the magnetostratigraphic sections. (a) La Pilona anticline (sections I-IV). ^{40}Ar - ^{39}Ar isotopic dates of interbedded tuff horizons come from the following samples: Md-12 from the Mariño Formation in section Ib; Md-13 from the La Pilona Formation in section II; Md-6 and Md-10 from the La Pilona and Tobas Angostura formations respectively, in section III; and Md-5 from the Tobas Angostura unit in section IV. (b) Map showing the location of the Quaternary air-fall horizon (sample Md-9) dated to constrain the time of motion of the La Pilona structure. (c) Refugio and Tupungato anticlines with inset to show the location of section V (eastern flank of the Tupungato anticline, along the Quebrada de la Angostura dry riverbed), from which sample Md-15 in the Tobas Angostura unit was obtained. Samples Md-8 comes from the Quaternary tephra dated to constrain the time of motion of the Refugio and Tupungato anticlines. Geology in parts a and b is based on a compilation of mapping from this study and from unpublished reports (e.g., Roller, 1949 and 1950; De la Motta, 1957; Rebori, 1979), and published reports (e.g., Varela et al., 1993). Discussions with Rebay, G. (pers. comm., 1992) provided understanding of the Tertiary stratigraphy. In part c, regional map is based on unpublished reports from Y.P.F. and the detailed map is from Yrigoyen (1993b).

Frontal Cordillera to a maximum of more than 2000-4000 m (Cortés, 1985) and cover a thick pile of Carboniferous strata. Middle to Upper Triassic nonmarine strata and subordinate volcanic rocks representing graben-fill sediments of the Cuyo basin overlie the volcanics of the Choiyoi Group. These sediments, which range from coarse alluvial fan deposits to laminated lacustrine mudstones (Rolleri and Fernández Garrasino, 1979), are considered to be the sedimentary fill of a continental rift basin (Uliana and Biddle, 1988; Kokogian and Mancilla, 1989; Ramos and Kay, 1991). Hydrocarbon reserves generated in the Mendoza Triassic rocks are hosted in several oil fields, one of which is the La Pilona anticline in the study area. More than 3000 m of deformed Paleogene to Pliocene continental deposits disconformably overlie the Triassic strata (Rolleri and Fernández Garrasino, 1979). Neogene strata are deformed into two north-northwest-striking anticlinal systems, the westernmost of which comprises the La Pilona and Tupungato anticlines of the study area (fig. 1.5). The more easterly Lunlunta and Barrancas anticlines mark the edge of the modern deformation front. The stratigraphy of the Tertiary deposits is summarized in the next section. Quaternary deposits are represented by widely developed pediments and several terraces which contain intercalations of air-fall deposits (Polanski, 1963).

(d) Tertiary deposits in the Cacheuta-Tupungato area

The formational nomenclature used in this research is extracted from Yrigoyen (1993a), who recently made a stratigraphic revision of the Tertiary units in the Mendoza province. Six outcropping formations have been recognized in the study area. The

Fig. 1.5. (a) Photo-mosaic of the study area. (b) The line drawing below superimposed on a scan of the above photo highlights the most distinctive geological features of the southern extension of the Precordillera fold-and-thrust belt. Neogene strata are deformed into two north-striking anticlinal systems, the westernmost of which comprises the La Pilona and Tupungato anticlines. The more easterly Lunlunta and Barrancas anticlines mark the edge of the modern deformation front. The Frontal Cordillera to the west is characterized by an imbricate structure. Triangle symbol shows the Cacheuta peak.

a

b

stratigraphy of the Neogene units, as determined in this study and by previous researchers, is summarized in table 1.1.

Divisadero Largo Formation (Chiotti, 1946)

A detailed lithological description of this unit is provided by Chiotti (1946) and Minoprio (1951). As summarized by Yrigoyen (1993a) this unit comprises two members, a basal section ('Zona de anhidrita'), consisting of sandstones with gypsum- and anhydrite-bearing horizons, and an upper section ('Arcillas abigarradas'), which mostly comprises varicoloured mudstones. This formation reaches a thickness of up to 150 m. Based on the study of vertebrate fossils collected from the upper member of the unit, the Divisadero Largo Formation is assigned to the Upper Eocene (e.g., Pascual and Odreman Rivas, 1971, 1973), Upper Eocene - Lower Oligocene (e.g., Pascual and Bondesio, 1981) or Upper Oligocene - Lower Miocene (e.g., Baez, 1986). Based on regional studies, Legarreta et al. (1993) suggest that the lower member of the unit ('Zona con anhidrita') is Upper Cretaceous - Paleocene and the upper member is Upper Eocene - Lower Oligocene.

Mariño Formation (Rolleri and Criado Roque, 1970)

As summarized by Yrigoyen (1993a), this formation is subdivided into three members. The lower member, 'Conglomerados violáceos' (Purple conglomerates), may attain a thickness of approximately 50 m and consists of purple conglomeratic beds alternating with greenish sandstones. The intermediate member, 'Areniscas entrecruzadas' (Cross-bedded sandstones), is a section comprised of a distinctive 120 to 250 m thick pile of eolian cross-bedded

Table 1.1. Stratigraphy of the Neogene deposits in the Cacheuta-Tupungato area

Time (Ma) ¹		Rolleri and Fernán- dez Garrasino, 1979 ²	Jordan and Ortiz, 1987 ³	Yrigoyen, 1993 a,b ⁴	This study ⁵		
Pleistocene		Quaternary	Quaternary	Pleistocene	Pleistocene		
1.8				----- ? ----- 1.2	----- ? ----- 1.2		
Pliocene	Late	Mogotes	Mogotes	Mogotes	Mogotes		
	3.5	Serie Amarilla	~~~~~ 3.8			~~~~~ 3.5	
	Early	Tobas Grises Superiores				~~~~~	
La Pilona		Tobas Grises Superiores		~~~~~			
	5.3	~~~~~ ?			~~~~~		
Miocene	Late	?	La Pilona	Río de los Pozos (=Serie Amarilla)	~~~~~		
		↑			~~~~~		
		Tobas Grises Inferiores	Tobas Grises Inferiores	?	~~~~~		
		↓		~~~~~	~~~~~		
	Middle	?	Mariño	----- ? ----- 8.4	~~~~~ 8.7		
		?		Tobas Angostura (=Tob.Gr.Sup.)	~~~~~ 9.0		
		~~~~~ ?		~~~~~ 10.5	La Pilona		
		~~~~~ ?		~~~~~ 10.8			
	Early	?		Toba La Higuera (=Tob.Gr.Inf.)	~~~~~ 12.2		
		?		↓	~~~~~		
		?		?	~~~~~		
		~~~~~ ?		~~~~~ 15.5	~~~~~		
Oligocene	Early		Estratos de Mariño	Mariño	Estr. de Mariño		
			~~~~~ 21.0			Areniscas entrecruzadas	
							Conglomerados violáceos
23.8							
Eocene	Early						
				~~~~~ 35.5			

¹ Cenozoic geochronology based on Cande and Kent (1992).

² Compilation of previous information, mainly from Truempy and Lhez (1937), Rolleri and Criado Roque (1970), and Pascual and Odreman Rivas (1973). Formations are assigned to epochs (i.e., Middle Miocene) based on mammal fossils and are not precisely dated.

³ Age assignments derived from stratigraphic correlations with Tertiary formational units from Sierra de Huaco, San Juan Province.

⁴ Age assignments by Yrigoyen (1993a) are based on Haq et al. (1987) time scale, which places the Eocene-Oligocene boundary at 36.0 Ma. Thus, Mariño strata are considered to be Oligocene-Miocene.

⁵ Stratigraphic nomenclature is based on Yrigoyen (1993a). The basal part of the La Pilona Formation includes the deposits that have been previously mapped as Tobas Grises Inferiores (Lower gray tuffs) (e.g., De la Motta, 1957). For explanation, see text.

sandstones with intercalations of mudstones and conglomeratic beds. The upper member, 'Estratos de Mariño' (Mariño strata) consists of a ~1000 m thick pile of sandstones interbedded with red mudstones and gray conglomerates representing fluvial-alluvial facies.

Based on magnetic polarity stratigraphy and ash chronology (see Chapters 2 and 3) the Mariño Formation is assigned to the interval 15.7-12.2 Ma, a middle Miocene age. Based on vertebrate remains collected from strata of the eastern limb of La Pilona anticline, this unit has been previously assigned to the Miocene (Rolleri, 1950; Pascual and Odreman Rivas, 1973) and Early Oligocene - Early Miocene, between 35.5 and 15.5 Ma (Yrigoyen, 1993a).

La Pilona Formation (Truempy and Lhez, 1937; Fossa Mancini, 1938)

This unit, which is up to ~800 m thick in the Cacheuta region, is composed in its lower part of gray sandstones and conglomerates overlain by a tuffaceous finer-grained succession containing ash-fall horizons. The uppermost part of this unit consists of conglomerates alternating with thin intervals of mudstone.

The basal part of this unit includes strata previously mapped as 'Tobas Grises Inferiores' (Lower gray tuffs) (Truempy and Lhez, 1937; De la Motta, 1957). According to criteria used in this study, there is no 'lower gray tuff' unit (= Toba La Higuera; Yrigoyen, 1993a) bounding the base of the La Pilona Formation, and thus, it should be excluded from the stratigraphic nomenclature. Based on age constraints provided by radiometric data, La Pilona Formation has been assigned to the Late Miocene, between 11.1 and 9.8 Ma (Yrigoyen, 1993 a and b). In this study, the magnetostratigraphic results for these deposits support a Late Miocene age assignment, but it is considered that these deposits accumulated

between ~11.7 and ~9 Ma.

Tobas Angostura unit (Yrigoyen, 1993a)

This unit, widely known as 'Tobas Grises Superiores' (Upper gray tuffs) (Truempy and Lhez, 1937), unconformably overlies the La Pilona Formation. It comprises very distinctive volcanogenic deposits characterized by their gray colour and weathered appearance and reaches a thickness of 100-120 m in the Tupungato area (Yrigoyen et al., 1954). These rocks are mainly reworked tuffs, sandstones with varying amounts of pyroclastic material and air-fall horizons. Conglomeratic intercalations are more abundant in the basal and upper parts of the unit.

Based on recent radiometric data consisting of a ^{40}K - ^{40}Ar hornblende age of 9.7 ± 1.1 Ma and a biotite age of 10.0 ± 0.5 Ma, Yrigoyen (1993 a and b) assigned the Tobas Angostura Formation to the Late Miocene (~9.8 Ma). In this study, the magnetostratigraphic results for these deposits support a Late Miocene age assignment but indicate that these deposits accumulated between 8.9 and 8.7 Ma.

Río de los Pozos Formation (Yrigoyen, 1993a)

This unit, widely known as 'Serie Amarilla' (Yellow sequence) (Truempy and Lhez, 1937), is composed of mudstones, tuffaceous mudstones and sandstones of yellowish colouration, and fine-grained conglomerates which become more abundant towards the top of the unit. In the Tupungato area, this formation exhibits a variable thickness, from 230 to 400 m, due to the unconformity that separates this unit from the overlying Mogotes

Formation. Based on radiometric dating and new paleontologic discoveries of fossil mammals, as well as on revision of fossils collected previously in the Tupungato area, Yrigoyen (1993 a and b) assigned the Río de los Pozos Formation to the Upper Miocene - Lower Pliocene. The present magnetostratigraphic study carried out on these sediments supports the assignment to the Upper Miocene. On the basis of the thickness of this formation and the rate at which the sediments accumulated (cf. Chapter 4, sediment accumulation rates), it is considered unlikely that sedimentation continued into the Lower Pliocene.

Mogotes Formation (= *Conglomerados de los Mogotes*, Truempy and Lhez, 1937)

This unit, which rests unconformably upon the Río de los Pozos Formation, is comprised of a ~2000 m thick sequence of medium to coarse conglomerates. They are interbedded with red mudstones and sandstones and a few tuffaceous horizons. These deposits probably represent proximal alluvial-fan facies, although a possible glacial origin has been suggested (Cortés, in Ramos 1990). Although the stratigraphic age assignment of this formation has been a matter of discussion for years, several authors have assigned this unit to the Upper Pliocene (e.g., Truempy and Lhez, 1937; Polanski, 1963; Pascual and Odreman Rivas, 1973). According to Yrigoyen (1993a), this age designation is supported by stratigraphic correlations with units outcropping in surrounding areas in conjunction with radiometric data. The fossiliferous Bajada Grande Formation of the Huayquerías region is considered stratigraphically equivalent to the Mogotes Formation of the La Piloná - Tupungato region (cf. Yrigoyen 1994). Radiometric data provided by Marshall et al. (1986) from air-tuff horizons interbedded with strata from the basal and middle members of the

Bajada Grande Formation yielded ages of 2.6 ± 0.1 Ma and 2.4 ± 0.3 Ma, respectively, and therefore both formations are of Late Pliocene age (cf. Yrigoyen, 1994).

3. METHODOLOGY

(a) Magnetic stratigraphy

Because the primary objective of the current study is to establish a precise chronology of the Neogene strata to link thrust motions in the western thrust belts with the sedimentary record in the adjacent foreland basin, these nonmarine strata have been dated using magnetic polarity stratigraphy calibrated with ^{40}Ar - ^{39}Ar isotopic dates of several interbedded tuff horizons.

Magnetostratigraphy utilizes the facts that the Earth's magnetic field reverses at irregular intervals and that sedimentary rocks record the pattern of magnetic reversals as they are deposited. The Earth's magnetic field approximates a dipole and is roughly symmetrical about the Earth's rotation axis (cf. Butler, 1992). It is generated by electrical currents in the Earth's core. For reasons that are poorly understood, the currents flowing in the core reverse their direction at irregular intervals (longer than 10^5 years), producing a reversal in the polarity of the dipole magnetic field. By convention, the present configuration of the dipole field is referred to as normal polarity. The dip or inclination of the field is directed downward in the northern hemisphere and upward in the southern hemisphere. The opposite configuration is defined as reversed polarity (cf. Butler, 1992).

As magnetic reversals are globally synchronous, they are recorded simultaneously in rocks all over the world. When the pattern of normal and reversed polarity intervals is established for a particular section, it can be correlated with the global Magnetic Polarity Time Scale (MPTS), a continuous record of polarity changes which is based on the marine magnetic record. The main reason for the potentially higher fidelity of the marine magnetic record, when compared with the continental magnetic record, is the notable continuity of the geologic processes by which new crust is formed along the mid-ocean ridges.

If a correlation between the two is achieved, the chronologic data from the MPTS can be applied to the local stratigraphic section. In many cases, a reliable correlation is achieved with the assistance of isotopic ages from volcanic ashes or igneous rocks in the local section. Using this technique, the age of events occurring within the local stratigraphy can potentially be dated with a resolution on the order of 10^5 years. This is particularly useful in terrestrial environments where the precision of biostratigraphic ages is not high because faunal stage boundaries are not reliably tied to a chronometric scale.

(b) Sedimentation and tectonics

Back-arc foreland-basin development results from deformation of a fold-thrust belt because of flexure of the lithosphere under tectonic loading. A variety of studies have quantified the relationship between thrust loading and basin subsidence (e.g., Beaumont, 1981; Jordan, 1981), some of which include more refined models that predict the stratigraphic response to thrust deformation (e.g., Flemings and Jordan, 1990). Because of the causal

relation between thrust emplacements and tectonically induced subsidence, the history of thrusting can be derived by analyzing the subsidence history of the basin. The tectonic subsidence history is measured by examining the thickness of preserved strata, which largely reflects the regional subsidence pattern induced by thickening in the thrust belt. This approach, explained in detailed by Jordan et al. (1988), requires well constrained stratal accumulation histories derived from a magnetic polarity stratigraphy of the sedimentary fill of the foreland basin.

The recognition of thrust events by the use of foreland-basin stratigraphy is critically dependent on the availability of detailed time-controlled stratigraphic records which allow for inference of the subsidence history and on the interpretation of stratigraphic indicators of tectonic activity (e.g., unroofing petrology and facies migration). As shown by Jordan et al. (1988) and Jordan (1995) the relationship between thrusting and sedimentation is not straightforwardly interpretable because the basin-fill stratigraphy also records the influence of other independent controls (i.e., climate, bedrock lithology, eustasy) that may imprint such an strong signature that thrust motion cannot be discerned. A hierarchy of accuracy of the methods for dating motion on thrust faults is suggested by Jordan et al. (1988) so that the correct deductions can be made when using indirect correlations.

(c) Location of the study sections

In the La Pilona area, five Neogene formational units are recognized (for description of the units, see above). Magnetic-polarity stratigraphy was established for all of the

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formations except the youngest (Mogotes Formation).

Five magnetostratigraphic sections were compiled (fig. 1.4 a and c). Three of the sections are located on the western flank of the La Pilona anticline. Section I (Tosca de Mariño) consists of strata of the Divisadero Largo and Mariño formations. It includes two separate segments (Ia, Ib) lying 1.5 km apart and sharing a sandstone unit that is physically continuous from one segment to the other. Section II (Arroyo del Agua Blanca) covers the upper part of the Mariño Formation and a portion of the La Pilona Formation. Section III (La Higuera) is approximately 6 km from section II and consists of three formational units: the upper part of the Mariño Formation, the La Pilona and Tobas Angostura formations. Section IV (El Molino) is a short column measured in the eastern flank of the La Pilona anticline and is comprised of strata of the upper part of the Tobas Angostura unit and a portion of the Río de los Pozos Formation. This section was compiled to establish the polarity pattern of the latter unit as it could not be sampled in section III. Section V (Quebrada de la Angostura) consists of strata of Tobas Angostura and Río de los Pozos Formation. This section is located in the eastern flank of the Tupungato anticline, which is found ~25 km south from the La Pilona anticline. The regionally extensive ash-rich unit within the Neogene sequence, the Tobas Angostura unit, establishes the correlation between this section and sections III and IV.

4. APPROACH OF THIS INVESTIGATION

Before the beginning of this research project at Carleton University, in March of 1991,

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there was one week of reconnaissance in the field during which oriented rock samples were collected for testing suitability for paleomagnetic procedures. The project itself required two field seasons totalling 6 to 7 months of work. The first took place between September and December of 1992 and the second from April to June of 1994. In both, logistic support, such as field equipment and a van, was provided by the Geological Survey of Argentina. The field work involved geologic mapping at 1:50000 scale in the La Pilona anticline, making regional observations and constructing sedimentological profiles in the Tertiary units. One of the profiles was made in the eastern flank of the Tupungato anticline, located about 25 km south of the study region. Numerous rock samples were collected for three purposes: (a) paleomagnetic processing, (b) ^{40}Ar - ^{39}Ar dating and (c) provenance analysis.

To establish the magnetic polarity stratigraphy, samples were processed at the Paleomagnetism Laboratory of the Geological Survey of Canada. Time-consuming cleaning experiments, using mostly thermal demagnetization, and measurements of natural remanent magnetizations were conducted under the supervision of Dr. Ken Buchan, with the assistance of Garry Freda. Mineral separation of the samples collected for radiometric dating was undertaken at Carleton University under the supervision of Dr. Sharon Carr. Drs. Randall Parrish, Chris Roddick and Mike Villeneuve (GSC) provided assistance with acquiring, processing and interpreting data. Thin sections of samples for provenance study were made at Carleton University by Anett Briggs, and Ronald Conlon provided assistance in staining feldspars. In the study of thin sections, Dr. Pablo Guerstein was of invaluable help. The counting of sandstone clasts was made with an automatic point counter at Carleton University.

CHAPTER 2

⁴⁰Ar-³⁹Ar GEOCHRONOMETRY

1. INTRODUCTION

Neogene strata exposed in the southernmost part of the Precordillera fold-and-thrust belt, Mendoza province, Argentina, accumulated during the Miocene and Pliocene in response to deformation of the western thrust belts, i.e., the Principal and the Frontal cordilleras. In order to link thrust motions in these mountain belts with the sedimentary record in the adjacent foreland basin, these sedimentary rocks have been dated using magnetic polarity stratigraphy calibrated with ⁴⁰Ar-³⁹Ar isotopic dates. This chapter reports on the age determinations obtained by ⁴⁰Ar-³⁹Ar isotopic dating of interbedded tuff beds. Sample locations are shown in figure 1.4 (Chapter 1).

Eight samples were collected from air-fall deposits and reworked tuff beds. All but two of the samples were extracted from horizons interbedded in the nonmarine sections in order to tie the local magnetostratigraphy to the global magnetic polarity time scale (Baksi, 1993). The other two samples were collected from Quaternary levels with the purpose of constraining the timing of formation of the major structures, the La Pilona and Tupungato anticlines (fig. 1.4 b and c)

Mineral separates of visually unaltered glass, biotite, hornblende and plagioclase were dated using the ⁴⁰Ar-³⁹Ar step-heating method. Only three samples underwent total fusion

^{40}Ar - ^{39}Ar analyses. A first set of mineral separates was analyzed in 1993-1994 using both resistance furnace step-heating and laser total fusion techniques at the Massachusetts Institute of Technology (MIT) facility. The second set was processed in 1995 using the laser step-heating technique at the Geological Survey of Canada (GSC). The results of 15 ^{40}Ar - ^{39}Ar age determinations are summarized in table 2.1.

A complete overview of the ^{40}Ar - ^{39}Ar age-dating method including interpretations of results and technical aspects of this technique is provided by McDougall and Harrison (1988). A shorter review of the method, principles and applications is found in Hanes (1991). A brief summary of the ^{40}Ar - ^{39}Ar method and of the step-heating technique is described below.

2. OVERVIEW

(a) Basis of the ^{40}Ar - ^{39}Ar dating method

Both conventional K-Ar and ^{40}Ar - ^{39}Ar methods are based on the radioactive decay of parent ^{40}K to stable daughter ^{40}Ar . Potassium has three naturally occurring isotopes, whose abundances are 93.26% (^{39}K), 0.01% (^{40}K) and 6.73% (^{41}K); only ^{40}K is radioactive and experiences branching decay to ^{40}Ca and ^{40}Ar . Therefore, this method is most commonly used for dating K-bearing whole-rock samples and minerals that contain potassium in lattice sites, such as micas, amphiboles and feldspars.

In the conventional K-Ar method, potassium concentrations are measured by one of several well established methods, including flame photometry, atomic absorption spectrometry

Table 2.1. Summary of $^{40}\text{Ar}/^{39}\text{Ar}$ results

Sample	Mineral	Method	Apparent age (Ma) ^a	$^{39}\text{Ar}_p$ (%) ^b	Isochron age (Ma) ^c	$^{39}\text{Ar}_i$ (%) ^d	$(^{40}\text{Ar}/^{39}\text{Ar})_o$ ^e	MSWD ^f
MD-5	Hornblende	Laser step heating	8.74 ± 0.10	94.9	8.64 ± 0.17	94.9	295.5 ± 2.6	20.9
MD-5	Biotite	Laser total fusion	8.85 ± 0.37	100	–	–	–	–
MD-6	Hornblende	Laser step heating	10.34 ± 0.13	98.9	10.42 ± 0.20	100	295.5 ± 2.0	12.4
MD-6	Hornblende	Furnace	9.79 ± 0.75	81.3	11.04 ± 1.68	100	295.5 ± 2.5	18.4
MD-6	Biotite	Laser total fusion	10.75 ± 0.29	100	–	–	–	–
MD-8	Biotite	Laser step heating	0.39 ± 0.23	81.5	0.26 ± 0.10	100	295.7 ± 0.6	1.2
MD-8	Plagioclase	Laser step heating	0.25 ± 0.02	100	0.24 ± 0.05	100	295.6 ± 1.9	11.6
MD-9	Glass	Laser step heating	0.38 ± 0.01	89.0	0.39 ± 0.03	100	295.3 ± 1.7	8.7
MD-9	Biotite	Laser step heating	0.39 ± 0.06	80.0	0.33 ± 0.10	87.8	295.6 ± 1.0	3.6
MD-10	Hornblende	Furnace	9.18 ± 0.68	99.5	9.15 ± 0.42	99.5	295.5 ± 0.6	0.5
MD-10	Biotite	Laser total fusion	8.39 ± 0.18	100	–	–	–	–
MD-12	Hornblende	Laser step heating	–	–	12.03 ± 0.45	100	344.6 ± 18.0	22.5
MD-13	Hornblende	Laser step heating	9.70 ± 0.11	96.9	9.63 ± 0.22	95.9	298.0 ± 9.3	6.2
MD-15	Hornblende	Laser step heating	9.54 ± 0.13	95.1	9.54 ± 0.15	100	295.9 ± 1.1	3.9
MD-15	Biotite	Laser step heating	8.62 ± 0.95	78.0	8.48 ± 0.19	90.8	295.7 ± 2.6	25.1

Note: All uncertainties are quoted at the 2 σ confidence level. ^a Calculated using formulae and error assessments detailed in Roddick (1988). ^b Percentage of total ^{39}Ar evolved from heating steps on the plateau. ^c Determined from correlation plot, as explained in Roddick et al. (1980) and Roddick (1988). Inverse isochron ages were calculated using York's modified regression treatment as outlined in Parrish et al. (1987). Ages are reported with both realistic error and uncertainty in J factor (0.5% at 1 σ level). ^d Percentage of total ^{39}Ar evolved from heating steps used for isochron age regression. ^e Reported with realistic error. ^f MSWD (mean square weighted deviation) for regression.

and isotope dilution. In ^{40}Ar - ^{39}Ar dating the ratio of daughter ($^{40}\text{Ar}^*$) to parent (^{40}K) is measured in a single isotopic analysis, avoiding the need for separate potassium analysis.

The basis of this method is the conversion of a portion of naturally occurring ^{39}K into ^{39}Ar (denoted by $^{39}\text{Ar}_k$) by the reaction $^{39}\text{K} (n,p) \rightarrow ^{39}\text{Ar}_k$. This reaction, along with many others, occurs in the sample when it is irradiated with fast neutrons in a nuclear reactor. After an optimum amount of $^{39}\text{Ar}_k$ has been produced, the sample is placed in an ultra-high-vacuum system, where it is fused in one step or heated in several steps to release gasses within the sample. The gasses are then purified and analyzed isotopically in a mass spectrometer, in the same way as in conventional K-Ar dating. However, no spike is required in the ^{40}Ar - ^{39}Ar method because only isotope ratios are determined.

Upon measuring the $^{40}\text{Ar}^*/^{39}\text{Ar}_k$ ratio (proportional to $^{40}\text{Ar}/^{39}\text{K}$) and correcting for some interferences (reactions induced by neutron bombardment), the age of the sample can be determined by using the following equation:

$$t = (1/\lambda) \ln [(^{40}\text{Ar}^*/^{39}\text{Ar}_k)J + 1]$$

where t is the age of the unknown sample, λ is the total decay constant, and J is a dimensionless irradiation parameter.

It can be inferred from the above equation that, in contrast to the conventional K-Ar method, a sample with a uniform distribution of potassium and radiogenic argon does not need to be completely outgassed in order to obtain a valid age.

Because of the difficulties in determining the integrated fast neutron dose that a

sample receives, 'J' is calculated by irradiating a standard sample of well known K-Ar age together with the unknown sample in order to monitor the dose. By measuring the ratio of $^{40}\text{Ar}^*$ to ^{39}Ar extracted from the standard sample, 'J' can be solved as follows:

$$J = (\exp(\lambda t_s) - 1) / (^{40}\text{Ar}^*/^{39}\text{Ar})_{\text{standard}}$$

where t_s is the age of the standard.

During the irradiation of the sample, interfering argon isotopes are produced mainly from calcium and potassium. Significant amounts of ^{36}Ar and ^{39}Ar are derived from calcium, and therefore, proper corrections must be made (for details, see McDougall and Harrison, 1988). One useful by-product of the irradiation is the transformation of a proportion of ^{40}Ca to ^{37}Ar , which, along with the transformation of ^{39}K to ^{39}Ar , permits a measurement of the Ca/K ratio of the sample.

As in the case of the K-Ar method, a correction for the presence of atmospheric ^{40}Ar must be made. This correction is based on the assumption that all ^{36}Ar is atmospheric. By measuring the content of ^{36}Ar that evolved during the experiment, the ^{40}Ar of atmospheric origin can be estimated from the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio of the present-day atmosphere, which has a value of 295.5. Thus,

$$^{40}\text{Ar}^*/^{39}\text{Ar} = [^{40}\text{Ar}/(^{36}\text{Ar})_{\text{measured}}] - 295.5(^{36}\text{Ar}/^{39}\text{Ar})_{\text{measured}}$$

Due to the principles outlined above, ^{40}Ar - ^{39}Ar age estimates are potentially more

precise than K-Ar age estimates. Both potassium and argon are analyzed on the same sample aliquot, avoiding problems of sample inhomogeneities; mass spectrometric measurements of ratios allow more accurate and precise results than determinations of absolute abundances of potassium and argon. Samples with very low potassium content (less than 80 ppm) and very small samples may be analyzed (Hanes, 1991). However, the main advantage of the ^{40}Ar - ^{39}Ar dating method is that argon can be partially released by step-wise heating of irradiated samples. This property affords the greatest potential of the technique, i.e., resolution of the concentration profile of ^{40}Ar within minerals.

(b) The ^{40}Ar - ^{39}Ar step-heating approach

This method relies upon the assumption that during the step-heating experiment, ^{39}Ar and ^{40}Ar are evolved from the sample by a thermal diffusion process, be it controlled by grain dimension (e.g., Turner et al., 1966) or a thermal activation barrier (e.g., Fitch et al., 1969). Many release patterns, however, display behaviour apparently not attributable to diffusion theory (McDougall and Harrison, 1988).

The sample to be analyzed is heated inside an extraction line, evacuated to ultra-high-vacuum conditions in a series of steps at temperatures starting as low as 400°C and continuing up to 1400°C (in some cases higher temperatures are required). The argon isotopes released at each step are analyzed in order to obtain a ^{40}Ar - ^{39}Ar date (apparent age). The plot of these apparent ages against the fraction of ^{39}Ar released is called an 'age spectrum plot'.

In the ideal case of a sample which has suffered no argon loss since it passed through

its closure temperature (e.g., a rapidly cooled mineral), $^{40}\text{Ar}^*/^{39}\text{Ar}$ ratios should be constant and the age spectrum plot should yield a flat pattern called a 'plateau'. This is readily taken to indicate that the sample has remained a closed system to argon diffusion since the time indicated by the plateau dates. A plateau has been thought to reflect uniform distribution of argon throughout the crystal, assuming that ^{39}Ar is homogeneously distributed throughout the sample (Turner, 1968) (fig. 2.1).

If a sample has lost a proportion of $^{40}\text{Ar}^*$ some time after initial cooling, there will be variations in the $^{40}\text{Ar}/^{39}\text{Ar}_k$ ratios measured on the gas fractions successively released by diffusion and the result is a spectrum of ages. If the higher temperature dates remain relatively constant, the high temperature part is also called a plateau (see fig. 2.1 b and c). The lowest apparent age would indicate an upper estimate of the thermal overprinting. Such age spectra have been thought to reflect partial resetting due to radiogenic argon loss near the edges of a grain by reheating (Turner, 1968).

There are few disturbed age spectra from natural samples that match Turner's (1968) theoretical diffusive-loss model. Most commonly, the release pattern of samples that have suffered a post-crystallization thermal event do not clearly reflect the ages of crystallization and reheating. Several experimental artifacts, analysis of material crushed below its effective diffusion ratios, mixed phases, structural changes in the mineral during the step-wise degassing and incorporation of ambient argon during a reheating episode, among other factors, may complicate an age spectrum.

Fig. 2.1. Schematic diagrams showing model $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra. Top diagram portrays an idealized crystal in cross section, the middle diagram in each panel shows the concentration of radiogenic argon and neutron-induced ^{39}ArK across the crystal, and the lower diagram illustrates the $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum expected from measurement of argon extracted in successive steps at progressively higher temperatures from the idealized crystal. (a) The case of a crystal undisturbed subsequent to initial crystallization and rapid cooling. (b) The case in which partial loss of radiogenic argon has occurred from the crystal in geologically recent times, so that there is a marked gradient of radiogenic argon across the crystal from essentially zero at the grain boundary. (c) The same case as in b, except that significant accumulation of radiogenic argon has occurred since the reheating event owing to the passage of time. A maximum age for the time of the reheating is given by the $^{40}\text{Ar}/^{39}\text{Ar}$ age for the gas released in the first step of the experiment, and a minimum age for the primary crystallization of the crystal is provided by the apparent age measured on the gas released at the highest temperature. From McDougall and Harrison (1988).

3. METHODOLOGY

(a) Sample preparation

Eight full rock pails (each ~3 kg) of fresh material were collected in the study area. Thin section examination of the samples in the laboratory revealed which minerals were sufficiently fresh to be dated.

Crushing and grinding procedures were required for only half the samples collected because the other four samples were unconsolidated tuffs. Mineral separations for all samples involved hand magnetic and Frantz separation, shaking water table, sieving and heavy liquid techniques. The order in which these steps were performed varied slightly from sample to sample depending on the minerals present in each sample. The last step involved handpicking to achieve >99% purity in all the mineral separates.

Aliquots of 20-40 mg of high quality glass from Md-8 and Md-9 were achieved by handpicking glass shards 100-300 μm in diameter and without vesicles or impurities. The same methodology was used to obtain 20-40 mg plagioclase crystals from the same samples.

Hornblende separates weighing 130 mg were obtained from all samples except Md-8 and Md-9. All the separates were sized between 160 and 300 μm in diameter, and >99% purity was achieved by picking out by hand mainly biotite grains.

Pure biotite separates sized between 160 and 300 μm in diameter were obtained from all the samples. High quality biotite separates were achieved by selecting dark crystals free of inclusions. Total fusion analyses required 15-20 mg aliquots of biotite, and laser step-

heating experiments required 5 mg aliquots.

(b) Laboratory procedures and instrumentation

The separates were washed consecutively in distilled water, acetone and ethanol prior to being wrapped in individual aluminium foil packets for neutron irradiation. Mineral irradiations were performed in the research reactor at McMaster University in Hamilton, Ontario. The first set of samples consisted of 14 separates, five biotite (2-5 mg each separate), two glass (15-18 mg), three hornblende (86-119 mg) and four monitors (1.5-3.5 mg) arranged in an aluminium can 40 mm x 19 mm in diameter, which was irradiated for 1.4 h in the central peak flux position 5C. Fast neutron flux was monitored using Fish Canyon Tuff Biotite, FCT-3 (27.68 ± 0.03 Ma; M.J. Kunk, pers. comm., as quoted in Roddick, 1990), and as a result of the neutron fluence, the J factor was 3.5×10^{-4} .

The second set of samples sent for irradiation consisted of 12 mineral separates, five hornblende (40 mg each separate), four biotite (2.5-5 mg), two plagioclase (18-45 mg) and one glass (10 mg), which were arranged along with other unrelated samples in the Al can and irradiated in central peak flux position 5C for 2 h. Neutron flux was monitored with 10 aliquots (0.20 mg each) of Fish Canyon Tuff Sanidine, FCT-Kf (28.03 ± 0.18 Ma; Renne et al., 1994), and the resulting J parameter was 4.85×10^{-4} .

Extraction of argon from the first set of samples was accomplished using resistance furnace step-heating and laser fusion techniques at the MIT laboratory. Furnace step-heating extractions were performed using a double-vacuum assembly built by Modifications Ltd. The

furnace temperature was measured with a Re-W thermocouple and controlled by a programmable system. The duration of the extraction was 5 min for each step.

Hornblende separates from samples Md-6, Md-10 and Md-5 of about 100 mg each were analyzed using the step-heating technique. Sample Md-6 was step-heated at 950, 1150, 1250, 1350, 1400, 1500 and 1600°C, resulting in seven analyses. Sample Md-10 was step-heated at 950, 1275, 1350 and 1600°C, resulting in four analyses. No usable data were obtained from sample Md-5, in which step-heating resulted in gas above background only for the last increment.

Laser fusion experiments were performed on three mineral separates using a Coherent 10 W Ar-ion laser in multi-line mode. Several aliquots of each sample were placed into milled wells in a copper planchet which was inserted into a windowed vacuum chamber. The extraction procedure involved heating each aliquot 10 min by means of a defocused laser beam at a power of 5-8 W, which produced uniform glass beads of all analyzed material.

Biotite separates from samples Md-5, Md-6 and Md-10 were analyzed by the total fusion method. Each mineral separate was split into nine aliquots (Md-5, Md-6) or five aliquots (Md-10), which consisted of approximately four to five crystals each. Each aliquot was fused, and the gas evolved was purified and analyzed.

In both furnace step-heating and laser fusion experiments, evolved gases were purified using a SAES AP-10 Al-Zr getter operated at about 973°C and three SAES ST-172 Fe-Zr-V getters operated at varying temperatures. Gettering time was 15 min for furnace extractions and 5 min for laser experiments. The purified gases were analyzed with an MAP 215-50 mass spectrometer using a Faraday collector for furnace argon extractions and an

electron multiplier for laser samples. This instrument has a sensitivity at M/e 40 of $6.8 \times 10^{-6} \text{ cm}^3 \text{ STP/V}$ for a trap current of 200 μA and an accelerating voltage of 3 kV with the Faraday configuration, and $8.4 \times 10^{-10} \text{ cm}^3 \text{ STP/V}$ for the electron multiplier configuration. More details on these procedures and instruments are found in Hodges et al. (1994).

Laser step-heating experiments were performed for the second set of samples at the laboratory of the Geological Survey of Canada. After irradiation, each sample was split into several aliquots which were loaded into separate 1.5 mm diameter x 4 mm deep wells in a copper planchet. This planchet was placed under vacuum in a chamber topped with a Zn-Se window. This window, which is transparent to the beam of a Weck CO_2 , 45 W surgical laser, was attenuated by 20X and stepped incrementally from 2 to 45 W power. The edges of the beam clipped by a metal mask restricted the beam to a diameter of approximately 150 μm with roughly constant power distribution. The beam was manually panned for 1 min around each hole to provide an even heating of the sample. Quantitative temperature calibration was not possible since sample geometry, size and opacity to laser, among other factors, shift the heating spectrum to higher or lower absolute temperatures within any one hole.

Laser step-heating analyses were carried out on 12 mineral separates: 40 mg hornblende separates from samples Md-5, Md-6, Md-12, Md-13 and Md-15; 5 mg biotite separates from samples Md-8, Md-9, Md-13 and Md-15; a 20 mg glass separate from Md-9; and 20 mg plagioclase separates from Md-8 and Md-9. Each mineral separate was split into four to seven aliquots which were put into separate holes in a metal planchet. The CO_2 laser beam was focused on each hole, and the power level was increased in several steps (six to nine in most cases) until total fusion. In each step, the gas was extracted, cleaned and

analyzed.

The gas released was cleaned by passive equilibration of the gas with 3 SAES NP-10 of ST707 alloy at 400°C as well as a cold getter of SAES alloy 201 pellets for 2-5 min. Purified gasses were analyzed with a VG3600 mass spectrometer, a magnetic sector mass spectrometer with 60° extended geometry equipped with a Faraday collector and electron multiplier. Signal on the latter was measured across a 1×10^9 ohm resistor with a resulting gain, relative to Faraday, of approximately 50. Although sensitivity of the multiplier is gain dependent, it was estimated at $1.900 \times 10^{-9} \text{ cm}^3 \text{ STP/V}$.

Isotopic ratios were corrected by on-line computer for mass fractionation, mass-spectrometer sensitivity, blanks, interference reactions from K and Ca, and decay since irradiation.

Data obtained from samples Md-13 (biotite) and Md-9 (plagioclase) using the laser step heating method were too scattered to provide meaningful age results.

(c) Method of data presentation

Data from step-heating analyses are displayed in age spectrum diagrams, correlation plots, and tables 2.2 to 2.13. These tables provide complete argon analytical data including an integrated age that results from averaging all steps. Correlation plots and age spectrum diagrams show the best estimated age derived from analytical results, which are summarized in table 2.1. Data from laser total fusion analyses are shown in table 2.14.

In the age spectrum diagram, the apparent ^{40}Ar - ^{39}Ar age of each heating step is plotted

against the cumulative amount of ^{39}Ar released, thus defining an age spectrum for each sample. This diagram assumes a $^{40}\text{Ar}/^{36}\text{Ar}$ ratio of 295.5 for the non-radiogenic gas. Each filled box in the diagram represents a single step-heating fraction, with the vertical width of the box being a measure of the analytical error.

In age spectrum diagrams of samples processed using the furnace step-heating technique, the apparent ^{40}Ar - ^{39}Ar age of each heating step is plotted against the fraction of ^{39}Ar released over the entire experiment. For example, data from sample Md-10 hornblende consist of four analyses obtained from step-heating the mineral separate in four steps, thus resulting in four separate ^{40}Ar - ^{39}Ar dates from the sample.

In age spectrum diagrams of samples processed using laser step-heating technique, multiple step-heating analyses are combined to form a single result, whereby the fraction of ^{39}Ar released relative to the sum total of ^{39}Ar released for all analyses forms the x-axis. Thus, the apparent ^{40}Ar - ^{39}Ar age of each heating step is plotted against the cumulative amount of ^{39}Ar released from the sample. For example, data from sample Md-13 hornblende consist of 45 analyses, and each set of nine steps, corresponding with one of the five aliquots, is in effect a replicate of the same procedure.

The $^{40}\text{Ar}/^{39}\text{Ar}$ ratio, age and errors for each gas fraction were calculated using formulae detailed in Roddick (1988). Apparent ^{40}Ar - ^{39}Ar ages for each step were calculated using the decay constants recommended by Steiger and Jäger (1977). Analytical uncertainties for individual steps are reported at the 2σ level and do not include uncertainty in irradiation calibration of the amount of ^{39}K converted to ^{39}Ar (J factor). The error limit in the final age determination includes uncertainty in the J parameter, typically 0.5% (1σ).

On the isotope correlation plots (Roddick et al., 1980), the x-intercept (inverse $^{40}\text{Ar}/^{39}\text{Ar}$, comprising K-derived radiogenic ^{40}Ar and neutron-induced ^{39}Ar) defines the age of the sample, and the y-intercept (inverse $^{40}\text{Ar}/^{36}\text{Ar}$) represents the trapped Ar composition. Binary mixing between the two end members defines a linear array with negative slope. A regression line fitted through all the steps is defined using a modified regression treatment (York, 1969) in which the errors are multiplied by the square root of MSWD (mean square of weighted deviations) to account for scatter, as outlined in Parrish et al. (1987).

(d) Results and interpretation

Age spectrum analysis of **Md-12** hornblende (fig. 2.2a, table 2.2) from a reworked tuff horizon of Mariño Formation displays a pattern of incremental release for each aliquot that defines a profile ascribable to the presence of excess ^{40}Ar . The age spectrum of each aliquot is characterized by high apparent ages for the initial gas followed by a monotonic decrease and flatter release pattern with increasing temperature, eventually producing a reasonably well defined plateau.

The first three steps of each aliquot and also the fourth and fifth steps in aliquots b and c give an apparent age older than the plateau age. These steps, when totalled, constitute ~30% of the ^{39}Ar released. In most aliquots (a, d, e), a flat release pattern is exhibited after the third heating step which, when totalled, represents ~70% of the ^{39}Ar released.

Although the pattern of incremental release reveals the presence of excess argon, it is constructed with the assumption that any trapped argon has a $^{40}\text{Ar}/^{36}\text{Ar}$ value of 295.5.

Fig. 2.2. Sample Md-12 hornblende: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.2. Argon analytical data, sample Md-12 hornblende^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca} (x10 ⁻¹¹ cm ³ STP) ^c	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
5	.020	.071	.002	.009	6.74	88.5	75.4	171.9	0.0
10	.010	1.392	.084	.144	7.05	43.8	23.9	.9	.5
15	.008	5.998	.527	.653	12.40	18.6	13.5	.2	2.1
20	.008	5.325	.369	.538	10.13	23.9	12.5	.6	1.7
25	.019	4.845	.371	.531	13.34	42.4	12.6	.3	1.7
30	.005	5.294	.475	.564	9.38	14.9	12.4	.4	1.8
35	.007	4.485	.302	.445	8.31	23.5	12.5	.6	1.4
40	.003	2.072	.188	.230	4.17	22.8	12.2	.3	.7
45	.003	3.117	.256	.315	5.17	16.3	12.0	.3	1.0
5	.024	.115	.003	.014	8.32	86.0	73.5	65.8	0.0
10	.023	.529	.042	.077	8.77	76.9	23.0	4.7	.2
15	.063	7.356	.545	.776	30.98	60.5	13.7	.5	2.4
20	.036	8.700	.579	.922	26.50	40.0	15.0	.3	2.9
25	.017	7.258	.569	.777	16.98	28.8	13.5	.5	2.4
30	.018	10.856	.836	1.126	21.35	25.1	12.4	.4	3.5
35	.009	5.721	.392	.600	11.34	24.4	12.5	.7	1.9
45	.029	14.494	1.018	1.452	28.22	30.2	11.8	.3	4.6
5	.028	.128	.006	.013	9.15	89.7	63.7	30.7	0.0
10	.026	.435	.040	.054	9.23	83.5	24.6	15.2	.2
15	.104	5.367	.456	.610	41.93	73.5	15.8	1.5	1.9
20	.037	9.224	.821	1.050	26.19	41.3	12.8	.3	3.3
25	.022	7.528	.692	.860	20.43	31.3	14.2	.6	2.7
30	.025	10.717	.905	1.128	23.54	31.0	12.6	.6	3.5
35	.023	8.643	.705	.913	20.02	33.6	12.7	.5	2.9
45	.023	14.194	.991	1.461	27.64	24.8	12.4	.3	4.6
5	.027	.073	.010	.031	8.85	91.6	20.9	10.2	.1
10	.025	.954	.075	.104	9.27	78.3	16.8	2.4	.3
15	.057	11.644	.997	1.287	37.34	45.0	13.9	.5	4.0
20	.020	12.171	.780	1.233	24.06	25.1	12.8	.3	3.9
25	.013	9.509	.669	1.002	18.59	21.2	12.7	.5	3.2
30	.010	8.327	.494	.799	14.41	19.6	12.6	.3	2.5
35	.022	17.002	1.181	1.821	32.09	20.0	12.3	.4	5.7
45	.023	20.403	1.423	2.146	37.48	18.3	12.4	.2	6.7
5	.046	.148	.002	.023	14.06	96.2	19.9	30.4	.1
10	.040	1.665	.124	.208	15.47	76.2	15.4	1.7	.7
15	.056	10.582	.725	1.173	35.59	46.4	14.2	.5	3.7
20	.022	11.705	1.007	1.293	25.39	25.6	12.7	.4	4.1
25	.015	11.923	.767	1.296	23.64	19.1	12.9	.6	4.1
30	.020	14.528	1.030	1.506	27.72	21.0	12.7	.4	4.7
35	.016	7.925	.568	.803	15.94	29.8	12.1	.6	2.5
45	.019	17.537	1.199	1.815	31.06	17.9	12.3	.2	5.7
Total ^e	1.02	299.96	22.23	31.80	778.2	38.7	13.1	.2	
Conc.(/g)	54.55	16040	1189	1701	41620				

^a18.7 mg, J = .000485 ± 0.50 % (1σ). ^b Prior to 20 times optical attenuation. ^{c-e} See table 2.6

Therefore, in this particular case, the apparent age thus derived reflects an inappropriate choice of trapped argon composition (and no value is given in table 2.1 or on fig. 2.2a).

The correlation plot (fig. 2.2b) has the advantage, when compared with the age spectrum diagram, of allowing assessment of the composition of the trapped $^{40}\text{Ar}/^{36}\text{Ar}$ component. The regression line fitted through the 45 analyses defines an age of 12.03 ± 0.45 Ma from the intersection with the x-axis and a trapped $^{40}\text{Ar}/^{36}\text{Ar}$ value of 344.6 ± 18.0 from the intersection with the y-axis. There is significantly more scatter of the data (MSWD = 41.6, relative to 22.5) when the regression assumes a present-day atmospheric argon composition for the trapped component. The better fit to a line corresponding to 12.03 ± 0.45 Ma as well as the essentially non-atmospheric $^{40}\text{Ar}/^{36}\text{Ar}$ intercept indicate that not only the gas released in the first steps but also the remaining ~70% of gas has been affected by excess ^{40}Ar .

Initial steps with anomalously high apparent ages may reflect small amounts of excess ^{40}Ar in less retentive sites of hornblende (Harrison and McDougall, 1980). The origin of more radiogenic trapped argon is unclear, but degassing of K-rich minerals and incorporation of excess ^{40}Ar in hornblendes under conditions of thermal disturbance of xenoliths or wall rocks of a magma chamber is a plausible explanation (cf. Harrison and McDougall, 1981).

For **Md-13** hornblende (table 2.3), the ^{40}Ar - ^{39}Ar age spectrum from a tuff horizon of La Pilona Formation (fig. 2.3a) displays a release pattern for each aliquot that conforms to a profile consistent with excess ^{40}Ar causing high apparent ages for the initial gas followed by a flat pattern with increasing temperature producing a plateau. These initially high apparent ages correspond to the first one or two steps of each aliquot, which, when totalled, represent

Fig. 2.3. Sample Md-13 hornblende: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.3. Argon analytical data, sample Md-13 hornblende^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
	(x10 ⁻¹¹ cm ³ STP) ^c								
5	.031	.286	.031	.097	10.36	88.1	11.1	1.3	.2
10	.029	3.013	.301	.421	13.51	63.3	10.3	.9	.7
15	.037	8.101	.810	1.093	22.86	48.1	9.47	.21	1.8
20	.031	7.014	.687	.899	19.29	47.9	9.75	.39	1.5
25	.018	5.187	.480	.654	12.62	42.9	9.62	.38	1.1
30	.013	4.806	.439	.571	10.39	37.5	9.93	.30	.9
35	.019	5.177	.480	.643	12.75	43.3	9.80	.20	1.0
40	.012	7.895	.807	1.031	15.19	23.3	9.86	.26	1.7
45	.016	10.279	1.030	1.292	19.17	24.5	9.77	.35	2.1
5	.054	.119	.005	.007	16.23	98.9	21.6	684.1	0.0
10	.077	4.719	.466	.625	30.47	74.8	10.7	.6	1.0
15	.084	18.157	1.776	2.405	51.70	47.9	9.76	.11	3.9
20	.029	12.849	1.246	1.654	27.34	30.8	9.97	.19	2.7
25	.031	9.709	.937	1.245	23.08	39.6	9.77	.36	2.0
30	.036	14.845	1.484	1.937	32.74	32.7	9.92	.22	3.1
35	.022	10.642	1.065	1.399	22.05	29.4	9.70	.27	2.3
40	.026	17.387	1.719	2.194	31.60	24.1	9.54	.28	3.5
45	.021	11.916	1.227	1.476	21.62	28.9	9.09	.26	2.4
5	.066	.190	.005	.020	19.22	101.8	-14.78	38.38	0.0
10	.059	2.474	.237	.329	22.27	78.3	12.8	.9	.5
15	.097	17.697	1.666	2.258	55.20	51.9	10.3	.3	3.6
20	.064	16.939	1.663	2.171	42.52	44.8	9.43	.28	3.5
25	.047	17.076	1.697	2.189	38.41	35.9	9.82	.29	3.5
30	.042	17.375	1.696	2.152	36.50	33.7	9.82	.26	3.5
35	.025	14.933	1.502	1.880	28.08	26.8	9.53	.21	3.0
40	.028	16.598	1.667	2.134	31.90	25.7	9.69	.18	3.4
45	.019	7.304	.761	.901	15.18	36.9	9.28	.38	1.5
5	.046	.101	.002	.018	15.85	85.3	111.2	23.7	0.0
10	.057	3.585	.334	.447	22.65	73.9	11.5	.9	.7
15	.091	17.696	1.705	2.266	52.65	51.3	9.87	.16	3.7
20	.056	14.455	1.388	1.814	37.01	44.5	9.88	.25	2.9
25	.043	11.620	1.120	1.458	28.32	44.7	9.38	.13	2.4
30	.045	14.105	1.413	1.827	33.37	39.7	9.61	.25	3.0
35	.036	16.683	1.670	2.159	35.19	30.1	9.95	.19	3.5
40	.026	19.887	1.956	2.497	36.21	21.4	9.94	.14	4.0
45	.020	12.486	1.285	1.589	23.35	25.7	9.53	.16	2.6
5	.064	.237	.010	.032	20.01	94.7	28.5	22.9	.1
10	.053	4.312	.428	.571	23.72	66.6	12.1	.7	.9
15	.100	20.347	1.982	2.603	59.60	49.8	10.0	.4	4.2
20	.080	19.649	1.865	2.474	50.25	46.9	9.41	.40	4.0
25	.056	13.474	1.302	1.688	34.65	47.7	9.37	.28	2.7
30	.045	14.946	1.472	1.885	33.29	39.6	9.31	.21	3.0
35	.028	13.053	1.304	1.651	26.11	31.2	9.49	.18	2.7
40	.017	10.362	1.069	1.349	19.46	25.3	9.39	.24	2.2
45	.020	15.725	1.571	1.924	26.99	21.4	9.62	.18	3.1
Total ^e	1.91	485.41	47.76	61.93	1260.9	44.9	9.80	.1	
Conc.(/g)	52.6	13320	1311	1700	34610				

^a36.4 mg, J = .000485±0.50 % (1σ). ^bPrior to 20 times optical attenuation. ^{c-e} See table 2.6

<5% of total ^{39}Ar released. In most aliquots, the flat portion of each spectrum is a six- or seven-step plateau; integrating over all plateaus yields an age of 9.70 ± 0.11 Ma, representing ~95% of the total gas.

The isochron treatment of the portion of the released gas represented by the flat part of the graph (fig. 2.3b) yields an age of 9.63 ± 0.22 Ma with a trapped $^{40}\text{Ar}/^{36}\text{Ar}$ value of 298.0 ± 9.3 (MSWD = 6.2). The good fit of the regression line as well as the essentially atmospheric intercept indicates that this portion of the gas has not been affected by excess ^{40}Ar . Only the first one or two steps of each aliquot provide evidence of an excess ^{40}Ar component.

Initial steps with anomalously high apparent ages probably reflect minor absorption of excess ^{40}Ar in less retentive sites of hornblende (cf. Harrison and McDougall, 1980). In contrast to sample Md-12, later heating steps fail to show evidence of more radiogenic trapped argon. The isochron age is interpreted as representing the time of closure of the hornblende to release of radiogenic ^{40}Ar .

The **Md-6** hornblende collected from an air-fall horizon interbedded in La Pilona Formation (plate 2.1) was first analyzed using the resistance furnace technique (table 2.4), and later a duplicate was processed by laser step heating (table 2.5).

In the former case (fig. 2.4a), the age spectrum yields an age of 9.79 ± 0.75 Ma (which includes the 'J' uncertainty), which corresponds with the step where the bulk of the gas (>80%) was released. This pattern is not a plateau since it does not fulfil the criteria outlined in Fleck et al. (1977). The correlation plot of this sample (fig. 2.4b) yields an age of

Plate 2.1. (a) Crystal-rich fallout tephra interbedded with strata of the La Piona Formation, termed the White ash horizon ('Horizonte de Ceniza Blanca', Truempy and Lhez, 1937). This horizon is laterally traceable for at least 400-500 m. (b) Detail of the tephra interbedded with red mudstone.

a

b

Fig. 2.4. Sample Md-6 hornblende (resistance furnace step-heating method): (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.4. Argon analytical data, sample Md-6 hornblende^a, resistance furnace step-heating method

Temp. (°C)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos.	Apparent age (Ma) (+/- 2σ) ^c		³⁹ Ar (%)
	(x10 ⁻⁹ cm ³ STP) ^b								
950	.058	.013	.001	.006	17.36	97.9	41.7	650.2	.6
1150	.006	.018	.002	.007	2.03	86.9	24.5	29.9	.8
1250	.011	1.250	.061	.107	6.25	52.7	17.6	4.8	12.2
1350	.017	7.974	.368	.715	15.92	31.2	9.79	.50	81.3
1400	.002	.324	.021	.027	.81	72.0	5.42	2.36	3.0
1500	.003	.122	.006	.012	.86	94.3	2.72	8.83	1.3
1600	.002	.043	.003	.007	.54	95.5	2.37	8.06	.8
Total ^d	.10	9.74	.46	.88	43.8	66.1	10.8	4.3	
Conc.(/g)	.98	97.7	4.63	8.82	438.8				

^a 99.73 mg, J = .000355 ± 0.50 % (1σ).

^b All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.

^c Errors are analytical only and do not include the error in the irradiation parameter J.

^d Includes the integrated age. The uncertainty in J is included in the error.

a

b

Fig. 2.5. Sample Md-6 hornblende (laser step-heating method): (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.5. Argon analytical data, sample Md-6 hornblende^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
	(x10 ⁻¹¹ cm ³ STP) ^c								
5	.030	.111	.006	.018	8.78	100.8	-3.67	28.97	.1
10	.013	4.086	.203	.367	7.91	46.9	9.99	.86	1.5
15	.010	9.182	.433	.840	12.37	24.5	9.70	.36	3.4
20	.007	4.586	.213	.429	6.72	31.9	9.30	.36	1.8
25	.005	1.664	.073	.159	3.25	45.0	9.77	1.51	.7
30	.005	1.237	.056	.094	2.56	58.0	9.92	3.78	.4
35	.006	.674	.035	.064	1.98	82.5	4.74	2.91	.3
45	.005	1.117	.056	.098	2.40	60.5	8.41	2.52	.4
5	.069	.202	.012	.079	21.36	95.0	11.8	6.0	.3
10	.076	5.470	.270	.506	28.91	77.8	11.1	.7	2.1
15	.086	20.606	.981	1.837	47.60	53.6	10.5	.2	7.5
20	.036	17.241	.821	1.523	28.39	37.9	10.1	.2	6.2
30	.031	28.071	1.336	2.457	38.40	23.9	10.4	.3	10.0
45	.028	34.148	1.621	3.003	42.65	19.2	10.0	.2	12.3
15	.058	3.907	.227	.492	22.12	77.0	9.01	.79	2.0
20	.037	4.723	.229	.442	16.09	67.9	10.2	.6	1.8
25	.034	6.221	.291	.546	15.90	62.6	9.51	.48	2.2
30	.042	8.236	.396	.724	22.23	55.6	11.9	1.0	3.0
35	.030	9.498	.428	.805	18.69	48.0	10.5	.4	3.3
45	.027	15.290	.724	1.357	23.51	33.7	10.0	.3	5.5
50	.018	11.277	.520	.939	16.67	32.8	10.4	.3	3.8
55	.016	14.112	.663	1.229	20.49	23.8	11.1	.3	5.0
60	.012	15.361	.725	1.357	19.43	18.4	10.2	.1	5.5
20	.023	4.515	.217	.432	12.35	55.2	11.2	.5	1.8
25	.051	11.551	.558	1.104	28.35	52.8	10.6	.4	4.5
30	.026	12.361	.547	1.124	21.51	36.0	10.7	.3	4.6
35	.019	8.374	.376	.789	15.28	37.2	10.6	.4	3.2
40	.018	10.883	.507	1.011	17.17	31.1	10.2	.2	4.1
45	.008	7.592	.344	.647	10.50	22.5	11.0	.1	2.6
Total ^e	.83	272.30	12.87	24.48	533.5	45.8	10.3	.1	
Conc.(/g)	40.18	13240	625.6	1190	25940				

^a 20.57 mg, J = .000485 ± 0.50 % (σ).
^b Prior to 20 times optical attenuation.
^c All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.
^d Errors are analytical only and do not include the error in the irradiation parameter J.
^e Includes the integrated age. The uncertainty in J is included in the error.

11.04 ± 1.68 Ma, assuming a present-day atmospheric argon composition, with MSWD = 18.4.

Age spectrum analysis on Md-6 hornblende using the laser step-heating technique (fig. 2.5a) displays a flat pattern of incremental release for each aliquot conforming to a well developed plateau. In general, each plateau is defined by five step-heating increments which, when integrated over all aliquots, yield an age of 10.34 ± 0.13 Ma. These steps represents ~99% of the ^{39}Ar released.

The correlation plot of this sample (fig. 2.5b) yields an age of 10.42 ± 0.20 Ma assuming a present-day atmospheric argon composition, with MSWD = 12.4. Data points with large errors ($\sigma > 1\%$) correspond to the aliquot (a) containing the least amount of ^{39}Ar .

Total fusion analyses performed on nine aliquots of Md-6 biotite yield an integrated age of 10.75 ± 0.29 Ma. These results are shown in table 2.14a.

When hornblende and biotite ^{40}Ar - ^{39}Ar age results are compared, there is an overlap at 10.4 Ma. As the highest quality data were obtained from the laser step-heating analyses performed on hornblende, it is considered that 10.42 ± 0.20 Ma is the best isochron age estimate. This age records the time of the original crystallization or the time the hornblende cooled through the closure temperature (cf. Dodson, 1973).

Age spectrum analysis of **Md-15** hornblende (table 2.6, fig. 2.6a) from an air-fall horizon of Tobas Angostura Formation displays a flat pattern of incremental release for each aliquot conforming to a well developed plateau. In most aliquots, each plateau is defined by

Fig. 2.6. Sample Md-15 hornblende: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.6. Argon analytical data, sample Md-15 hornblende^a, laser step-heating method

60

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
	(x10 ⁻¹¹ cm ³ STP) ^c								
5	.076	.381	.003	.037	24.13	93.1	38.8	20.8	.1
10	.058	2.712	.347	.353	21.74	78.4	11.6	.8	1.4
15	.085	8.665	1.209	1.275	39.33	64.2	9.64	.36	4.9
20	.038	4.596	.529	.625	18.41	61.1	10.00	.53	2.4
25	.022	3.706	.404	.459	11.56	57.1	9.43	.72	1.8
30	.009	3.529	.457	.417	7.39	37.0	9.74	.63	1.6
35	.010	4.794	.629	.670	10.07	28.0	9.45	.27	2.6
45	.011	6.991	.941	.929	13.60	24.6	9.64	.12	3.6
5	.083	.642	.016	.086	26.06	93.8	16.2	17.7	.3
10	.035	1.807	.269	.299	14.68	71.1	12.3	1.0	1.2
15	.107	8.891	1.293	1.428	47.40	66.5	9.70	.77	5.5
20	.054	4.380	.578	.688	23.67	66.8	9.96	.66	2.7
30	.030	6.906	.920	1.031	20.07	43.6	9.59	.40	4.0
40	.021	9.144	1.290	1.437	21.62	28.2	9.43	.26	5.5
45	.012	6.582	.959	.955	13.98	25.8	9.47	.31	3.7
5	.063	.713	.024	.060	20.14	92.5	21.9	18.8	.2
10	.045	2.661	.293	.308	16.98	78.7	10.2	1.3	1.2
15	.068	6.202	.962	.961	30.60	65.8	9.49	.49	3.7
20	.042	4.744	.641	.589	18.82	66.0	9.47	.18	2.3
30	.035	4.762	.728	.697	17.79	57.6	9.44	.67	2.7
40	.038	9.350	1.423	1.391	25.99	42.7	9.34	.29	5.4
45	.012	5.773	.863	.758	11.95	30.8	9.52	.37	2.9
5	.079	.747	.016	.038	24.23	95.8	23.3	38.6	.1
10	.082	1.875	.287	.292	27.47	88.4	9.56	1.60	1.1
15	.060	7.481	1.148	1.092	29.67	59.6	9.57	.30	4.2
20	.044	5.756	.836	.824	22.18	58.5	9.73	.47	3.2
30	.046	7.585	1.094	1.051	25.04	54.0	9.57	.41	4.0
40	.062	11.767	1.689	1.571	34.96	52.4	9.24	.22	6.1
45	.060	5.828	.875	.856	26.98	65.9	9.37	.56	3.3
5	.077	.955	.052	.091	24.76	92.4	18.0	3.9	.3
10	.052	3.547	.400	.418	19.80	77.8	9.15	1.23	1.6
15	.064	7.821	1.152	1.174	31.42	60.3	9.28	.34	4.5
20	.029	3.922	.506	.519	14.26	60.0	9.58	.25	2.0
30	.046	9.893	1.338	1.314	28.06	48.9	9.52	.42	5.1
40	.035	7.384	1.129	1.047	22.29	46.8	9.88	.30	4.0
45	.013	1.499	.190	.221	6.42	61.4	9.81	.99	.8
Total ^e	1.70	183.99	25.49	25.96	793.5	63.4	9.75	.2	
Conc.(/g)	112	12049	1669	1700	51960				

^a 15.3 mg, J = .000485 ± 0.50 % (1σ). ^b Prior to 20 times optical attenuation.

^c All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.

^d Errors are analytical only and do not include the error in the irradiation parameter J.

^e Includes the integrated age. The uncertainty in J is included in the error.

six increments; the plateaus represent about 90% of the ^{39}Ar and yield an age of 9.54 ± 0.13 Ma. Large errors in the apparent ages correspond to the lowest step of each aliquot representing $\sim 1.5\%$ of the total ^{39}Ar .

The regression line fitted through the 36 analyses (fig. 2.6b) defines an age of 9.54 ± 0.15 Ma assuming a present-day atmospheric argon composition, with $\text{MSWD} = 3.9$.

Age spectrum analysis of Md-15 biotite (table 2.7, fig. 2.7a) displays a release pattern for each aliquot that defines a profile that can be ascribed to partial loss of radiogenic ^{40}Ar (Turner, 1968). The age spectrum of each aliquot is characterized by a low apparent age for the initial gas followed by a steep age gradient and then a flatter pattern with increasing temperature, eventually producing a reasonably well defined plateau.

In most cases, the first three lowest steps of each aliquot gave an apparent age younger than the flatter portion of the spectrum. These steps, when totalled, represent about 15-20% of the ^{39}Ar released. Only two aliquots (b, g) display flat release patterns defining well developed plateaus over four increments. The rest exhibit flat release patterns in the last two highest steps. The flat portions, when totalled, constitute about 75-80% of the ^{39}Ar released, yielding an integrated age of 8.62 ± 0.95 Ma.

The correlation plot of 24 selected data points corresponding to the flatter portion of each spectrum (fig. 2.7b) yields an age of 8.48 ± 0.19 Ma assuming a present-day atmospheric argon composition, with $\text{MSWD} = 25.1$.

Having evidence of loss of radiogenic ^{40}Ar in Md-15 biotite and thus a disturbed age spectrum, the hornblende age (9.54 ± 0.15 Ma) is interpreted as representing the best estimate of the closure age of the Md-15 sample. According to the Turner (1968) model for episodic

a **b**

Fig. 2.7. Sample Md-15 biotite: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.7. Argon analytical data, sample Md-15 biotite^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca} (x10 ⁻¹¹ cm ³ STP) ^c	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
3	.118	.032	.080	.569	38.19	91.3	5.12	.74	.4
5	.161	.045	.146	1.181	55.93	85.1	6.22	.42	.7
10	.321	.062	.409	3.458	125.11	75.8	7.70	.25	2.1
20	.461	.125	.884	7.405	209.13	65.1	8.66	.14	4.6
45	.221	.268	1.676	13.650	194.89	33.5	8.35	.05	8.5
3	.245	.030	.106	.738	76.92	94.2	5.34	1.82	.5
5	.198	.054	.176	1.352	70.54	83.1	7.75	.68	.8
10	.234	.060	.457	3.733	106.21	65.1	8.75	.24	2.3
20	.306	.180	.944	7.637	166.52	54.3	8.76	.16	4.7
30	.172	.505	.523	3.379	84.08	60.4	8.68	.25	2.1
45	.090	.324	.185	.676	33.63	79.3	9.07	.51	.4
3	.175	.014	.094	.723	55.45	93.2	4.60	1.30	.4
5	.215	.060	.191	1.607	75.55	84.1	6.60	.47	1.0
10	.348	.080	.494	3.921	138.10	74.4	7.95	.33	2.4
20	.486	.142	.915	7.362	220.66	65.0	9.22	.17	4.6
45	.204	.256	.348	2.270	82.84	72.6	8.79	.29	1.4
3	.214	.055	.117	.840	65.64	96.2	2.65	1.09	.5
5	.195	.062	.155	1.271	66.92	86.2	6.39	.46	.8
10	.347	.108	.451	3.587	135.05	75.9	7.99	.14	2.2
20	.699	.333	1.419	10.362	312.17	66.1	8.98	.09	6.4
45	.464	.604	1.772	14.052	277.68	49.4	8.80	.09	8.7
3	.259	.057	.129	.912	82.09	93.4	5.25	.99	.6
5	.426	.081	.323	2.282	141.73	88.8	6.15	.95	1.4
10	.382	.051	.395	2.990	140.78	80.2	8.19	.31	1.9
20	.783	.174	1.073	8.035	306.05	75.6	8.17	.19	5.0
45	1.420	.418	2.639	20.278	620.10	67.7	8.71	.14	12.6
3	.132	.065	.066	.484	41.85	93.0	5.31	.26	.3
5	.192	.061	.168	1.339	66.60	85.1	6.54	.61	.8
10	.225	.066	.349	2.944	92.68	71.8	7.81	.31	1.8
20	.471	.142	1.074	8.891	227.03	61.3	8.70	.14	5.5
45	.577	.515	1.940	16.411	327.52	52.0	8.42	.11	10.2
3	.140	.054	.090	.766	47.16	87.4	6.80	1.10	.5
5	.127	.074	.193	1.567	52.77	71.3	8.51	.29	1.0
10	.127	.062	.252	2.139	58.21	64.4	8.51	.28	1.3
20	.091	.134	.211	1.552	42.63	63.2	8.90	.54	1.0
45	.040	.137	.092	.548	16.94	68.9	8.46	.59	.3
Total ^e	11.266	5.490	20.536	160.911	4855.4	68.6		8.35	.1
Conc.(/g)	1547	754	2821	22103	672400				

^a 7.28 mg, J = .000489 ± 0.50 % (1σ). ^b Prior to 20 times optical attenuation.

^c All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.

^d Errors are analytical only and do not include the error in the irradiation parameter J.

^e Includes the integrated age. The uncertainty in J is included in the error.

^{40}Ar loss, the biotite isochron age would represent a minimum estimate of the original age, and the lowest apparent age (2.63 ± 1.10 Ma) would indicate an upper estimate of a thermal disturbance.

There is regional evidence for post-crystallization tectonism in the area that may account for the young ages in the low temperature release. This sample was collected from the western flank of the Tupungato anticline, whose deformation occurred prior to deposition of an air-tuff horizon dated at 0.39 ± 0.03 Ma (Md-9, glass) and after or during deposition of the Mogotes Formation, whose basal part has been dated at 2.6 ± 0.1 Ma (Yrigoyen, 1993 a and b).

Age spectrum analysis of **Md-10** hornblende (table 2.8, fig. 2.8a) from the Tobas Angostura Formation displays a flat release pattern conforming to a well developed plateau at 9.18 ± 0.68 Ma. The plateau is defined by three steps representing >99% of the total ^{39}Ar released, and most of the gas (~80%) evolved in the second step. The correlation plot of this sample (fig. 2.8b) yields an age of 9.15 ± 0.42 Ma assuming a present-day atmospheric composition, with MSWD = 0.5.

Total fusion analyses performed on five aliquots of Md-10 biotite yield an integrated age of 8.39 ± 0.18 Ma. These results are shown in table 2.14b.

The biotite age result is not in agreement with the hornblende isochron age, but total fusion analyses on biotite preclude an interpretation. However, it is worth noting that this sample was collected from the same unit as sample Md-15, where the biotite shows evidence of loss of radiogenic ^{40}Ar . If Md-10 had the same systematics as Md-15, then the biotite age

Fig. 2.8. Sample Md-10 hornblende: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.8. Argon analytical data, sample Md-10 hornblende^a, resistance furnace step-heating method

Temp. (°C)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar	Apparent age (Ma) (+/- 2σ) ^c		³⁹ Ar (%)
	(x10 ⁻⁹ cm ³ STP) ^b								
950	.151	.133	.002	.008	45.22	98.5	56.8	169.1	.5
1275	.174	8.982	1.389	1.331	70.70	72.9	9.21	.83	78.9
1350	.005	2.087	.230	.277	5.58	28.3	9.21	.59	16.4
1600	.012	.574	.108	.071	4.52	79.2	8.48	1.68	4.2
Total ^d	.34	11.78	1.73	1.69	126.0	80.3	9.40	1.0	
Conc.(/g)	2.88	99.06	14.54	14.19	1060				

^a 118.88 mg, J = .000355 ± 0.50 % (1σ).

^b All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.

^c Errors are analytical only and do not include the error in the irradiation parameter J.

^d Includes the integrated age. The uncertainty in J is included in the error.

would represent a minimum estimate of the original age. Based on these grounds, the hornblende isochron age (9.15 ± 0.42 Ma) is thought to be the best estimate of the closure age of the Md-10 sample.

The ^{40}Ar - ^{39}Ar age spectrum of **Md-5** hornblende (table 2.9, fig. 2.9a) from the Tobas Angostura Formation (plate 2.2a) displays a release pattern for each aliquot that conforms to a profile consistent with excess ^{40}Ar causing high apparent ages for the initial gas followed by a flat release pattern defining a well developed plateau. These initially high apparent ages corresponding to the first one or two steps of each aliquot ($\sim 5\%$ of total ^{39}Ar) decrease rapidly onto three- to seven-step plateau segments which represent, when totalled, about 75% of the ^{39}Ar release. The plateau portions of the aliquots yield, when integrated, an age of 8.74 ± 0.10 Ma.

The isochron treatment of the flat portion of the gas released (fig. 2.9b) yields an age of 8.64 ± 0.17 Ma assuming a present-day atmospheric composition, with MSWD = 20.9. When the regression line is not fixed to the atmospheric composition, the resulting age is statistically the same and the initial argon ratio does not deviate within the error from the expected value of 295.5, thus indicating that this portion of the gas has not been affected by excess ^{40}Ar . As in sample Md-13, the best evidence of an excess ^{40}Ar component is found in the first one or two steps of each aliquot.

Total fusion analyses performed on nine aliquots of Md-5 biotite yield an integrated age of 8.85 ± 0.37 Ma. These results are shown in table 2.14c.

When biotite and hornblende age results are compared, they overlap at 8.64 ± 0.17

Plate 2.2. (a) Air-fall horizon (white layer at the top of the hammer) interbedded with mudstone in the Tobas Angostura unit. (b) Quaternary air-fall deposit overlain by alluvial cover.

a

b

a **b**

Fig. 2.9. Sample Md-5 hornblende: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.9. Argon analytical data, sample Md-5 hornblende^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	%atmos. ⁴⁰ Ar.	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
	(x10 ⁻¹¹ cm ³ STP) ^c								
5	.023	.320	.004	.003	7.21	94.2	136.2	1356.4	0.0
10	.023	2.127	.246	.236	9.82	68.6	11.4	1.3	.8
15	.022	7.822	1.245	1.141	17.87	36.7	8.65	.21	4.1
20	.013	4.846	.804	.694	10.41	36.0	8.39	.17	2.5
25	.008	4.065	.562	.540	7.48	31.0	8.34	.18	1.9
30	.007	5.129	.708	.711	8.77	23.8	8.20	.16	2.5
35	.009	4.800	.669	.615	8.36	32.1	8.06	.25	2.2
40	.011	8.239	1.182	1.174	14.32	22.2	8.29	.08	4.2
45	.011	8.689	1.108	1.143	14.49	22.5	8.58	.22	4.1
5	.012	.584	.005	.011	4.59	78.6	79.0	106.3	0.0
10	.021	2.350	.269	.258	8.97	68.3	9.62	.68	.9
15	.019	7.242	1.178	1.065	16.56	34.5	8.89	.31	3.8
25	.014	6.955	1.104	.986	14.55	29.4	9.09	.26	3.5
35	.012	9.512	1.342	1.278	16.35	21.9	8.72	.25	4.6
40	.012	7.361	.999	.983	13.31	27.3	8.59	.16	3.5
45	.010	7.089	1.024	.963	12.45	24.0	8.59	.21	3.4
5	.024	.692	.007	.029	6.91	101.8	-3.70	20.63	.1
10	.018	1.784	.237	.240	8.62	62.7	11.7	1.2	.9
15	.038	7.478	1.038	.966	21.04	54.0	8.74	.60	3.5
25	.027	7.368	1.033	1.050	18.40	43.2	8.69	.13	3.8
35	.041	11.041	1.592	1.462	26.44	45.8	8.55	.13	5.2
40	.014	9.250	1.352	1.256	17.16	24.6	8.99	.14	4.5
45	.011	6.248	.861	.849	12.19	25.8	9.30	.36	3.0
5	.019	.792	.013	.021	6.60	86.4	37.0	9.6	.1
10	.020	3.674	.405	.410	11.07	52.1	11.3	.4	1.5
15	.025	9.268	1.172	1.219	20.86	35.5	9.63	.23	4.4
25	.022	7.027	1.044	1.007	16.74	38.5	8.92	.28	3.6
35	.016	6.795	.973	.935	14.46	32.2	9.16	.11	3.3
40	.016	8.918	1.140	1.142	16.81	28.9	9.13	.31	4.1
45	.011	7.888	1.142	1.135	15.11	22.3	9.04	.24	4.1
5	.015	.365	.010	.030	4.66	96.3	5.12	10.83	.1
10	.016	1.979	.200	.199	6.88	68.8	9.42	1.14	.7
15	.031	8.977	1.384	1.288	21.95	41.8	8.66	.16	4.6
20	.015	4.351	.518	.489	9.45	47.7	8.82	.34	1.7
25	.008	2.606	.276	.256	5.06	49.4	8.74	.44	.9
35	.014	5.303	.751	.780	11.67	36.6	8.27	.19	2.8
40	.027	6.334	.998	.974	17.49	46.3	8.42	.33	3.5
45	.012	3.567	.476	.445	7.66	45.2	8.23	.26	1.6
Total ^e	.67	208.835	29.071	27.983	482.7	41.0	8.89	.2	
Conc.(/g)	52.92	16496	2296.2	2210	38130				

^a 12.7 mg, J = .000485 ± 0.50 % (1σ).^b Prior to 20 times optical attenuation.^{c-e} See table 2.11.

Ma (hornblende isochron age), which is considered to be the best estimate for the closure age of the Md-5 sample.

The correlation plot of **Md-8** biotite from a Quaternary air-fall horizon (table 2.10, fig. 2.10b) yields an age of 0.26 ± 0.10 Ma assuming a present-day atmospheric argon composition, with MSWD= 1.2. Age spectrum portrayal of the data (fig. 2.10a) displays a flat release pattern with increasing temperature conforming to a well defined plateau segment over six increments in one aliquot (d). A plateau spectrum is also defined over three steps in aliquots a and b. All these plateau steps account for 80-85% of the total ^{39}Ar released, yielding an integrated age of 0.39 ± 0.23 Ma.

The correlation plot of Md-8 plagioclase (table 2.11, fig. 2.11b) yields an age of 0.24 ± 0.05 Ma assuming a present-day atmospheric argon composition, with MSWD = 11.6. The age spectrum plot of the data (fig. 2.11a) yields an integrated age of 0.25 ± 0.02 Ma, but no plateau spectrum is defined by any aliquot.

The isochron age results from biotite and plagioclase are in good agreement and overlap at the plagioclase age. Therefore, it is considered that 0.24 ± 0.05 Ma records the closure age of Md-8 sample.

Age spectrum analysis of **Md-9** glass (table 2.12, fig. 2.12a) from a Quaternary air-fall deposit (plate 2.2b) displays a flat pattern of incremental release for each aliquot conforming to a well defined plateau. Each plateau is defined by from three to eight increments (aliquot b has eight). These plateau steps, when totalled, represent 85-90% of the ^{39}Ar release,

Fig. 2.10. Sample Md-8 biotite: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.10. Argon analytical data, sample Md-8 biotite^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar.	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
	(x10 ⁻¹¹ cm ³ STP) ^c								
5	.911	.061	.212	1.417	271.24	99.3	1.26	1.19	5.1
20	.450	.197	.584	4.648	134.04	99.2	.21	.13	16.8
40	.036	.038	.092	.760	10.94	96.7	.42	.41	2.7
5	1.109	.070	.190	1.269	329.26	99.5	1.15	2.07	4.6
20	.610	.219	.501	3.553	181.31	99.5	.24	.27	12.8
40	.102	.082	.137	.906	30.44	99.1	.27	.11	3.3
5	2.357	.110	.281	1.514	698.94	99.6	1.46	3.66	5.5
40	.224	.102	.200	1.335	67.29	98.2	.81	.38	4.8
3	1.414	.261	.167	.909	419.41	99.6	1.50	6.79	3.3
5	.745	.449	.244	1.432	221.78	99.3	.94	1.62	5.2
10	1.271	.120	.381	2.880	376.17	99.8	.18	1.32	10.4
20	.713	.134	.444	3.286	212.29	99.3	.41	.61	11.9
25	.172	.108	.243	1.850	52.49	96.7	.82	.61	6.7
35	.107	.054	.170	1.256	31.85	99.1	.19	.46	4.5
45	.054	.133	.106	.655	16.25	98.5	.33	.81	2.4
Total ^e	10.275	2.138	3.952	27.670	3053.7	99.4	.56	.4	
Conc.(/g)	5304	1103	2039	14284	1.588E+6				

^a 1.94 mg, J = .0004885 ± 0.50 % (1σ).^b Prior to 20 times optical attenuation.^c All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.^d Errors are analytical only and do not include the error in the irradiation parameter J.^e Includes the integrated age. The uncertainty in J is included in the error.

Fig. 2.11. Sample Md-8 plagioclase: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.11. Argon analytical data, sample Md-8 plagioclase^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos. ⁴⁰ Ar	Apparent age (Ma) (+/- 2σ) ^d	³⁹ Ar (%)
	(x10 ⁻¹¹ cm ³ STP) ^c							
10	.022	1.713	.205	.999	6.63	98.6	.08	.11
20	.030	6.591	.390	2.442	10.05	89.2	.39	.11
45	.035	26.419	.486	6.124	12.01	86.1	.24	.03
20	.064	6.781	.559	3.104	19.72	95.3	.26	.10
45	.032	25.283	.483	5.895	11.58	81.6	.32	.03
20	.066	6.764	.531	3.051	20.71	93.7	.38	.10
45	.071	26.119	.761	6.717	21.85	96.2	.11	.04
20	.078	9.438	.697	4.108	24.52	94.0	.31	.09
45	.033	24.427	.526	5.992	11.04	87.3	.21	.03
Total ^e	.431	133.535	4.638	38.432	138.1	92.1	.25	.0
Conc.(/g)	3.81	1182	41.0	340	1222			

^a 113 mg, J = .000485 ± 0.50 % (1σ).
^b Prior to 20 times optical attenuation.
^c All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.
^d Errors are analytical only and do not include the error in the irradiation parameter J.
^e Includes the integrated age. The uncertainty in J is included in the error.

a

b

Fig. 2.12. Sample Md-9 glass: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.12. Argon analytical data, sample Md-9 glass^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca} (x10 ⁻¹¹ cm ³ STP) ^c	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos.	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
5	.037	.318	.690	2.744	12.95	84.3	.65	.09	2.5
7	.055	.516	1.535	5.880	18.82	85.8	.40	.03	5.4
10	.031	.510	1.103	4.202	11.32	81.0	.45	.02	3.8
15	.025	.355	.841	3.239	8.76	83.8	.38	.08	3.0
20	.009	.109	.207	.823	2.92	86.6	.41	.10	.8
45	.005	.187	.065	.280	1.72	88.0	.65	.79	.3
3	.024	.161	.279	1.138	7.48	94.9	.30	.12	1.0
4	.049	.407	1.048	4.183	15.98	91.4	.29	.08	3.8
5	.031	.444	.707	2.801	10.14	89.5	.33	.06	2.6
6	.014	.169	.217	.836	4.36	93.1	.32	.43	.8
8	.014	.225	.238	.948	4.55	92.4	.32	.29	.9
10	.029	.322	.649	2.424	9.88	86.8	.47	.16	2.2
15	.043	.328	1.332	5.115	14.49	86.8	.33	.03	4.7
45	.060	.767	2.563	9.654	22.07	80.8	.38	.03	8.8
2	.003	.045	.003	.015	.88	86.0	7.25	13.51	0.0
3	.053	.424	1.151	4.472	17.60	88.8	.39	.04	4.1
4	.051	.302	.741	2.851	16.37	91.5	.43	.09	2.6
5	.038	.411	.871	3.471	12.62	89.6	.33	.07	3.2
6	.017	.129	.276	1.050	5.48	91.8	.38	.24	1.0
10	.039	.401	.988	3.834	12.75	90.3	.28	.04	3.5
15	.034	.452	1.156	4.373	12.16	82.2	.43	.10	4.0
20	.029	.433	1.138	4.337	10.97	79.3	.46	.10	4.0
45	.016	.260	.552	2.104	6.02	79.2	.52	.06	1.9
2	.001	.091	.002	.005	.62	56.9	47.8	55.3	0.0
3	.014	.176	.135	.550	4.14	96.9	.20	.21	.5
4	.025	.273	.520	2.023	7.88	91.9	.28	.06	1.8
5	.035	.304	.685	2.668	11.09	93.5	.24	.06	2.4
6	.010	.169	.215	.809	3.02	93.2	.22	.05	.7
10	.046	.246	.476	1.894	14.59	92.8	.49	.17	1.7
15	.082	.799	2.268	8.658	28.46	85.0	.43	.03	7.9
20	.019	.260	.581	2.232	6.76	82.1	.47	.13	2.0
45	.014	.197	.367	1.409	4.53	88.3	.33	.06	1.3
2	.002	.123	.004	.012	.69	88.5	5.75	12.79	0.0
3	.014	.163	.158	.632	4.36	92.4	.46	.49	.6
4	.017	.176	.261	1.039	5.32	92.3	.35	.13	.9
5	.020	.320	.478	1.852	6.38	91.4	.26	.06	1.7
6	.009	.162	.190	.719	2.86	96.3	.13	.09	.7
10	.015	.186	.204	.811	4.50	97.8	.11	.14	.7
15	.037	.348	.920	3.541	12.63	86.9	.41	.05	3.2
20	.031	.313	.684	2.653	10.47	87.6	.43	.06	2.4
45	.063	.643	1.934	7.324	22.46	83.5	.44	.04	6.7
Total ^e	1.16	12.62	28.43	109.61	391.0	87.4	.39	0.0	
Conc.(/g)	28.8	313	706	2720	9705				

^a 40.29 mg, J = .000485 ± 0.50 % (1σ). ^b Prior to 20 times optical attenuation.^{c-e} See table 2.12.

Fig. 2.13. Sample Md-9 biotite: (a) ^{40}Ar - ^{39}Ar age spectrum; (b) correlation plot.

Table 2.13. Argon analytical data, sample Md-9 biotite^a, laser step-heating method

Laser power ^b (W)	³⁶ Ar _{tr}	³⁷ Ar _{Ca}	³⁸ Ar _{Cl}	³⁹ Ar _K	⁴⁰ Ar	% atmos.	Apparent age (Ma) (+/- 2σ) ^d		³⁹ Ar (%)
	(x10 ⁻¹¹ cm ³ STP) ^c								
3	.170	.020	.295	1.009	51.17	98.4	.69	.63	2.7
5	.067	.039	.197	1.052	20.39	97.3	.45	.31	2.8
10	.101	.069	.311	2.010	30.85	96.8	.43	.17	5.4
20	.088	.094	.330	2.575	26.51	97.6	21	.10	6.9
30	.011	.014	.032	.232	3.56	89.2	1.45	.83	.6
45	.006	.015	.013	.090	1.75	99.2	.13	1.12	.2
3	.303	.088	.394	.958	89.61	99.9	.10	1.41	2.6
5	.111	.056	.308	1.398	33.82	96.6	.72	.31	3.7
10	.084	.076	.233	1.450	25.16	98.4	.24	.19	3.9
20	.133	.137	.495	3.517	42.91	91.3	.93	.26	9.4
45	.039	.063	.146	1.177	12.02	96.8	.28	.16	3.2
3	.129	.018	.248	.743	38.55	98.7	.59	.59	2.0
5	.034	.037	.102	.629	9.95	99.6	.05	.30	1.7
10	.032	.053	.107	.706	9.81	97.7	.28	.28	1.9
20	.049	.046	.128	.841	14.73	97.5	.39	.37	2.3
45	.011	.055	.034	.214	3.71	90.8	1.39	1.12	.6
3	.372	.082	.428	1.736	110.90	99.2	.43	1.05	4.7
5	.198	.098	.421	2.325	60.06	97.6	.53	.31	6.2
10	.250	.159	.657	3.709	74.90	98.8	.21	.09	9.9
20	.336	.290	.939	6.219	102.53	96.9	.44	.19	16.7
30	.153	.166	.574	3.984	47.05	96.4	.37	.12	10.7
45	.033	.064	.100	.726	10.56	93.3	.85	.30	1.9
Total ^e	2.710	1.739	6.492	37.297	820.5	97.6	.45	.1	
Conc.(/g)	1606	1030	3846	22100	486100				

^a 1.688 mg, J = .000485 ± 0.50 % (1σ).^b Prior to 20 times optical attenuation.^c All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ⁴⁰Ar denotes trapped plus radiogenic Ar. Atmos. ⁴⁰Ar assumes a trapped Ar component of atmospheric composition.^d Errors are analytical only and do not include the error in the irradiation parameter J.^e Includes the integrated age. The uncertainty in J is included in the error.

Table 2.14. Argon analytical data, laser total fusion method

Aliquot	$^{36}\text{Ar}_{\text{tr}}$	$^{37}\text{Ar}_{\text{Ca}}$	$^{38}\text{Ar}_{\text{Cl}}$	$^{39}\text{Ar}_{\text{K}}$	^{40}Ar	% atmos. ^{40}Ar	Apparent age (Ma) (+/- 2σ) ^b	^{39}Ar (%)	
(x10 ⁻¹¹ cm ³ STP) ^a									
(a) Sample Md-6 biotite (4.67 mg, J = .0003445 ± 0.50 % (1σ))									
1	1.182	.326	.263	3.495	414.23	84.3	11.5	.6	7.6
2	1.365	.256	.247	3.779	475.84	84.8	11.9	1.0	8.2
3	1.251	.260	.252	3.974	436.49	84.7	10.4	.5	8.6
4	2.137	.544	.485	6.822	753.43	83.8	11.1	.7	14.8
5	2.232	.606	.577	8.522	810.33	81.4	11.0	.6	18.5
6	1.835	.476	.383	5.721	639.51	84.8	10.5	.8	12.4
7	1.609	.438	.361	4.890	554.57	85.7	10.0	.9	10.6
8	2.022	.371	.369	5.593	690.78	86.5	10.3	.9	12.1
9	1.447	.217	.235	3.390	482.12	88.7	9.97	1.13	7.3
Total ^c	15.08	3.50	3.17	46.19	5257.3	84.8	10.75	.29	
Conc.(/g)	3229	748.5	679.2	9890	1.126E+6				
(b) Sample Md-10 biotite (2.12 mg, J = .0003445 ± 0.50 % (1σ))									
1	.831	.252	.887	5.913	322.38	76.1	8.07	.56	22.7
2	1.156	.296	1.146	7.161	438.57	77.9	8.40	.15	27.5
3	.354	.110	.514	3.306	153.00	68.3	9.08	.43	12.7
4	.800	.209	.869	5.499	312.77	75.6	8.60	.24	21.1
5	.791	.155	.711	4.131	286.67	81.5	7.96	.31	15.9
Total ^c	3.93	1.02	4.13	26.01	1513.4	76.8	8.39	.18	
Conc.(/g)	1854	482	1947	12270	713900				
(c) Sample Md-5 biotite (5.2 mg, J = .0003445 ± 0.50 % (1σ))									
1	1.574	.404	.879	4.847	538.51	86.4	9.38	2.79	9.2
2	1.699	.432	1.011	5.736	580.02	86.6	8.43	.62	10.9
3	1.970	.521	1.347	7.314	682.22	85.3	8.49	.56	13.8
4	1.549	.378	1.151	7.121	554.60	82.5	8.44	.26	13.5
5	1.110	.322	.802	4.626	401.85	81.6	9.89	.82	8.8
6	2.026	.363	1.137	6.398	688.14	87.0	8.67	1.59	12.1
7	2.082	.429	1.177	6.590	703.19	87.5	8.27	.30	12.5
8	.981	.235	.712	4.340	355.58	81.5	9.40	.38	8.2
9	1.331	.257	1.020	5.876	482.22	81.5	9.39	.66	11.1
Total ^c	14.32	3.34	9.23	52.84	4986.3	84.9	8.85	.37	
Conc.(/g)	2754	643	1776	10160	958900				

^a All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar, and Ca, Cl and K denote Ar derived from these elements. ^{40}Ar denotes trapped plus radiogenic Ar. Atmos. ^{40}Ar assumes a trapped Ar component of atmospheric composition.

^b Errors are analytical only and do not include the error in the irradiation parameter J.

^c Includes the integrated age. The uncertainty in J is included in the error.

yielding an integrated age of 0.38 ± 0.01 Ma. Large errors in the apparent ages correspond to the lowest step of each aliquot and represent $\sim 2.5\%$ of the total gas.

The correlation plot of this sample (fig. 2.12b) yields an age of 0.39 ± 0.03 Ma assuming a present-day atmospheric argon composition, with MSWD = 8.7.

Age spectrum analysis on **Md-9** biotite (table 2.13, fig. 2.13a) displays a flat release pattern for each aliquot and produces a reasonably well defined plateau. The plateaus, defined by three to five step-heating increments, represent, when totalled, about 80% of the ^{39}Ar released. When integrated, these steps yield an age of 0.39 ± 0.06 Ma.

The regression line (fig. 2.13b) defines an age of 0.33 ± 0.10 Ma assuming a present-day atmospheric argon composition, with MSWD = 3.6. Four data points with large errors ($\sigma > 1\%$) are not shown in fig. 2.13b and are excluded from the calculation.

When glass and biotite age results are compared, they overlap at the glass age. Therefore, it is considered that 0.39 ± 0.03 Ma records the time of quenching of the Md-9 sample.

4. CONCLUSIONS

The ^{40}Ar - ^{39}Ar geochronometry carried out on samples collected from air-fall deposits and reworked tuffs interbedded with Neogene nonmarine strata and from Quaternary deposits of the Mendoza Precordillera provides age constraints on the accumulation history of these strata, as follows:

(1) The occurrence of a reworked tuff horizon in the uppermost part of the Mariño Formation indicates that the ~800 m thick sedimentary pile below the Md-12 horizon was deposited prior to 12.03 ± 0.45 Ma.

(2) Sedimentation of the La Pilona Formation was, in part, coeval with explosive volcanic episodes as evidenced by the occurrence of air-fall horizons dated at 9.63 ± 0.22 Ma (Md-13) and 10.42 ± 0.20 Ma (Md-6), which are found interbedded with nonmarine strata placed stratigraphically in the middle part of the unit.

(3) The oldest strata of the La Pilona Formation must be younger than 12.03 ± 0.45 Ma (Mariño Formation) and not older than strata of the overlying Tobas Angostura Formation.

(4) Deposition of the Tobas Angostura unit was, in part, coeval with eruptive episodes, as evidenced by tephra layers preserved in the lower part of the unit and dated at 9.15 ± 0.42 Ma (Md-10) and 8.64 ± 0.17 Ma (Md-5) in both limbs of the La Pilona anticline. Approximately 20-25 km to the south, in the eastern flank of the Tupungato anticline, an air-fall horizon from the basal part of the unit was dated at 9.54 ± 0.15 Ma (Md-15).

(5) When dates of the Tobas Angostura unit in the La Pilona anticline are considered together, they indicate that accumulation of the lower part of this unit probably occurred between 9.47 and 8.47 Ma; the age of the Río de los Pozos Formation conformably overlying the upper portion of the Tobas Angostura Formation has yet to be dated.

(6) Motion on the La Pilona structure occurred prior to deposition of an ash layer dated at 0.39 ± 0.03 Ma (Md-9) and at the same time or after deposition of the Mogotes Formation, whose basal part has been dated at 2.6 ± 0.1 Ma (Yrigoyen, 1993 a and b).

(7) Motion on the Tupungato structure occurred prior to deposition of an ash layer dated at 0.24 ± 0.05 Ma (Md-8) and at the same time or after deposition of the Mogotes Formation, whose basal part has been dated at 2.6 ± 0.1 Ma (Yrigoyen, 1993 a and b).

CHAPTER 3

MAGNETIC STRATIGRAPHY

1. INTRODUCTION

The goal of the current study is to establish a precise chronology for the Neogene nonmarine strata outcropping in the Mendoza Precordillera in order to link the sequence of deformational events occurring in the western thrust belts with the sedimentary record in the adjacent foreland basin. These sedimentary rocks have been dated using magnetic polarity stratigraphy calibrated with ^{40}Ar - ^{39}Ar isotopic dates on several interbedded tuffs (cf. Chapter 2).

This chapter describes the paleomagnetic study carried out on the Tertiary deposits and the age assignment for the individual magnetic polarity sections based on the correlation with the standard reversal chronology (Baksi, 1993). The locations of the magnetic polarity sections are shown in figure 1.4. A discussion of the age determinations based on both sources of chronological data, the ^{40}Ar - ^{39}Ar geochronometry and the paleomagnetic study, is presented at the end of this chapter. Sedimentation rates derived from the magnetostratigraphy are described in Chapter 4.

The magnetic polarity stratigraphy technique pioneered by Johnson et al. (1975) has been applied extensively to assess the age of nonmarine deposits. Of particular interest are studies carried out in Andean foreland basins in which the time of thrusting was investigated

using this technique (e.g., Jordan et al., 1988 and references therein, 1990, 1993; Beer and Jordan, 1989; Beer, 1990; Reynolds et al., 1990).

2. OVERVIEW OF PALEOMAGNETISM

(a) Origin of remanence

Paleomagnetism is based on the phenomenon that rocks acquire a permanent magnetization parallel to the Earth's magnetic field at or near the time of formation. This original magnetism is termed the primary magnetization. Between the time of formation of the rock and the present, secondary magnetizations may also be acquired. The vector sum of these primary and secondary components is called the natural remanent magnetization or NRM (Collinson, 1983).

Magnetic minerals

In the majority of the rocks, the NRM is carried by members of the iron-titanium oxide series of minerals (Irving, 1964). In igneous rocks the NRM is usually carried by magnetite or titanomagnetite. In highly altered igneous and metamorphic rocks, pyrrhotite, maghemite and hematite may also be present. In sedimentary rocks, hematite is often the dominant carrier of the remanence. For example, in red beds hematite may occur both as the red pigment responsible for the distinctive colour of the rocks and as black, polycrystalline particles (specularite).

The identification of the minerals which carry the NRM is often difficult because only a small portion of the magnetic minerals in a rock contributes to the remanent magnetization. In addition, the stable remanence may reside in very-fine-grained particles which are difficult to identify under a microscope. Measurement of blocking temperature spectra in paleomagnetic thermal demagnetization experiments (see below) can be useful. Other techniques that help in the identification of minerals or mineral phases which carry the NRM are reviewed by Collinson (1983).

The magnetization of a rock may be considered at the level of atoms, domains and mineral grains (Irving, 1964). Each atom has a magnetization due to the orbital motion and spin of its electrons. Electrons with opposite spins are paired. As a result their spin magnetizations cancel each other out. Atoms with unpaired electrons are said to have a magnetic moment.

The atomic moments align with an external magnetic field. Upon removal of the field, the atomic moments in most substances become randomized due to thermal fluctuations. These substances are called paramagnetic. However, in a few substances (e.g., iron) the unpaired electrons of adjacent atoms are coupled, so that the atomic moments are aligned parallel (or antiparallel) to each other, even in the absence of an external magnetic field. These substances are called ferromagnetic.

The magnetization process of a ferromagnetic substance is explained by Néel's 1955 domain theory (Irving, 1964). Relatively small ferromagnetic grains are single domain. In such grains, all the atomic moments are aligned. In larger multi-domain grains it is energetically more favourable for the atomic moments to be grouped into small regions or domains, each

with a distinct magnetization direction. Domains are separated from each other by narrow zones called Bloch walls (Tarling, 1971). The dimensions of the domains vary depending on the mineralogy, imperfections in the lattice, shape and size of the grains.

Ferromagnetism, defined in a general sense, can be divided into three types of behaviour depending on the ordering of atomic moments within a magnetic domain: ferromagnetic (e.g., iron), antiferromagnetic (e.g., ilmenite) or ferrimagnetic (e.g., magnetite). Ferromagnetic substances, in the restricted usage of the term, have adjacent atomic magnetic moments aligned parallel to each other, yielding a net magnetization. In other materials the crystal lattice is divided into two sublattices. Within each sublattice, the moments have parallel alignment. However, atomic moments on the two sublattices are antiparallel, yielding no net magnetization if these atomic moments are of equal size (antiferromagnetic), but yielding a net magnetization if they are of unequal size (ferrimagnetic).

Types of remanent magnetization

The mechanism by which the primary NRM is acquired depends on the mode of formation of a rock (Irving, 1964). Igneous rocks acquire a thermoremanent magnetization (TRM) as their constituent magnetic minerals cool in the presence of an ambient geomagnetic field. At high temperatures thermal fluctuations destroy the coupling of the atomic moments, and the rock behaves as paramagnetic. As the temperature falls, spontaneous magnetization appears at the Curie temperature (T_C), which is $\sim 580^\circ\text{C}$ for magnetite and 680°C for hematite. This magnetization is free to follow the external field because the relaxation time, the time for the magnetization to relax to its thermal equilibrium value, is short. At the blocking

temperature T_b , which is at or below T_C , the relaxation time increases rapidly and a thermoremanent magnetization becomes "frozen in". Changes in the external field direction that occur at temperatures below T_b are ineffective in changing the direction of magnetization. Sedimentary rocks acquire a depositional remanent magnetization (DRM) when permanently magnetized grains become oriented parallel to the Earth's magnetic field as they settle in water. Although there may be some disorientation when the particles reach the bottom and during subsequent drying and consolidation, a statistical alignment parallel to the field direction remains.

Deposition experiments using crushed sediments or synthetic particles show that the inclination of DRM (i.e., the dip of the magnetic vector below the horizontal plane) is sometimes less than that of the applied field, whereas the declination (azimuth of the magnetic vector) is accurately recorded (Collinson, 1983). This 'inclination error' could result from the presence of elongated magnetic particles which tend to lie with their long axes parallel to the deposition surface or from consolidation during burial.

Chemical remanent magnetization (CRM) is acquired when magnetic minerals are formed at temperatures below the Curie temperature. CRM is thought to arise from a process of nucleation (Irving, 1964). In the early stages, sufficiently small newly magnetic particles behave superparamagnetically, that is, they act like paramagnetic materials but have high intensities like ferromagnetic substances. Their spontaneous magnetizations follow an applied magnetic field. As a grain grows, its relaxation time increases, and the magnetization is stabilized. Thus, there is a critical blocking volume above which the remanence becomes "frozen in". In a grain growth, three stages may be envisaged: a superparamagnetic stage with

no remanence, a single domain stage with remanence of high stability and a multi-domain stage with remanence of lower stability (Irving, 1964).

The CRM can be superimposed on primary TRM or DRM and may have higher blocking temperatures than the primary remanence. A common example of the acquisition of more than one type of magnetization is observed in weathered igneous rocks, which contain a component of CRM carried by a weathering product or a late stage exsolution product superimposed on a primary TRM. Another example occurs in sedimentary rocks that initially possess a DRM carried by magnetite. Oxidation of magnetite to hematite will produce a CRM. If this process occurs immediately after deposition, the CRM may be considered equivalent to a primary remanence; when it takes place long afterwards, CRM is considered a secondary magnetization.

Over long time periods primary remanence may decay and be partially (or totally) replaced by viscous remanent magnetization or VRM (Irving, 1964). VRM results from thermal agitation of the atoms that comprise a magnetic grain. Therefore, it affects grains with lowest relaxation times and generally has T_b 's lower than those of the primary remanence.

Other types of secondary magnetizations may be present in a rock but their description is beyond the scope of this section. Magnetic cleaning or demagnetization procedures that progressively remove the least stable components of magnetization, such as VRM, are described in the next section.

(b) Stability of remanence and cleaning techniques

In many studies (e.g., magnetostratigraphy), it is the primary remanence that is of greatest interest. Therefore, it is first necessary to remove any secondary NRM. The procedure by which the primary component is isolated is termed demagnetization or magnetic cleaning. It can be successfully employed because the magnetic hardness, or resistance to demagnetization, of secondary viscous magnetizations is generally less than that of the primary NRM.

Two magnetic cleaning techniques are used routinely in paleomagnetic studies: stepwise alternating field and stepwise thermal demagnetization. Individual specimens are subjected to increases in alternating field or temperature in order to progressively remove the least stable components of magnetization.

Alternating field demagnetization

Once the intensity and direction of NRM are measured, a specimen is placed in an alternating field of peak intensity H_p which is then reduced slowly to zero. At H_p the direction of magnetization of all grains with a coercive force (the opposing applied magnetic field necessary to reduce the remanent magnetization to zero) less than H_p will follow the field as it alternates. As the field decreases, the magnetization of grains with progressively lower coercivities become fixed in either of the two orientations assumed by the field as it alternates, cancelling each other out. This procedure is carried out at progressively higher and higher magnetic fields. Therefore, the total magnetic moment of the grains will cancel out. The results of such a stepwise demagnetization may be expressed graphically as changes in the intensity of magnetization or as changes in the directions (Irving, 1964).

Thermal demagnetization

A specimen is heated to a preset temperature, cooled to room temperature in zero magnetic field and measured. The components with blocking temperatures less than the preset temperature are destroyed. By cooling in a zero magnetic field no new magnetization is acquired. The procedure is repeated several times selecting progressively higher temperatures in order to demagnetize further components. The results of thermal demagnetization, like those for alternating field demagnetization, are expressed as changes in the intensity and in the direction of magnetization (Irving, 1964).

3. METHODOLOGY

(a) Field work and sample collection

Geological sections were selected to maximize stratigraphic coverage. Stratigraphic thicknesses were measured with a tape and/or a Jacob staff using a Brunton compass and bench marks. The sections were traversed, where possible, at right angles to the strike of the beds. Paleomagnetic samples were collected following the program proposed by Johnson et al. (1975), which is designed to assure adequate site statistics and reasonable coverage of the magnetic reversal record. One hundred and eleven paleomagnetic sites separated by an average of 30 m were sampled: 39 sites from the Tosca de Mariño profile (section I); 17 sites from the Arroyo del Agua Blanca profile (II); 43 sites from the La Higuera profile (III); four sites from the El Molino profile (IV) and eight sites from the Quebrada de la Angostura

profile (V). At each site three separately oriented specimens were spaced about 1.5 m apart within the same sedimentary bed.

The preferred lithologies for magnetic sampling were clay and silt. However, fine-grained sandstones were occasionally collected when clays and siltstones were not available in the stratigraphic column and/or when they were too fissile to be easily sampled. When necessary a hammer was used to remove the most weathered surface of the sample. Sandstones were collected throughout the sections for clast provenance studies, which are reported in detail in the next chapter.

Each hand sample was oriented by first filing a horizontal plane on its surface with a hand rasp. Then the orientation of the plane was measured with a Brunton compass and marked on the sample. Finally, the sample was removed from the outcrop, labelled, wrapped with paper and placed in a plastic bag to be transported to the laboratory.

(b) Laboratory procedures and instrumentation

In the laboratory, the hand samples were cut and ground into 2.5 cm cubes. This procedure was carried out without the use of water because of the poorly consolidated nature of the samples. According to Johnson et al. (1975), the error associated with sample preparation is about $\pm 5^\circ$ in both declination and inclination. Special plastic specimen holders in which to fit the cubic specimens were constructed. The plastic which was used has very low magnetic susceptibility and can be considered nonmagnetic for the purpose of this work.

Measurements were carried out in a magnetically shielded room at the Geological

Survey of Canada in Ottawa. The walls of the room consist of two layers of electrical steel and reduce the Earth's magnetic field to less than 2000 γ . All specimens were stored in the shielded room for the duration of the experimental work.

Specimen remanences were measured with a DSM-1 digital spinner magnetometer. This magnetometer-computer system has a measuring range of 10^{-4} - 10^3 A/m. The specimen spins at 5 rps near a flux-gate sensor located in a magnetic shield. The computer calculates and stores the magnetic vector data for each spin. After six spins, in which the sample holder is rotated to different positions, the computer automatically calculates the average remanent magnetic parameters for each step.

Before starting the systematic measurements, cleaning experiments on pilot specimens were conducted to determine whether thermal or alternating magnetic field (AF) demagnetization provided the most effective method of isolating the stable component of magnetization. AF demagnetization was undertaken using a Schonstedt single axis demagnetizer, model GSD-1. An alternating current is passed through a coil, thereby producing an alternating field along its axis. Individual specimens were placed within the coil and demagnetized along each of three perpendicular axes. The intensity and direction of remanence were measured after each demagnetization step.

Ten pilot specimens were AF demagnetized in 15 to 20 steps to a peak field of 100 mT. The most complete runs involved stepwise demagnetization at 2.5, at 5.0, at 10 to 90 mT at intervals of 5 mT and at 100 mT. AF demagnetization to 100 mT had little effect on the magnetization on some of the pilot specimens. Blocking temperatures on companion specimens suggested that hematite was the major carrier of the stable remanence in these

specimens. Thermal cleaning experiments, described below, proved much more efficient than AF cleaning at eliminating the remanence.

Thermal demagnetization was carried out with a Schonstedt thermal demagnetizer, model TSD-1, modified to maintain temperatures within $\pm 5^{\circ}\text{C}$. This instrument consists of separate heating and cooling chambers. Specimens are placed eight at a time on a glass boat and pushed into the furnace. The chambers are arranged coaxially so that after the specimens reach thermal equilibrium, the specimen boat can be pushed directly into the cooling chamber. Isolation of the two chambers allows one batch of eight specimens to be cooled while another batch is being heated. After each demagnetization step, the intensity and direction of remanence are measured at room temperature.

Thermal demagnetization was carried out in steps. In most sites, at least one specimen was thermally demagnetized in 16 steps using the following peak temperatures: 100, 200, 250, 300, 350, 400, 450, 500, 530, 560, 580, 600, 620, 640, 660 and 680°C . If the first specimen that was measured from a given site revealed a stable component of magnetization, the second specimen was subjected to fewer heating steps, usually 250, 350, 450, 500, 530, 550 and 630°C . Additional steps were included when appropriate. In those cases in which the first specimen displayed a complex pattern of demagnetization, the detailed step-heating treatment was also used for the second specimen. The third specimen of the stable sites was demagnetized in two or three steps, generally 500, 550 and 600°C . Occasionally, more steps were included.

Measurements of magnetic susceptibility (K) were performed on most specimens before thermal cleaning. The instrument used for this purpose was a SI-1 (Sapphire

Instrument), designed to measure magnetic susceptibility in different sample orientations. Six orientations were used for computation to determine the values of the maximum, intermediate and minimum K axes for each specimen.

(c) Method of data presentation and analysis

The basic data obtained for each specimen on demagnetization are the direction and intensity of remanence. Directions of magnetization are expressed in terms of declination, which by convention is measured clockwise from geographic north, and inclination from the horizontal, denoted positive downward and negative upward. Directional data are plotted and discussed with respect to paleohorizontal. The structural correction was made about the local strike line of the beds. It should be noted that most sampling sites were located on the limbs of an anticline which plunges very gently ($\sim 5\text{-}10^\circ$) to the south but no correction for the plunge was applied because this correction yields insignificant change in the declination (e.g., Butler et al., 1984).

The NRM directions obtained during demagnetization are displayed on a Lambert equal-area projection. Those with positive inclination are plotted as closed circles and those with negative inclination as open circles. Intensity plots are used to display intensity of magnetization as a function of peak demagnetizing temperature. The intensity shown in these diagrams is normalized to the initial intensity of NRM measured at room temperature.

Intensity and directional changes observed during demagnetization are combined on a single diagram, the Zijderveld plot (Zijderveld, 1967). In the Zijderveld plot, the

magnetization vector measured in each heating step is projected onto two orthogonal planes, the horizontal plane and the north-up vertical plane with axes normalized to the NRM intensity. Straight-line segments on both projections of the Zijderveld plot correspond to elimination of a single magnetization component. Curved segments correspond to elimination of two or more components at the same time. Examples of these types of behaviour are described below. Magnetization directions for individual specimens are calculated using the least-squares fit method of Kirschvink (1980), based on linear segments identified on Zijderveld plots.

Specimen 2402 from La Higuera (HI) profile (section III in fig. 1.4a) exemplifies the simplest case of magnetic cleaning. In this case, a single, stable component of NRM of normal polarity (fig. 3.1a) is interpreted to be primary. On the equal-area stereonet, the high stability of the primary NRM is evident from the tight grouping of magnetization directions with progressive demagnetization steps. The data points in the Zijderveld diagram approximate a straight line which trends toward the origin.

In other cases a secondary component of magnetization is superimposed on the primary remanence. Specimen 0801 from Quebrada de la Angostura profile (fig. 3.1b; section V in fig. 1.4c) illustrates the case where the blocking temperatures of the secondary magnetization (20-200°C) do not overlap the primary remanence (200-680°C). The Zijderveld diagram displays two straight lines, one associated with the secondary component and the other with the primary component. Directions of each component are calculated using the Kirschvink (1980) method.

A common problem is the separation of the primary and secondary magnetizations

Fig. 3.1. Examples of magnetic cleaning results for individual specimens from La Higuera (HI) and Quebrada de la Angostura (LA) profiles. a) Specimen 2402 shows one single and stable component of magnetization. b) Specimen 0801 displays primary and secondary components whose blocking temperature spectra do not overlap. c) Specimen 0102 represents the case where blocking temperature spectra of the primary and secondary components partially overlap. Directional data are plotted with respect to paleohorizontal on a Lambert equal-area projection. Open symbols indicate up directions and closed symbols down directions. On the intensity plot, intensity relative to that of the initial NRM (JNRM) is plotted against peak temperature. On the Zijderveld plot, intensity and directional changes observed during demagnetization steps are combined on a single diagram, where open circles represent projection of the total magnetization vector onto the north-east plane (horizontal plane) and closed circles represent projection onto the north-up vertical plane.

in cases where blocking temperature spectra overlap (cf. Dunlop, 1979). Specimen 0102 from the HI profile (fig. 3.1c; in section III) illustrates the case where both secondary and primary magnetizations are removed between 200 and $\sim 500^{\circ}\text{C}$. However, the direction of the primary remanence is well determined above 530°C , where it defines an approximate straight line towards the origin. In this case, the direction of the secondary component cannot be determined because there is no corresponding straight-line segment in the Zijdeveld plot.

If primary and secondary components have distinctly different directions, they can be readily differentiated (e.g., fig. 3.1 b, c). If, however, they have similar directions, they cannot be easily separated. For example, in rocks of Neogene age, a viscous secondary component along the direction of the present Earth's magnetic field cannot be separated from a normally magnetized primary component, since both have roughly the same direction. Therefore, it is difficult to be sure that the remanence of a normally magnetized sample is not entirely viscous in origin. To circumvent this problem in this study, it was assumed that blocking temperatures of viscous components in normally magnetized samples were no higher than blocking temperatures of viscous components in samples with a reversed primary remanence.

In a few specimens the magnetization directions of the characteristic component on thermal demagnetization were too scattered to employ the Kirschvink method. In such cases, an end point was calculated as the mean of the semistable directions using Fisherian statistics (Fisher, 1953).

Fisherian statistical analysis was used to average the stable remanence directions for each site and to calculate the mean paleomagnetic poles. Virtual geomagnetic poles (VGP) were calculated for Class I and II sites using the classification system of Johnson et al. (1982),

which was modified from that of Opdyke et al. (1977). Sites with three samples and k (precision parameter) ≥ 10 are designated as Class I sites; sites with $k < 10$, but whose polarity is not in doubt are classified as Class II sites. In those sites that do not meet Class II criteria, the cause is that (a) the sample size is too small ($N \leq 2$), (b) two of the three samples from the site have nearly parallel magnetic vectors, with the third being divergent, or (c) the three samples from the site have $k < 10$. Class III designates those sites where the three specimens have widely divergent directions. In some cases, Class III sites are interpreted as representing transitional sites between zones of opposite polarity.

Magnetic polarity zonation logs were constructed for each section using the VGP data and correlated with the magnetic polarity time scale of Baksi (1993).

4. RESULTS

(a) General characteristics

The NRM of most of the samples in this study includes two components of magnetization. The characteristic component, interpreted as a detrital remanence magnetization (DRM), has either a north and up or south and down direction. This component was revealed in most cases above peak temperatures of 250°C . It occupies the highest T_b range in each sample and is carried by titanomagnetite, titanohematite or both minerals. The second component is a viscous remanence (VRM), which is usually directed north and up, and was likely acquired in the present Earth's magnetic field. It occupies a T_b

range below that of the characteristic component or occasionally overlaps it.

The distribution of the highest blocking temperatures of the viscous components at all sites with a reversed characteristic component is shown in figure 3.2. In most of the samples, the VRM falls between 100 and 250°C. In only eight specimens is it possible that the blocking temperature range of the viscous component extends to 350°C or higher. Six specimens are from sites 07, 17 and 23 of the Mariño Formation. The presence of the viscous remanence at high temperatures in these specimens, however, is somewhat uncertain due to the scattered nature of the results. Two specimens from sites 01 and 51 of the La Pilona Formation are more stably magnetized and required thermal cleaning to 500 or 530°C to eliminate the normally magnetized viscous component. An alternative explanation for the anomalously high blocking temperatures of the normal components of these reversely magnetized samples is that two stable components of opposite polarity co-exist within these specimens.

As almost all of the well defined viscous components at reversely magnetized sites have blocking temperatures below 250°C, it will be assumed that normally magnetized sites with blocking temperatures of ~300°C or greater carry a normally magnetized primary component. Assuming that normal components with blocking temperatures to 560°C are viscous would have little effect on the interpretation of polarity stratigraphy in this study, except in the lower part of the Mariño Formation. As most of the normal sites in the lower part of the Mariño Formation have blocking temperatures below 500°C the predominance of normal polarity in the lower part of the Mariño Formation could reflect a normal overprint which has not been removed by the demagnetization treatment rather than the primary

Fig. 3.2. Histograms of the highest blocking temperature of the viscous components at all sites with reversed characteristic component from (a) Mariño, (b) La Pilona and (c) Río de los Pozos formations. Viscous blocking temperatures above 400°C for specimens of the Mariño Formation are rather uncertain, as discussed in the text.

polarity of the beds. This interpretation, although less satisfying, cannot be completely ruled out.

The distribution of the intensity of NRM in all sites is shown in figure 3.3a. The intensities range from 1×10^{-5} to 5×10^{-4} A/m with 40% of the determinations falling between 1×10^{-4} and 3×10^{-4} A/m, and 33% between 2×10^{-5} and 5×10^{-5} A/m.

Magnetic susceptibilities range from 1×10^{-3} to 3×10^{-2} SI (fig. 3.3b). Samples from the Mariño Formation display higher susceptibilities than samples from the other Tertiary units. Most determinations which fall in the range between 1×10^{-2} and 3×10^{-2} SI and 50% of those in the range between 7×10^{-3} and 1×10^{-2} SI are from rocks of the Mariño Formation. The bulk of the determinations (75%) falling in the range between 1×10^{-3} and 7×10^{-3} SI correspond to samples from La Pilona, Tobas Angostura and Río de los Pozos formations.

Of a total of 111 sites whose magnetic polarity was measured, 71 sites (64.0%) are classified as Class I sites (Johnson et al., 1982), 32 sampling locations (28.8%) are Class II sites, seven sites (6.3%) are Class III, and one site (0.9%) did not yield any information of the primary remanence direction. The number and hierarchy of the sites for each stratigraphic section are presented in the discussion of each profile.

Mean directions for each site, after correction for tilting about a horizontal axis, were calculated using Fisher statistical analysis (Fisher, 1953). They are clustered to the north and up or to the south and down (fig. 3.4a). The site mean direction for the normally magnetized Class I and II sites is $D = 15.4$, $I = -47.4$ (precision parameter $k = 14.3$, radius of the 95% confidence circle about the mean direction $\alpha_{95} = 5.0^\circ$) with corresponding paleopole at

Fig. 3.3. Histograms of (a) intensity of magnetization (NRM) and (b) magnetic susceptibility (K). In both diagrams each individual value represents the average of two or three samples per site (Class I and II sites).

76.0°N, 5.9°E ($dm = 6.5^\circ$ and $dp = 4.2^\circ$, where dm and dp are the semiaxes of the oval of 95% confidence). The site mean direction for the reversely magnetized Class I and II sites is $D = 200.0^\circ$, $I = 49.9^\circ$ ($k = 19.1$, $\alpha_{95} = 5.3^\circ$), with a corresponding paleopole at 72.8° N, 17.9° E ($dm = 7.1^\circ$, $dp = 4.7^\circ$). The normal and reversed means are antipodal (fig. 3.4b), supporting the interpretation that these samples are accurate recorders of the Earth's ancient magnetic field.

The mean characteristic directions of normal and reversed site vectors obtained from each paleomagnetic section are shown in table 3.1, and the corresponding data are plotted on a stereonet shown in figure 3.5. The stereographic plots of figs. 3.4 and 3.5 show that magnetization directions from the Tosca de Mariño, Arroyo del Agua Blanca and La Higuerita sections of the western limb of the La Pilona anticline are rotated clockwise ~ 16 - 23° from the present axial dipole field, which has a zero declination.

However, the magnetization direction of the El Molino section on the eastern limb of the anticline is rotated slightly counterclockwise from the present axial dipole field. This suggests that the observed difference in declination between the two limbs may be related to internal strains in the system that have not been accounted for or that it may have been small block rotations about a vertical axis.

(b) Stratigraphic sections

The magnetic data and the magnetic polarity stratigraphy obtained from each of the stratigraphic sections are described below.

Fig. 3.4. a) Mean site directions for characteristic component of Class I and II sites plotted on an equal area stereonet. Open circles are on upper hemisphere, filled circles are on lower hemisphere. The mean directions of magnetization for normally and reversely magnetized sites are represented by open and filled triangles respectively, with circles of 95% confidence. b) Mean directions of normally (open symbols) and reversely magnetized sites (closed symbols) with circles of 95% confidence plotted on the upper hemisphere of stereographic projection.

Table 3.1. Mean paleomagnetic results

Section	N	Direction				Paleopole			
		D (°)	I (°)	α_{95} (°)	k	Lat (°N)	Lon (°E)	dm (°)	dp (°)
Tosca Mariño-A.del Agua Blanca									
Normal	32	19.7	-45.8	7.0	14.1	72.0	7.1	9.0	5.7
Reverse	19	205.2	50.7	8.2	17.7	68.7	23.3	11.1	7.4
La Higuera									
Normal	21	15.8	-46.8	9.1	13.2	75.5	4.7	11.7	7.6
Reverse	18	198.5	48.3	8.0	19.6	73.8	12.32	10.5	6.9
Q. de la Angostura									
Normal	4	2.9	-48.3	10.9	72.1	85.3	324.5	14.3	9.3
Reverse	3	178.0	49.6	15.8	62.2	86.7	258.7	21.0	13.9
El Molino									
Normal	4	344.8	-56.8	14.7	39.8	77.0	177.6	21.4	15.5
Combined data									
Normal	61	15.4	-47.4	5.0	14.3	76.0	5.9	6.5	4.2
Reverse	40	200.0	49.9	5.3	19.1	72.8	17.9	7.1	4.7

NOTES: N, number of sites used in statistics; D and I, mean declination and inclination, respectively; α_{95} , radius of the circle of 95% confidence about the mean direction; k, precision parameter (Fisher, 1953); Lat and Lon, latitude and longitude of the mean paleopole, respectively; dm and dp are the semiaxes of the oval of 95% confidence about the mean paleopole.

Fig. 3.5. Mean site directions of Class I and Class II sites plotted on an equal area stereonet for (a) Tosca de Mariño and Arroyo del Agua Blanca sections, (b) La Higuera section, and (c) Quebrada de la Angostura and El Molino sections. Symbols as in fig. 3.4 except for the El Molino section where upper hemisphere projections are represented by open diamonds. In each case, the mean of site means is shown as a triangle with its circle of 95% confidence except for the El Molino section where a star is used. Large symbols correspond to Class I sites, smaller symbols to Class II sites.

Tosca de Mariño (MA) section

This section includes two formational units, the Divisadero Largo Formation at the base and the Mariño Formation above (for location, see fig. 1.4a). The position of the Divisadero Largo - Mariño Formation boundary within the stratigraphic column is based on field observations and information available in published and unpublished maps (e.g., Roller, 1950; De la Motta, 1957). At the top of the section is a fault which indicates that at least a portion of the Mariño Formation is missing. The stratigraphy, location of the paleomagnetic sites and the paleomagnetic data from the Tosca de Mariño section are summarized in figure 3.6. Because of difficult access it was necessary to sample the portion of the section above 625 m at a distance of 1.5 km from where the portion below 625 m was sampled. However, the sedimentary rocks are continuous between the two localities. The reversal pattern for the section was established by collecting samples at 39 sites that were separated by an average of 30 m. The stratigraphic thickness between sites was up to 15-20 m for Divisadero Largo Formation and the lower member of Mariño Formation (Conglomerados violaceos), whereas in the upper part of the unit (Mariño member) the stratigraphic separation had to be extended to as much as 42 m, as most of the strata are conglomeratic beds unsuitable for paleomagnetic study. Sample material was mostly claystone, mudstone and siltstone except for the eolian member of the Mariño Formation where fine-grained sandstones were sampled.

Paleomagnetic data from 38 sites were suitable for establishing the magnetic polarity stratigraphy. Thirty-five sites are from the Mariño Formation, 21 (60.0%) of which qualify as Class I, 11 (31.4%) as Class II and three (8.6%) as Class III. Three sites are from the Divisadero Largo Formation, and they qualify as Class II. Data from one sampling location

Tosca de Mariño section

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Fig. 3.6. Magnetic polarity stratigraphy of the Tosca de Mariño section. (a) Simplified stratigraphic section, showing formations, basic textures, faults (F), unconformities (U); the arrow indicates the locality where the section was offset by 1.5 km. (b) VGP latitudes. Plotted points are from Class I sites (solid circles) and Class II sites (open circles). (c) Magnetic polarity zonation derived from the VGP latitude plot (b). Black zones have normal polarity, and white zones have reversed polarity. ⊕ represents a Class III site interpreted as recording a magnetic transition; ⊙ represents a Class I or II site with anomalous magnetization direction.

in the Mariño Formation did not provide information on the direction of the primary remanence. The 14 Class II sites did not attain Class I status for the following reasons. In seven sites (50.0%) the sample size was insufficient ($N=2$); in five sites (35.7%) the third specimen was unstably magnetized; in two sites (14.3%) with three specimens, k (precision parameter) < 10 . The VGP latitudes from which the magnetic polarity zonation log was derived were calculated for Class I and Class II sites (filled and open circles, respectively, in fig. 3.6).

The magnetostratigraphy of this section is notable for the number of reversals encountered (fig. 3.6). The lowest 34 m corresponding with the Divisadero Largo Formation has a short normal magnetozone (N_0) followed by a longer reverse polarity zone (R_0). The Mariño Formation, which is the unit of interest in this study, starts with a short reversed magnetozone (R_1). This is followed by a long normal interval (N_1 and N_2) between 13 and 250 m, interrupted by a short reversed section (R_2). Between 250 and 457 m, two reversed magnetozones (R_3 , R_4) and two normal magnetozones (N_3 , N_4) are present. Between 457 and 508 m a reverse polarity zone (R_5) is followed by a normal magnetozone (N_5). Between 508 and 609 m a long reversed interval (R_6 and R_7) is punctuated by one short normal polarity zone (N_6). A normal magnetozone occurs between 609 and 680 m (N_7), followed by a reversed interval (R_8). A normal polarity zone (N_8) is recorded at site 38, which is located at 731 m, but the length and location of this interval are uncertain. This site has dual polarity (see discussion below) and is located between sites of reversed polarity (R_8 and R_9). Due to the uncertainty in the length of this interval, figure 3.6 shows its possible age range. The section ends with reversely magnetized sediments (R_9) cut by a fault at 772 m.

In summary, 17 geomagnetic reversals spanning a 772 m stratigraphic section are recorded by the Mariño Formation at the sampling locality. The boundary between zones of different polarity is placed mid-way between consecutive normal and reversed data points except for sites 24 (305 m), 27 (385 m) and 38 (731 m), which are interpreted as transitional sites.

All but three sites in the Mariño Formation carry a consistent primary remanence, either normal or reversed. Magnetization directions in most sites are north and up or south and down. Typical examples of stable normally and reversely magnetized specimens from the Mariño Formation are shown in figures 3.7 and 3.8. Directions at most sites remain stable in the temperature range between $\sim 300^\circ$ and $580\text{--}660^\circ\text{C}$. Most samples attain a stable magnetization direction after removal of a soft viscous component between room temperature and 250°C . In some specimens, the characteristic component displays some scatter with progressive demagnetization steps, but the polarity is not in doubt (fig. 3.9).

As noted above, samples of sites 24, 27 and 38 do not show a consistent remanence direction suggesting that these sites may record magnetic transitions. For example, at site 27 the three specimens that were demagnetized show widely divergent directions (fig. 3.10a). Its location between sites of reversed polarity below and normal polarity above supports its interpretation as a transitional site. At site 24, the three specimens have anomalous directions, i.e., north-down, steep south-up and southeast-down directions. As at site 27, site 24 is located between sites of reversed polarity below and normal polarity above. Paleomagnetic results from site 38 (fig. 3.10b) show that two specimens (3802, 3803) are reversely magnetized and that two specimens from a single block sample (3801 a and b) are normally

Fig. 3.7. (a-c) Typical paleomagnetic results for normally magnetized specimens from the Mariño Formation in the Tosca de Mariño section. Symbols and conventions as in fig. 3.1. Random directional data in the higher temperature steps are not plotted on the stereonet.

Fig. 3.8. Typical paleomagnetic results for reversely magnetized specimens from the Mariño Formation in the Tosca de Mariño profile. Symbols and conventions as in fig. 3.1.

Fig. 3.9. Example of a specimen from the Mariño Formation where the characteristic component shows considerable directional scatter although the polarity of the specimen is not in doubt. Symbols as in fig. 3.1.

Fig. 3.10. Paleomagnetic results for individual specimens from two sites of the Mariño Formation that do not show a consistent characteristic direction of magnetization. (a) At site 38, two specimens (02, 03) carry reversed magnetization directions and two specimens (01a, 01b) normal magnetization directions. (b) At site 27, three specimens show widely divergent directions, i.e., south-down (01), north-up (02) and west-down (03). Symbols as in fig. 3.1.

magnetized. In view of these results, site 38 probably spans a magnetic transition although it is located between sites that are both of reversed polarity.

The characteristic component in the Mariño Formation has a wide range of blocking temperatures. Specimens from sites 10, 12 (fig. 3.7c) and 35 (fig. 3.7a) and specimens from sites 22, 32 (fig. 3.8b) and 30 (fig. 3.8c) are examples where the blocking temperature spectrum of the characteristic component is close to 680°C, the Curie temperature of hematite, indicating that this mineral is the major carrier of the remanence. There are also many cases where blocking temperatures are mainly below 580-600°C, suggesting that magnetite may be responsible for much of the stable remanence. Such examples are illustrated by specimens 3701 and 3901 (fig. 3.8a). This agrees with earlier work based on isothermal remanent magnetization studies which established the presence of abundant magnetite in rocks of the Mariño Formation (Costa, 1994). In some of the samples with blocking temperatures below 580°C it is unclear whether the remanence is carried by hematite or magnetite (e.g., 2503 and 2803, in fig. 3.7b). In specimens from three sites (26 and 37 in fig. 3.8 c and a, respectively, and 38 in fig. 3.10b), both magnetite and hematite could be the carriers of the remanence. For example, specimen 3701 (fig. 3.8a) has an intensity of remanence which drops markedly at 580-600°C suggesting that the stable portion of the remanence is carried by magnetite. Subsequent demagnetization steps to 620-680°C indicate that hematite is also present and carries an unstable magnetization.

Many samples attain a stable magnetization direction after removal of a soft viscous component, commonly parallel to the present Earth's magnetic field (table A1 of the Appendix). In most of the reversed sites, the normally magnetized viscous component has a

blocking temperature spectrum below 250°C. However, in six specimens from three sites (07, 17, 23), it is possible that the viscous blocking temperatures are significantly above 250°C (fig. 3.2a). For example, three specimens from site 23 (fig. 3.11a) show a swing towards a reversed endpoint although the directional data are highly scattered (fig. 3.11a). However, because of the apparent consistency of the magnetization directions (i.e., in the three specimens the directions change upon thermal demagnetization in a systematic way), it is interpreted that a reversed polarity is obtained. Thus, a scattered reversed direction is attained at temperatures of about 500-530°C in specimens 2301 and 2303 and possibly at 560°C in specimen 2302. Another example is provided by two specimens from site 07 (fig. 3.11b). In specimen 0702, a scattered reversed magnetization direction is attained at temperatures of about 400-450°C. In specimen 0701 it appears that the blocking temperatures of the normal component are as high as 600-620°C.

At the sampling locality, the Mariño Formation unconformably overlies Divisadero Largo Formation. The latter formation was sampled for paleomagnetic study in order to compile a complete magnetostratigraphy of the Tertiary deposits outcropping in the area. Although the Neogene units are of primary interest in the current research, thermal cleaning results from two specimens of the Divisadero Largo Formation are shown in figure 3.12 to illustrate the thermal behaviour of specimens collected from this unit.

Arroyo del Agua Blanca (AB) section

This profile covers two formational units, the upper part of the Mariño Formation and a portion of the La Pilona Formation (see location in fig. 1.4a). The magnetic polarity

Fig. 3.11. Paleomagnetic results for individual specimens from two sites in the Mariño Formation where the removal of a probable viscous component required temperatures significantly above 250°C. (a) At site 23, three specimens show swings towards a reversed endpoint although the magnetization direction data are highly scattered. Specimens 01 and 03 attain a scattered reversed direction at temperatures of about 500-530°C, specimen 02 possibly at 560°C. (b) At site 07, specimen 02 shows a scattered reversed magnetization above 450°C. In specimen 01, it appears that the blocking temperatures of the viscous component are as high as 600-620°C. Symbols and conventions as in fig. 3.1.

Fig. 3.12. Paleomagnetic results for two specimens from Divisadero Largo Formation in the Tosca de Mariño profile. Symbols and conventions as in fig. 3.1.

stratigraphy, sampling locations and associated data for this section are shown in figure 3.13.

At the bottom of the section is a fault block comprised of strata of the upper part of the Mariño Formation, bounded at the bottom by fault F_1 with strike-slip and reverse displacements and at the top by a nearly vertical fault, F_2 . A regional unconformity, although locally a disconformity, separates the Mariño Formation from La Pilona unit. This boundary is located on the basis of the field interpretation in conjunction with geochronological constraints provided by ^{40}Ar - ^{39}Ar analysis (see below).

The magnetic reversal pattern was obtained for this profile by sampling at 17 sites, 11 of which are in the La Pilona unit. Thirteen sites (76.5%) qualify as Class I; three (17.6%) as Class II and one (5.9%) as Class III. The Class II sites failed to meet Class I criteria because there were insufficient samples ($N = 2$). The stratigraphic separation between sites was usually 30 m, but it was extended up to 40 m between some sampling locations due to the presence of thick conglomeratic beds. Sample material was mostly claystone, mudstone and siltstone.

The magnetic polarity zonation log derived from the VGP latitudes (fig. 3.13) shows that the 71 m thick faulted block composed of sediments of the Mariño Formation records normal and reversed magnetozones (N_F , R_F). Above the faulted block, between 71 and 200 m, there are two intervals of normal (N_0 , N_1) and reversed polarity (R_0 , R_1). Above the unconformity that separates La Pilona Formation from the underlying unit, there is a thick normal magnetozone that extends from 200 to 438 m. Above the long normal zone, there are reverse and normal polarity zones (R_2 , N_3).

In summary, this section is notable for the thick normal magnetozone that occurs from 200 to 438 m in sediments of the La Pilona Formation. It also contains two ash horizons

Fig. 3.13. Magnetic polarity stratigraphy of the 'Arroyo del Agua Blanca' section. a) Simplified stratigraphic section, showing formations, basic textures (see legend in fig. 3.6), faults (F), unconformities (U), dated tuff horizons (star with sample number). b) VGP latitudes. Plotted points are from class I sites (solid circles) and class II sites (open circles). c) Magnetic polarity zonation derived from the VGP latitude plot (b). Black zones have normal polarity, and white zones have reversed polarity. T represents a Class III site interpreted as recording a magnetic transition.

located at 174 m (Mariño Formation) and 553 m (La Pilona Formation) that have been dated by the ^{40}Ar - ^{39}Ar method.

Samples collected from the Mariño Formation along the Arroyo del Agua Blanca valley carry a stable primary remanence. All sites have a consistent primary direction directed either north and up or south and down. Typical examples of paleomagnetic results for individual specimens are shown in figure 3.14. As in samples from the Mariño Formation described above, the characteristic component of these samples shows a wide range of blocking temperatures. For example, in specimens from site 43 (fig. 3.14a) the blocking temperature spectrum is close to 580°C, suggesting that magnetite carries the stable remanence; in specimens from sites 40 and 42 (fig. 3.14b) and 45 (fig. 3.14a) blocking temperatures close to 680°C indicate that hematite is the major carrier of the remanence. In some cases, such as specimens from site 41 (not shown) and site 42 (fig. 3.14b), both magnetite and hematite probably carry the remanent magnetization, with magnetite responsible for much of the stable remanence. In most specimens the stable magnetization is attained after removal of a viscous component between room temperature and 200°C (table A1 of the Appendix).

Samples collected from La Pilona Formation in the Arroyo del Agua Blanca profile carry a very stable primary remanence magnetization. Typical examples of paleomagnetic results for individual specimens are shown in figure 3.15. Directions at most sites remain stable in the temperature range between 250 and 640-660°C. Most samples attain the stable direction after removal of a soft viscous component which is eliminated between room temperature and 250°C (fig. 3.2). In a single specimen (5101) demagnetization up to 500-

Fig. 3.14. Typical paleomagnetic results for a) reversely magnetized and b) normally magnetized specimens from the Mariño Formation in the Arroyo del Agua Blanca profile. Symbols and conventions as in fig. 3.1.

Fig. 3.15. Typical paleomagnetic results for (a) reversely magnetized and (b) normally magnetized specimens from La Pilona Formation in the Arroyo del Agua Blanca profile. Symbols and conventions as in fig. 3.1.

530°C was required to remove the viscous component (fig. 3.15a).

All but one site (53) from the La Pilona Formation carry either a stable normal magnetization directed in most sites north and up at about 45-50° or a stable reversed magnetization directed south and down with similar inclination. Blocking temperatures indicate that the stable magnetization in these sites is carried largely by hematite (fig. 3.15).

Specimens from one site (53) carry stable but scattered directions of magnetization upon stepwise demagnetization (fig. 3.16). The widely divergent magnetization directions and the location of this site between sites of reversed polarity below and normal polarity above suggest that this site probably records a magnetic transition. Most blocking temperatures at site 53 are in the range up to 580-600°C suggest that magnetite is the dominant carrier of the remanence.

La Higuera (HI) section

The La Higuera section is comprised of three formational units, the upper part of the Mariño Formation, the La Pilona Formation and a portion of the Tobas Angostura Formation (for map location see fig. 1.4a). The location of the unconformities that separate these three units is based on field observations and information available in published and unpublished maps (e.g., Roller, 1950; De la Motta, 1957). The upper 10 m of this section represents the gradual transition between the Tobas Angostura and Río de los Pozos formations. The stratigraphy, the location of the paleomagnetic sites and the paleomagnetic data from the La Higuera section are summarized in figure 3.17.

The general magnetic reversal pattern for the section was established by collecting

Fig. 3.16. Paleomagnetic results for individual specimens of site 53 in the Arroyo del Agua Blanca profile. These specimens carry fairly stable but scattered magnetization directions suggesting that this site may represent a transitional site. Symbols and conventions as in fig. 3.1.

Fig. 3.17. Magnetic polarity stratigraphy of the La Higuerita section. (a) Simplified stratigraphic section, showing formations, basic textures (see legend in fig. 3.6), faults (F), unconformities (U), dated tuff horizons (star with sample number). (b) VGP latitudes. Plotted points are from Class I sites (solid circles) and Class II sites (open circles). (c) Magnetic polarity zonation derived from the VGP latitude plot (b). Black zones have normal polarity, and white zones have reversed polarity. ① represents a Class III site interpreted as recording a magnetic transition; ② corresponds with a Class III site where the three specimens show anomalous, divergent magnetization directions. ③ corresponds with a Class II site with anomalous magnetization direction.

La Higuera section

samples at 43 sites that were separated by an average of 30 m. The stratigraphic thickness between sites was up to 25 m for the lower part of the La Pilona Formation but greater in some portions higher in the section due to the lack of suitable material for paleomagnetism and/or to poor exposure. Sample material from La Pilona and Tobas Angostura formations was claystone, mudstone and siltstone. In the stratigraphically lowest two sites of the Mariño Formation, -fine-grained sandstones were sampled.

Of a total of 43 sampling sites, 28 (65.1%) qualify as Class I site, 13 (30.2%) as Class II, and two (4.7%) as Class III. Class II sites failed to meet Class I criteria because in nine sites (69.2%) there was insufficient sample size; in four sites (30.8%) two out of the three samples from each site have nearly parallel magnetic vectors, with the third being divergent. The VGP latitudes from which the magnetic polarity zonation log was derived were calculated for Class I and II sites.

The paleomagnetic data yield the following results. No information on magnetic polarity of the Mariño Formation was obtained in the lowest 67 m as the first site provided an anomalous direction of magnetization, and the second site was a Class III site. Between 67 and 120 m, the Mariño Formation records a reversed and a normal magnetozone (R_0 , N_0).

Above the unconformity that separates the La Pilona Formation from the underlying Mariño Formation, there is a reversed polarity zone followed by a thick normal magnetozone that extends from 264 to 500 m (N_1). Above the long normal zone, there is a reverse polarity zone (R_2), followed by a normal interval (N_2), which extends from 653 to 715 m. The La Pilona unit ends with 197 m of reversely magnetized sedimentary rocks (R_3). The 120 m of the overlying Tobas Angostura Formation records an interval of normal polarity (N_3).

In summary, this section is notable for the thick normal magnetozone that occurs from 264 to 500 m in sediments of the La Pilona Formation. In addition, the section contains two air-fall ash beds located at 503 m (La Pilona Formation) and at 955 m (Tobas Angostura Formation) that have been dated by the ^{40}Ar - ^{39}Ar method.

Most samples from the La Pilona Formation of the HI profile carry a very stable characteristic magnetization of either normal (fig. 3.18) or reversed (fig. 3.19) polarity. Directions in most sites remain stable in the temperature range between 250° and 660-680°C. Most samples attain a stable magnetization after removal of a soft viscous component at temperatures of 250°C or less.

Three specimens from sites 02, 03 and 19 display some scatter of magnetization directions, but the polarity of each specimen is not in doubt. For example, specimen 0202 (fig. 3.20) has a large soft viscous component which is removed at 200°C and a semistable reversed direction between ~200 and ~500°C.

Stepwise demagnetization results from site 06 fail to show a consistent magnetization direction (fig. 3.21), suggesting that this site may record a magnetic transition. The variation of magnetic declination and inclination observed in the three specimens and the location of this site between sites of reversed polarity below and normal polarity above are the grounds for supporting this interpretation.

Stepwise thermal demagnetization demonstrates that in many samples the blocking temperature spectrum of the characteristic component is close to the Curie temperature of hematite (figs. 3.18 and 3.19). In some samples, hematite and magnetite may both be carriers of the stable remanence. Examples are specimens 1203 (fig. 3.18b) and specimen 2204 (fig.

Fig. 3.18. (a, b) Typical paleomagnetic results for normally magnetized specimens from the La Pilona Formation in the La Higuera profile. Symbols and conventions as in fig. 3.1.

Fig. 3.19. (a, b) Typical paleomagnetic results for reversely magnetized specimens from the La Pilona Formation in the La Higuerita profile. Symbols and conventions as in fig. 3.1.

Fig. 3.20. Specimen 0202 exemplifies one of the few cases in the La Higuera profile where the characteristic component is only semistable (cf. with typical paleomagnetic results, figs. 3.18 and 3.19). A semistable reversed direction is attained between ~200 and 500°C. Symbols and conventions as in fig. 3.1.

Fig. 3.21. Paleomagnetic results of three specimens from site 06 of the La Pilona Formation in the La Higuierita profile. These specimens carry scattered magnetization directions suggesting that this site may represent a transitional site. Symbols and conventions as in fig. 3.1.

3.19a). In these samples, the intensity plots display a marked decrease in the intensity of magnetization up to 580-600°C, the Curie blocking temperature of magnetite, suggesting that at least a portion of the remanence is carried by magnetite. A less steep gradient is displayed in the intensity plots with subsequent demagnetization steps to 660-680°C, the Curie blocking temperature of hematite, indicating that hematite is also present.

Samples from the lower part of the La Pilona Formation (sites 02-06) and a few from higher stratigraphic levels (sites 19 and 21) have blocking temperatures no higher than 580-600°C, suggesting that magnetite may be the major carrier of the stable remanence. Examples are specimens 0401 and 2102 (fig. 3.19b).

As noted above, viscous components in the HI profile are typically eliminated at less than 250°C (fig. 3.1a). One single specimen (0102) from this collection required a temperature of 450-500°C to remove the viscous component of magnetization (fig. 3.1c). In this specimen, both secondary and primary magnetizations are removed between 200 and ~500°C and the direction of the primary remanence is well defined in the 530-640°C interval.

Samples collected from the Tobas Angostura Formation in the HI profile also carry a very stable primary remanence. All sites carry a normal magnetization directed north and up. Typical examples of paleomagnetic results for individual specimens are shown in figure 3.22.

After removal of a soft viscous component between room temperature and 200-250°C, directions in all samples remain stable in the temperature range ~250-640°C. Blocking temperatures indicate that the characteristic component is carried by hematite in specimens from sites 32 and 36. In specimens from site 34, most of the stable remanence is

Fig. 3.22. (a, b) Typical paleomagnetic results of specimens from Tobas La Angostura Formation in the La Higuera profile. All sites carry a stable normal magnetization direction. Symbols and conventions as in fig. 3.1.

carried by magnetite with a small hematite component above 580°C. Blocking temperatures in three specimens from site 35 are dominantly below 600°C, suggesting that magnetite is probably the carrier of the stable remanence.

At one site (33), severe chemical alteration and partial destruction of the specimens occurred during the heating experiments at 300 to 450°C.

Quebrada de la Angostura (LA) section

This section covers two formational units, the Tobas Angostura and Río de los Pozos formations (see location in fig. 1.4c). The location in the field of the unconformity that separates La Pilona Formation from Tobas Angostura Formation is based on field observations and information available in published and unpublished maps (e.g., Yrigoyen, 1993 a and b). The stratigraphy, location of the paleomagnetic sites and the paleomagnetic data from the Quebrada de la Angostura section are summarized in figure 3.23.

A general reversal pattern was obtained for this section by sampling at eight sites, six within the Río de los Pozos unit. The stratigraphic separation between sampling locations was usually about 30 m although there is one gap of 48 m.

Of a total of eight sites, six (75%) qualify as Class I; one (12.5%) as Class II and one (12.5%) as Class III. Statistical significance was not obtained in the Class II site because two out of the three samples from the site have nearly parallel magnetic vectors whereas a third is divergent.

The paleomagnetic data yield the following results. The lowest 133 m of the section comprising the Tobas Angostura unit and the lower part of Río de los Pozos Formation

Fig. 3.23. Magnetic polarity stratigraphy of the 'Quebrada de La Angostura' section. a) Simplified stratigraphic section, showing formations, basic textures, unconformities (U), dated tuff horizons (star with sample number). Below the unconformity U₁, there are strata of the La Piona Formation which were not sampled in this profile. b) VGP latitudes. Plotted points are from class I sites (solid circles) and class II sites (open circles). c) Magnetic polarity zonation derived from the VGP latitude plot (b). Black zones have normal polarity, and white zones have reversed polarity. Ⓢ represents a Class III site interpreted as recording a magnetic transition.

records a normal magnetozone (N_1). Above it, from 133 to 232 m the sedimentary rocks are reversely magnetized (R_1). The Tobas Angostura Formation contains several ash beds, one of which is located at 9 m and has been dated by the ^{40}Ar - ^{39}Ar method.

Samples collected from Tobas Angostura Formation in the LA profile show a stable primary remanence. Specimens from the two sites carry magnetizations directed north and up (fig. 3.24b).

Directions in both samples remain stable in the temperature range from ~ 20 - 100° to ~ 580 - 640°C (fig. 3.24b). Blocking temperatures in three specimens from site 01 are mainly below 600°C , suggesting that magnetite is probably the carrier of the stable remanence. In site 02, blocking temperatures indicate that hematite is the carrier of the remanence.

Samples collected from Río Los Pozos Formation carry a very stable and consistent primary magnetization, directed either north and up or south and down. Typical examples of paleomagnetic results for individual specimens are shown in figure 3.24a.

Directions at most sites remain stable in the temperature range between 200 and 640°C . Most samples attain the stable magnetization direction after removal of a soft viscous component which is eliminated in temperature less than 200°C . Blocking temperatures obtained in thermal cleaning experiments indicate that in most sites the stable remanence is carried by hematite.

Thermal demagnetization results from three specimens of site 03 show magnetization directions that are too scattered to allow calculation of a meaningful mean direction. As this site is located between sites of normal polarity below and reversed polarity above, it is interpreted as a transitional site.

Fig. 3.24. Typical paleomagnetic results for individual specimens from Río de los Pozos Formation (a) and Tobas La Angostura Formation (b) in the Quebrada de la Angostura profile. Symbols and conventions as in fig. 3.1.

El Molino (EM) section

In the EM profile samples from two sites were collected for paleomagnetic study from the Tobas Angostura Formation and two from the Río de los Pozos Formation (for location of the section, see fig. 1.4a). The aim was to establish the polarity pattern of the latter unit as it could not be sampled in the La Higuera profile due to lack of exposure. The stratigraphic separation between sampling locations was between 15 and 25 m.

The stratigraphy, location of the paleomagnetic sites and the paleomagnetic data from the El Molino section are summarized in figure 3.25. The Tobas Angostura unit lies unconformably above strata of the La Pilona Formation. Because about 10-15 m of strata at the base of the Tobas Angostura Formation is covered, no paleomagnetic data were obtained. The boundary between Tobas Angostura and the overlying Río de los Pozos Formation is gradual. It is defined according to convention by the first appearance of medium- to coarse-grained conglomeratic beds.

Of the four sampling sites, three qualify as Class I and one as Class II ($N = 2$). Paleomagnetic data from specimens of all the sites yield a normal polarity magnetozone. The basal part of this section contains an air-fall ash bed that has been dated by the ^{40}Ar - ^{39}Ar method.

Specimens collected from the Tobas Angostura Formation carry a very stable primary magnetization directed north and up (fig. 3.26b). They remain stable in the temperature range between 20 and 660°C, indicating that the stable remanence is carried by hematite.

Specimens collected from the Río de los Pozos Formation also carry a very stable primary remanence directed north and up (fig. 3.26a). Directions at these sites remain stable

Fig. 3.25. Magnetic polarity stratigraphy of the 'El Molino' section. a) Simplified stratigraphic section, showing formations, basic textures, dated tuff horizon (star with sample number). b) VGP latitudes. Plotted points are from class I sites (solid circles) and class II sites (open circles). c) Magnetic polarity zonation derived from the VGP latitude plot (b). The black zone represents normal polarity. The base of the Tobas Angostura unit is covered and lack data for magnetic polarity zonation. It has been estimated that there is about 15 m of no stratigraphic coverage, below which there are strata from the La Pilona Formation. As the boundary between Tobas Angostura unit and the overlying Río de los Pozos Formation is gradual, it is defined by convention, with the first appearance of medium conglomerates.

Fig. 3.26. Typical paleomagnetic results for individual specimens from Río de los Pozos Formation (a) and Tobas La Angostura Formation (b) in the El Molino profile. Symbols and conventions as in fig. 3.1.

in the temperature range between 20 and $\sim 640^{\circ}\text{C}$. Blocking temperatures obtained in thermal cleaning experiments indicate that the stable remanence is probably carried by both magnetite and hematite.

(c) Uncertainties

Magnetic polarity stratigraphy has been widely used as a chronostratigraphic tool to assess geological time in nonmarine deposits after the definitive contribution of Johnson et al. (1975). This technique is based on a comparison of the magnetic history of a local stratigraphic section and the global magnetic polarity time scale (MPTS) based on the marine magnetic record. If a correlation between the two is achieved, then the chronologic data from the MPTS can be applied to the local stratigraphic section.

Uncertainties in magnetostratigraphic correlations are related to the accuracy of the MPTS itself, the sampling strategy used to compile the local magnetostratigraphy and the isotopic dates obtained from the sampling section. A brief analysis of the uncertainties is presented here.

Magnetic polarity time scale

Several magnetic polarity time scales have been published in recent years (e.g., Berggren et al., 1985; Harland et al., 1982, 1990; Haq et al., 1987; Cande and Kent, 1992; Baksi, 1993). They show considerable variation in the length of the normal and reversed polarity intervals and in the ages that are assigned to these intervals. For example, the age of

the older boundary of anomaly 5 has been estimated in a number of studies in recent years, including the following: Saemundsson et al. (1980), ~10.47 Ma; McDougall et al. (1984), ~11.1 Ma; Tauxe et al. (1985), ~10.0 Ma; Deino et al. (1990), 10.30 Ma; and Berggren et al. (1985), ~10.42 Ma. Differences in the various MPTS's result from the variety of methods used in the construction of the time scales and in age calibration. In particular, there are uncertainties in estimating the widths of the sea floor magnetic anomalies on which the time scale is based, uncertainties in the isotopic dates of the sea floor calibration points and uncertainties in the derived spreading rates. In addition, the data base from which any version of the magnetic polarity time scale has been compiled is incomplete, lacking the very short polarity intervals which cannot be detected in the sea floor record (cf. Cande and Kent, 1992). In some cases, local magnetostratigraphy may uncover short polarity zones that are not defined in the established MPTS. An example of such short polarity zones is found in the Miocene sections outcropping in the Bermejo basin, Argentina (Beer, 1990).

Cande and Kent (1992) have recently reassessed the relative widths of the magnetic polarity intervals from magnetic profiles of the world's ocean basins for the Late Cretaceous and Cenozoic. They derived a composite geomagnetic polarity sequence based primarily on data from the South Atlantic. They assumed that spreading rates in the South Atlantic were smoothly varying but not necessarily constant. The ages of magnetic anomalies were determined using a cubic spline function to fit a set of nine age calibration points dated by the K-Ar method and the present-day mid-ocean ridge axis to the composite polarity sequence.

Cande and Kent's (1992) two youngest tie-points are at 2.60 Ma, based on astrochronologic calibration of the Matuyama/Gauss boundary, and 14.8 Ma, based on K-Ar

dates from rocks intruding strata containing a specific genus of Foraminifera. The ages of the older and younger limits of chron C5n (of particular interest in this study) are obtained from interpolation between these two calibration points. As chron C5n is located halfway between them, the uncertainties in the determination of its age limits are rather large (at least ± 0.5 Ma), even without considering the uncertainty in the age determination of the tie-points themselves.

The MPTS proposed by Baksi (1993) for the interval 0-17 Ma is used as a time frame in the current study because it is based on ^{40}Ar - ^{39}Ar dating, rather than the less reliable K-Ar technique used in earlier scales (fig. 3.27). In addition, Baksi et al. (1993) have dated chron C5n, which has been clearly identified in the stratigraphic section of the present study. Baksi's time scale was constructed using the same intervals of normal polarity magnetization as did Cande and Kent (1992), but the ages of the polarity chrons in the interval 10-17 Ma are considerably older than those of Cande and Kent (fig. 3.27b). The discrepancy stems in part from the choice of different tie-points in the middle Miocene.

Baksi's (1993) MPTS was constructed using four tie-points at ~1, 2, 10 and 16 Ma derived from ^{40}Ar - ^{39}Ar plateau ages. These ages are in agreement with values derived by the astrochronological technique (Shackleton et al., 1990; Hilgen, 1991) for chrons in the time frame 0-6 Ma. The ^{40}Ar - ^{39}Ar incremental heating age determined at the 9.67 Ma tie-point agrees with Cande and Kent's (1992) interpolated age of 9.59 Ma. With increasing age, however, the two time scales become increasingly discordant (fig. 3.27b).

Local magnetostratigraphic section

Johnson and McGee (1983) determined statistically that time-uniform sampling is the

Fig. 3.27. (a) Magnetic polarity time scale (MPTS) proposed by Baksi (1993) for the interval 0-17 Ma, which is used as a time frame in the current study (ages for normal polarity intervals given in table at end of caption). (b) Comparison of the MPTS proposed by Baksi (1993) and Cande and Kent (1992). Detailed nomenclature of chrons and polarity intervals are based on Cande and Kent (1992).

Ages for normal polarity intervals in the time frame 0-17 Ma,
as proposed by Baksi (1993)

Polarity Chron	Normal Polarity interval (Ma)	Polarity Chron	Normal Polarity interval (Ma)
C1n	0.00-0.78	C4An	8.59-8.92
C1r.1n	0.99-1.05	C4Ar.1n	9.13-9.21
C2n	1.78-2.02	C4Ar.2n	9.50-9.57
C2r.1n	2.24-2.27	C5n.1n	9.67-9.83
C2An.1n	2.64-3.10	C5n.2n	9.87-11.08
C2An.2n	3.17-3.27	C5r.1n	11.21-11.27
C2An.3n	3.38-3.61	C5r.2n	11.73-11.80
C3n.1n	4.12-4.23	C5An.1n	12.30-12.48
C3n.2n	4.37-4.55	C5An.2n	12.61-12.88
C3n.3n	4.74-4.82	C5Ar.1n	13.22-13.26
C3n.4n	4.94-5.17	C5Ar.2n	13.35-13.41
C3An.1n	5.82-6.06	C5AAn	13.62-13.82
C3An.2n	6.19-6.49	C5ABn	14.02-14.28
C3Bn	6.86-7.01	C5ACn	14.53-15.00
C3Br.1n	7.05-7.09	C5ADn	15.12-15.66
C3Br.2n	7.25-7.28	C5Bn.1n	15.89-15.99
C4n.1n	7.33-7.46	C5Bn.2n	16.17-16.32
C4n.2n	7.54-7.96	C5Cn.1n	17.33-17.65
C4r.1n	8.11-8.14	C5Cn.2n	17.70-17.88

Magnetic Polarity
Time Scale
Baksi, 1993

most efficient method of capturing all the principal reversals in a given interval. Unfortunately, sampling at increments of equal stratigraphic thickness will only yield data at equal time intervals if sedimentation rates are constant. However, sedimentary accumulation rates may vary with time. There may also be time gaps in the sedimentary record caused by periods of non-deposition or erosion (cf. Sadler, 1981; Tipper, 1983). The lack of suitable rocks for paleomagnetic study may also affect the time increment between sampling sites. Therefore, in any given magnetic polarity stratigraphy some polarity intervals may be artificially lengthened or shortened, whereas others may be missing entirely.

In the compilation of a local magnetic polarity stratigraphy, three assumptions must be made because of the information gap between consecutive sampling sites. First, it is assumed that magnetic transitions occur mid-way between consecutive normal and reversed data points. Second, it is assumed that the interval between adjacent, like-polarity sites is occupied by the same polarity. Third, between two consecutive sites of different polarity, it is assumed that there is only one magnetic reversal. These assumptions introduce uncertainties not only in the length of the polarity intervals but also in the actual number of reversals encountered (i.e., some short polarity intervals may be missed between consecutive data points).

The age uncertainty in the location of the boundary between zones of different polarity is dependant on the sediment accumulation rate and the stratigraphic separation between sampling sites. An increase in the stratigraphic separation between sites or a decrease in sedimentation rate results in a higher degree of age uncertainty. Its magnitude can be roughly estimated by multiplying the thickness of a particular sampling interval by the inverse of the

sedimentation rate inferred for this part of the section.

Isotopic dates

Isotopic ages are helpful in correlating local magnetic polarity stratigraphy with the established MPTS. Ages determined by the ^{40}Ar - ^{39}Ar step-heating method are considered to be the most accurate; ^{40}Ar - ^{39}Ar total fusion ages, while less accurate, provide important constraints on the age of the stratigraphic sections.

Correlation between local magnetostratigraphy and the MPTS

There are distinctive signatures in the marine magnetic record, such as the long normal polarity magnetozones (chron C5n). When local magnetostratigraphy captures such an interval, a direct correlation can be made. When no distinctive polarity magnetozones are identified in the local section, the correlation with the MPTS can only be achieved with the assistance of isotopic ages.

If a correlation between the local section and the MPTS is achieved, each magnetic reversal boundary serves as a unique time marker (Johnson et al., 1988). Only the magnetic reversal events that have been uniquely identified in the MPTS have designated ages, and the assigned ages are only as accurate as the calibration of the time scale itself (Johnson et al., 1988). If the same magnetic reversal can be traced between two nearby stratigraphic sections, then the sections can be correlated.

(d) Chronology of the Neogene units

This section describes the correlation of the local magnetic polarity stratigraphy of the present study with the magnetic polarity time scale (Baksi, 1993; fig. 3.28). The nomenclature of the polarity chrons described below and in figure 3.27 is based on Cande and Kent (1992) and Baksi (1993).

Tosca de Mariño section

The magnetic polarity stratigraphy of the Tosca de Mariño section can be most easily matched with the MPTS between 15.8 and 12.2 Ma (from chrons C5ADr to C5r.3). The dominantly normal interval which occurs in the marine record between about 15.7 and 14.5 Ma, broken by a short reversed interval at 15.1-15.0 Ma, is observed near the base of the sampled section. Although the Tosca de Mariño section lacks isotopic ages, the sample Md-12 dated at 12.03 ± 0.45 Ma from a tuff horizon collected from the upper portion of the Mariño Formation in the nearby Arroyo del Agua Blanca section, confirms the correlation.

Arroyo del Agua Blanca section

This section, which consists of strata of the Mariño and La Pilona formations is notable for two reasons: (a) it captures a long normal polarity magnetozone, and (b) it contains two isotopically dated ash horizons. The ash horizons yield ages of 9.63 ± 0.22 Ma (Md-13) and 12.03 ± 0.45 Ma (Md-12). The former is located at the top of the section in strata of the La Pilona unit, and the latter in sedimentary rocks of the Mariño Formation.

Fig. 3.28. Correlation of the local magnetostratigraphic sections of the Neogene strata of the Mendoza Precordillera to the magnetic polarity time scale (Baksi, 1993). Strata from the fault block (FB) in the lower part of the Arroyo del Agua Blanca section may be placed at similar stratigraphic level as strata above fault F2, based on lithostratigraphic correlation. The ^{40}Ar - ^{39}Ar isotopic dates shown with an asterisk represent the best estimate of the age of the sample (for discussion of ^{40}Ar - ^{39}Ar age assignments, see Chapter 2).

⁴⁰Ar-³⁹Ar isotopic dates

⑫	12.03 ± 0.45 Ma (Hb)*
⑬	9.63 ± 0.22 Ma (Hb)*
⑥	10.42 ± 0.20 Ma (Hb)*
	9.75 ± 0.50 Ma (Hb)
	10.75 ± 0.29 Ma (Bt)
⑩	9.15 ± 0.42 Ma (Hb)*
	8.38 ± 0.18 Ma (Bt)
⑤	8.64 ± 0.17 Ma (Hb)*
	8.85 ± 0.37 Ma (Bt)
⑮	9.54 ± 0.15 Ma (Hb)*
	8.48 ± 0.19 Ma (Bt)

Based on these age constraints and the configuration of the local magnetostratigraphy, the long normal polarity zone is readily correlated with the long normal chron C5n (anomaly 5) of the MPTS.

The dated Md-13 horizon falls within a normally magnetized interval that is stratigraphically above the long normal magnetozone. The dated interval is correlated with chron C4Ar.2a, which spans the interval 9.57-9.50 Ma (Baksi, 1993). In this interpretation the tiny reversed polarity zone (C5n.1r) in the long normal chron C5n of the MPTS has not been sampled in the local section. This is not surprising when the stratigraphic separation of ~35 m between sampling sites in this part of the section is considered.

Several lines of evidence suggest that the short section of the Mariño Formation at the base of the Arroyo del Agua Blanca profile brackets the interval ~12.7-12.2 Ma. The isotopic age constraint provided by sample Md-12 at 12.03 ± 0.45 Ma and a comparison between the local magnetostratigraphy and the MPTS suggest that the two normal magnetozones found in the section likely correspond with the upper part of chron C5An.2 and chron C5An.1. In addition, structural and stratigraphic arguments favour this correlation. Most of the stratigraphic column in the nearby Tosca de Mariño section is comprised of strata of the Mariño Formation which are stratigraphically below the strata outcropping in the Agua Blanca valley. Repetition of strata of the upper part of the Mariño Formation is assured by the reversed fault encountered in the Tosca de Mariño section.

La Higuera section

This section, which consists of strata of the Mariño, La Pilona and Tobas Angostura

formations is notable for two reasons: (a) it contains a long normal polarity magnetozone, and (b) it includes two isotopically dated ash horizons. One ash horizon is located at the bottom of Tobas Angostura Formation and is dated at 9.15 ± 0.42 Ma (Md-10). The other is located within ~ 1.5 m of the younger limit of the long normal magnetozone and yields an age of 10.42 ± 0.20 Ma (Md-6).

The long normal polarity magnetozone in the La Pilona Formation is assigned to chron C5n (11.08-9.67 Ma) based on (a) its distinctive length and (b) the age constraints provided by the isotopic ages. It should be noted that although the isotopic ages clearly indicate that the long normal interval is equivalent to chron C5n, the Md-6 isotopic age is a little older than the age assigned by the marine magnetic record. This topic will be discussed below.

The remainder of the La Pilona Formation stratigraphically above the long normal interval is assigned to reversed chron C4Ar.2n. The age assignment of the normal magnetozone within the long reversed interval is ambiguous as the MPTS shows two short polarity intervals punctuating the long reversed interval. Tentatively, this normal polarity zone is assigned to chron C4Ar.1n based on an anomalous direction of magnetization found in one site between the long normal magnetozone and this short normal interval. This anomalous direction suggests that a reversal in the Earth's magnetic field may have occurred in this part of the section.

There are no isotopic ages to constrain the age of the Mariño strata below the unconformity that separates this unit from La Pilona Formation. Tentatively, the normal polarity magnetozone recorded in these strata is correlated with chron C5An.1 based on a lithostratigraphic correlation of this part of the section with the exposures at the Arroyo del

Agua Blanca section.

The normal polarity magnetozone located above the unconformity that separates Tobas Angostura Formation from La Pilona unit is assigned to chron C4An. This correlation is based on (a) an isotopic age of 9.15 ± 0.42 Ma (Md-10) obtained from an ash horizon located in strata of Tobas Angostura and (b) the fact that chron C4An is the first relatively longer normal interval above the long reversed chron C4Ar in the MPTS.

When discussing the age assignment of the long normal polarity magnetozone to chron C5n, it was noted that the isotopic age of sample Md-6 is in disagreement with the age assignment provided by the MPTS. The younger limit of chron C5n in the MPTS is 9.67 ± 0.22 Ma (2σ) based on ^{40}Ar - ^{39}Ar plateau ages obtained from three lava flows in New Zealand (Baksi et al., 1993; Baksi, 1993). The younger limit of chron C5n from ^{40}Ar - ^{39}Ar dating of the local stratigraphy of the present study is 10.42 ± 0.20 Ma. Even when the uncertainties in the two ages are taken into account, a discrepancy of about 0.3 Ma remains.

The reason for the discrepancy is not clear. Baksi et al. (1993) dated samples from the transition zone itself. Because reversals of the magnetic field are likely to take no more than a few thousand years (Harland et al., 1990), little additional uncertainty can be assigned due to their stratigraphic position. Uncertainty in the location of the boundary of the normal magnetozone in the local section accounts for an age uncertainty of perhaps 0.1-0.2 Ma.

It is also possible that the uncertainty assigned to one or both of the isotopic ages is too low. In the case of sample Md-6, hornblende has been dated using both the furnace and laser step-heating ^{40}Ar - ^{39}Ar techniques. The former yields an apparent age of 9.79 ± 0.75 Ma corresponding with the step where the bulk of the gas ($>80\%$) was released, and the latter

yields an isochron age of 10.42 ± 0.20 Ma. The choice of the latter estimate as representative of the age of the sample was based on the quality of the data and the age constraint provided by total fusion analyses on biotite. However, with three discrepant isotopic dates from the same stratigraphic horizon, the final age assignment is determined in part by subjective judgment of whether or not one isotopic date is more reliable than the other.

El Molino and Quebrada de la Angostura sections

These two profiles contain strata of the same units, the Tobas Angostura Formation and the lower part of the Río de los Pozos Formation. The normal polarity magnetozone encountered in both sections is assigned to chron C4An, and the reversed polarity zone recorded in the Quebrada de la Angostura section is correlated with chron C4r. The grounds for this interpretation are (a) the ^{40}Ar - ^{39}Ar isotopic age results from both sections, (b) the isotopic age obtained from strata placed at equivalent stratigraphic position in the La Higuera section, and (c) the consistent normal polarity observed in the Tobas Angostura in all three sections.

In the El Molino section, the isotopically dated ash horizon (Md-5) collected ~15 m from the base of the unit yields an age of 8.64 ± 0.17 Ma, which is in agreement with the age of 9.15 ± 0.42 Ma for sample Md-10 collected in the La Higuera section. The horizon dated in the Quebrada de la Angostura section yields an age of 9.54 ± 0.15 Ma (Md-15), which is ~0.5 Ma older than the isotopic age results from the other sections and the MPTS. In the MPTS the beginning of the normal chron C4An is placed at 8.92 Ma based on interpolation between two calibration points at 1.78 and 9.67 Ma (Baksi, 1993). The discrepancy between

the ages from the MPTS and the Quebrada de la Angostura section may be largely accounted for by uncertainties in the isotopic age of sample Md-15 and in the isotopic ages of the MPTS tie-points which, when totalled, account for an error of about 0.4 Ma. It is also possible that strata from the Quebrada de la Angostura section record somewhat older stratigraphic levels as this section is located ~25 km south of the other two sections.

5. CONCLUSIONS

Based on both sources of chronological data, the ^{40}Ar - ^{39}Ar geochronometry and the paleomagnetic study carried out on the Neogene nonmarine strata of the Mendoza Precordillera, the following conclusions are made:

- (1) Deposition of the Mariño Formation started at ~15.7 Ma and continued to ~12.2 Ma.
- (2) In the La Higuera section, deposition of the La Pilona Formation started at or before ~11.7 Ma and continued to approximately 9.0 Ma. The older age limit is derived by interpolation from the plot of stratigraphic position versus age (fig. 4.8a in Chapter 4) assuming that sediment accumulation rate is constant for the lower portion of the La Pilona unit.
- (3) In the Arroyo del Agua Blanca section, deposition of the La Pilona Formation started no earlier than 11.1 Ma.

(4) Deposition of the Tobas Angostura unit occurred between ~8.9 and ~8.7 Ma.

(5) Deposition of the Río de los Pozos Formation extended from ~8.7 to at least ~8.3 Ma at the Quebrada de la Angostura locality. A few kilometres north of this locality, the Río de los Pozos Formation is approximately 400 m thick. If the sedimentation rate was the same as that estimated in the study area (0.41 mm/a; see fig. 4.9 in Chapter 4) and if it remained constant throughout the 400 m of Río de los Pozos unit, deposition of this unit could be as young as ~7.2 Ma.

(6) The unconformity that separates strata of the Mariño Formation from the La Pilona unit appears to be time transgressive. The lowest strata of the La Pilona Formation are ~0.6 Ma older in the La Higuera section (11.7 Ma) than at the Arroyo del Agua Blanca section (11.1 Ma). The unconformity likely represents a time interval of ~0.6 Ma in the La Higuera section and ~ 1.1 Ma in the Arroyo del Agua Blanca and Tosca de Mariño sections.

(7) The unconformity that separates strata of the La Pilona Formation from Tobas Angostura Formation probably represents a short time interval of the order of 10^4 - 10^5 years.

CHAPTER 4

FORELAND BASIN STRATA

1. INTRODUCTION

This chapter focuses on the physical description of the Neogene nonmarine deposits that crop out at the southernmost extension of the Precordillera thrust belt, in northern Mendoza (fig. 1.1). Clastic strata are ~4000 m thick and consist of five Miocene and Pliocene formational units, which represent the Tertiary molasse deposits derived from uplift and deformation of the Principal and Frontal cordilleras. This succession was deformed in the Late Pliocene and is exposed in a series of north-trending anticlines and synclines that extends beyond the limits of the study area (fig. 1.5).

The Neogene succession comprises a variety of fluvial, alluvial and eolian deposits. This chapter reports on the sedimentary facies and the environment of deposition of these strata. Precise geochronology based on the magnetostratigraphy and ^{40}Ar - ^{39}Ar geochronometry (Chapters 2 and 3) allows determination of the timing of changes in depositional environment and sediment accumulation rate. In addition, a petrographic study establishes sediment provenance. The chapter concludes with the integration of a multiple data set in which sedimentation rates, facies variation and timing of changes of sediment provenance in the foreland are linked to episodes of thrust activity in the Principal and Frontal cordilleras.

This methodology has been previously applied in other foreland basins of the Andean chain to document the history of thrusting in the adjacent thrust belts. Those studies began with Johnsson et al. (1984) and Johnson et al. (1986) in the Sierra de Huaco, San Juan, and subsequent work took place there and in other areas (e.g., Bercowski et al., 1986; Reynolds et al., 1987, 1990; Jordan et al., 1988, 1990, 1993 a and b, 1996; Beer and Jordan, 1989). In particular, the paper of Jordan et al. (1988) provides a complete overview of the approach, including how to interpret the data and limitations of the method.

2. LITHOFACIES AND DEPOSITIONAL ENVIRONMENT

Six Tertiary formations are recognized in the La Pilona area (fig. 1.4a); their most distinctive features and ages are summarized in Chapter 1 (section 2d). All but the oldest unit, the Divisadero Largo Formation, are Neogene in age. The Tertiary succession is bounded above and below by two major stratigraphic unconformities. The basal unconformity is clearly visible along the western and southern flank of the Cacheuta peak, north and south of the Mendoza River. Near the Arroyo de las Minas dry riverbed, the Divisadero Largo Formation lies unconformably on Triassic strata that onlap andesitic rocks of the Choiyoi Group at an angle of about 15° (fig. 1.4a). The upper unconformity is well exposed in the eastern part of the study area where conglomerates of the youngest unit, the Mogotes Formation, are overlain by unconsolidated Quaternary sediments that in some areas contain intercalated air-fall deposits.

Within the Tertiary succession, there is an unconformity that separates strata of the

Upper Eocene - Lower Oligocene Divisadero Largo Formation (e.g., Pascual and Bondesio, 1981) from younger units. This unconformity is well exposed in a few localities, such as near the Arroyo de las Minas riverbed at the locality of the Tosca de Mariño section or north of the Mendoza River to the west of the Cacheuta peak. At these localities, the Divisadero Largo Formation is overlain by the conglomerate of the lower member of the Mariño Formation (plates 4.1 and 4.2b). The stratigraphy of the Neogene succession exhibits regional unconformities that separate formational units. These bounding surfaces are paraconformities that mark a hiatus ranging from $\sim 10^5$ to $\sim 4 \times 10^6$ Ma.

Formational units are described below in ascending stratigraphic order and follow the nomenclature recommended by Yrigoyen (1993a). Although the Neogene units are of primary interest in this study, a brief description of the Divisadero Largo Formation is included because it represents the base of the Tertiary succession in the study area.

Interpretation of lithofacies and environment of deposition of each unit is based upon textures, sedimentary structures and architectural styles. Sections were selected to maximize vertical stratigraphic continuity, a prerequisite for magnetostratigraphic studies, rather than lateral continuity or three-dimensional exposure, which are more convenient for facies analysis. Therefore, the limited data set allows for recognition of the most distinctive lithofacies and a general interpretation of the environment of deposition for each of the study units. In some cases, sedimentological data from other exposures were collected to add information on the study sections.

Plate 4.1. (a) Photograph looking to the southwest to the Mariño Formation showing the three members of this unit: L, lower member, Purple conglomerates (Conglomerados Violáceos); M, middle member, Cross-bedded sandstones (Areniscas Entrecruzadas); and U, upper member, Mariño strata (Estratos de Mariño). The photo is taken from a road that leads south from the international route (connecting Chile and Argentina) and runs parallel to the Arroyo de las Minas dry riverbed. DL = Divisadero Largo Formation. (b) Closer view of the lower member (above the line) and the Divisadero Largo Formation (below the line). Note that the lower member comprises two distinctively coloured packages: PC, the lower unit, from which this member derives its name (Purple conglomerates); and DS, the upper unit, termed the Druse strata (Serie de las Drusas; Truempy and Lhez, 1937), consisting of gray tuffaceous sandstones alternating with reddish mudstones.

a

b

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Plate 4.2. (a) View of the Divisadero Largo Formation. Gypsum horizons that form continuous or discontinuous masses of coalesced nodules traceable for several meters are interbedded with varicoloured mudstones. (b) The rock outcrop represents the base of the Mariño Formation and is composed of fine purple conglomerate and sandstone which overlie the red mudstone of the Divisadero Largo Formation. Note that these deposits are characterized by almost flat, erosive bases, occurrence of internal scour surfaces (e.g. see arrows) and they taper into mudstone (toward the upper left). Just below the contact between the two formations are horizons of nodular chert (broken line of white fragments), along which there is evidence of bedding-shear displacements.

a

b

(a) Divisadero Largo Formation

This unit reaches a thickness of up to ~80 m in the study area (e.g., near the town of Potrerillos) but is locally ~35 m thick in the Tosca de Mariño section (fig. 4.1, blow up #1). At this locality, three main lithofacies are recognized:

Lithofacies 1: Red sandstone and pebbly sandstone with nodular gypsum

This lithofacies characterizes the lower 18 m of the unit. It consists of medium sandstone and pebbly sandstone which typically contain lenses or horizons of nodular gypsum, locally forming meshes and networks with linking veins. Bedding surfaces are difficult to identify. Where observed, bed thicknesses range from ~0.5 to ~1.2 m, and they have planar bases, sometimes erosive with low-relief downcutting. Beds are massive but locally fine upward. The coarsest material, found at the base of the unit and concentrated in scour fills, consists of pebbly sandstone or locally a conglomerate with abundant angular volcanic pebble clasts ranging in diameter from 0.5 to 2 cm, most commonly <1 cm.

Lithofacies 2: Varicoloured mudstone with gypsum horizons

It consists of laminated tuffaceous siltstone and claystone and thin siltstone to fine sandstone interbeds. Composite bed sets show flat-based sandstone beds with flat or wavy tops. Layers rich in organic matter are locally found in alternation with wavy laminated carbonate-rich layers. Distinctively, there are ~2-15 cm thick gypsum horizons which occur parallel to bedding. They form continuous or discontinuous masses of coalesced nodules that

Fig. 4.1. Tosca de Mariño (a) and Arroyo del Agua Blanca (b) profiles, showing details of textures and sedimentary structures of parts of the Mariño Formation (1-12 = detailed local sedimentological profiles). Based on lithologies and the remanent magnetization carried by strata of the Arroyo del Agua Blanca profile, the stratigraphic repetition above faults 1 and 2 is likely less than a few tens of metres. The tentative correlation is given in figure 3.28 (chapter 3). Strata below 0 stratigraphic level correspond to the Divisadero Largo Formation.

can be traced for several metres (plate 4.2a). This lithofacies overlies lithofacies 1 and alternates in gradational contact with lithofacies 3.

Lithofacies 3: Thin-bedded greenish and gray mudstone

This lithofacies is mainly comprised of thin-bedded or laminated greenish tuffaceous claystone and siltstone and interbeds of mudstone and lenticular bedded fine sandstone. Locally, there are normally-graded, sharp-based medium to coarse sandstone beds up to ~5 cm thick, which may contain mud intraclasts. Gypsum laminates are common.

Paleoenvironmental characterization

Based on detailed facies analysis of the Divisadero Largo Formation in the Mendoza-Cacheuta-Potrerrillos region, Kokogian et al. (1988) and Kokogian and Mancilla (1989) interpret this unit to be an ancient continental sabkha, where sediments were deposited by ephemeral fluvial systems that flowed in a bajada towards the most depressed zones. These areas were characterized by the occurrence of standing water bodies and local dune fields (Kokogian et al., 1988).

In the study section, the lithofacies defined by sandstones and pebbly sandstones (lithofacies 1) is ascribed to the channel-fill deposits of the fluvial system, and the presence of nodular gypsum suggests the existence of supersaturated phreatic pore fluids. The occurrence of massive beds with no observed upward change to other sedimentary structures and the evidence of scour at the base of beds suggest rapidly decreasing flow power in short-lived, high energy flow events (cf., Picard and High, 1973; Turnbridge, 1981, 1984). The

dominantly muddy sequence indicates deposition under lower energy conditions. Depressed areas or shallow ponds in a mud flat were subjected to periods of dominantly subaqueous conditions where sedimentation was dominated by suspension fallout of mud occasionally interrupted by brief, higher-energy tractional events that deposited normally graded beds and interbeds of siltstone to very fine sandstone (lithofacies 3). Episodes of emergence are recorded by the laminated varicoloured mudstone interbedded with organic-rich layers (algal peats?) and gypsum horizons (lithofacies 2).

(b) Mariño Formation

This unit is subdivided into three members as summarized in Chapter 1, section 2d (plate 4.1a). The contact between members is marked by a general change in exposure characteristics and an easily recognized change of depositional facies. Lithofacies determinations are based on sedimentological data obtained from the (a) Tosca de Mariño and (b) Arroyo del Agua Blanca profiles (fig. 4.1; location of the sections are shown in fig. 1.4a). Vertical stratigraphic continuity of strata is interrupted by a thrust fault (F_1 , fig. 4.1) at the top of the Tosca de Mariño profile. This fault is traced to the base of the Arroyo del Blanca profile where repetition of the uppermost Tosca de Mariño strata occurs above the fault. In the Arroyo del Agua Blanca profile, part of the stratigraphy may also be repeated above fault 2 (F_2). Based on lithofacies and the primary remanent magnetization carried by strata of the Arroyo del Agua Blanca profile, the amount of repetition is likely less than a few tens of metres (fig. 3.28, Chapter 3).

In view of the limited lateral continuity of exposure in some stratigraphic intervals of the lower part of the upper member, additional data from the lowest 300 m of this member were collected from a locality situated ~5 km west of the Camping Santo Domingo, south of the international route that connects the towns of Cacheuta and Potrerillos (for location, see fig. 1.4b).

Lower member (≤ 63 m)

This member comprises two distinctively coloured packages (plate 4.1b). The lower is up to 18 m thick, and the upper is up to 45 m. The lower unit, from which this member derives its name (Conglomerados Violáceos = Purple conglomerates; Truempy and Lhez, 1937), is characterized by vertically stacked beds of purple and greenish conglomerate and sandstone that alternate with thinner intervals of red mudstone. In the study profile (section I), this unit thins laterally along the outcrop (from 18 to 13 m) due to displacement along a left-handed strike-slip fault. The upper unit, termed the 'Serie de las Drusas' (Truempy and Lhez, 1937), consists of gray tuffaceous sandstone beds alternating with red mudstones.

Lithofacies 1: Purple conglomerate and sandstone

This lithofacies comprises conglomerates and sandstones (fine to very coarse grained) arranged in stacked upward fining beds ranging in thickness from a few decimetres to ~2 m (fig. 4.1, blow up #2). They define conglomeratic sandstone bodies a few metres thick that extend throughout the ~250 m wide exposure. Conglomeratic sandstone bodies have almost flat (locally gently concave upward) erosional bases and a few internal erosion surfaces that

parallel the base (plate 4.2b). Although channel-like forms are rarely observed, two exposures allowed recognition of gentle sided channel margins. Units generally fine upward from fine conglomerate to medium sandstone. They include sequences that may show a basal lag of pebble clasts, locally with mud intraclasts, in some cases overlain by poorly defined sets of trough cross-bedded sandstone (facies St and Ss of Miall, 1978). These sequences may be capped by a thin, planar-stratified sandstone or may grade up to a thin interval of reddish siltstone. Locally, there are also thin limonite-rich horizons with gypsum concretions forming meshes and networks with linking veins. These layers only occur a few metres above the contact with the underlying Divisadero Largo Formation.

Conglomerates are clast-supported with a sandstone matrix. They are well sorted and composed mostly of extrabasinal subrounded andesite clasts from the Principal Cordillera (cf. provenance study, section 3). Clasts are up to ~3 cm in diameter but commonly range from 2 to 3 mm. The top of the purple section is marked by a poorly sorted, matrix-supported conglomerate that contains clasts up to 3 cm in diameter dispersed in a tuffaceous fine-grained matrix; its thickness is laterally variable from a few decimetres to ~1 m, and its base is planar (locally erosive?). Cross-beds, although rare due to the lack of suitable exposures and to poorly defined structures, suggest eastward paleoflow ranging from N60° to N150°.

Lithofacies 2: Massive gray tuffaceous sandstone

This lithofacies characterizes the upper 45 m of the member. It consists of sheet-like sandstone bodies ranging in thickness from a few decimetres to up to ~6 m (fig. 4.1, blow up #3), which are characterized by planar nonerosive or locally erosive bases and flat tops.

Sandstone bodies alternate with intervals of reddish mudstone from 10 cm to 3 m thick. Measurement of clast fabric and sandstone bed thickness was hindered by the weathered nature of the exposures. Beds, clearly discernible, are up to ~1.5 m thick and persist laterally over the ~200 m wide exposure.

Sandstones are medium- to fine-grained and tuffaceous. A few horizons consist of a light-coloured bentonite. Internally, sandstone units show subtle planar stratification highlighted by pebble clasts, or they are massive, locally with dispersed clasts of 1-2 cm in diameter, which may be concentrated in small pockets. Mud intraclasts and sparse fine gravels (up to 2 cm) are locally present at the base of some sandstone intervals and these units may show a subtle upward fining. In the uppermost part of this section, decimetre-thick massive beds bounded by thin mudstone layers are mottled, with patches of colour occurring on the bed surface. Bioturbation is suggested by the occurrence of discontinuous lumps or pockets with different texture (e.g., grain size) within beds. Gypsum is relatively abundant and is found disseminated on the surface of sandstones or as thin veins.

Lithofacies 3: Red mudstone

These deposits occur between purple sandstone bodies or alternating with flat-based gray sandstone units. Where interbedded with purple sandstone, the muddy facies consists mainly of red siltstone, which may grade upward into a massive or laminated claystone. These deposits define laterally extensive packages with thicknesses of usually up to ~2 m.

Fine-grained intervals between flat-based sandstone bodies consist of laminated or massive mudstone or a heterolithic association of thin massive or parallel-laminated sandstone

beds interbedded with mudstone, which may include flat-based sandstone units a few decimetres thick. Distinctively, there are several laterally persistent ~10 cm thick horizons of carbonate-rich mudstone whose highly indurated nature contrasts with the rest of the weathered exposures. These beds show rare sand-filled cracks. In addition, siliceous (chalcedony) concretions do occur (Truempy and Lhez, 1937).

Interpretation

The lithofacies recognized in this member indicate that these deposits are of fluvial origin. The lower ~18 m comprised of purple conglomerate and sandstone bodies (lithofacies 1) with alternating red mudstone (lithofacies 3) is interpreted to represent channel-fill and flood-plain deposits. The architectural style of the conglomerate and sandstone bodies (e.g., sheet-like geometry, basal and internal erosive surfaces) in conjunction with the internal characteristics of these deposits (e.g., normally graded beds, fining upward sedimentation units, absence of fine sediment within sandstone units, lack of evidence of lateral accretion surfaces) suggest that deposition in channels was dominated by vertical rather than lateral accretion. Deposition likely occurred in channelized but poorly confined parts of a stream system, as suggested by the rarity of clearly defined channel margins (cf., Friend, 1983). Sequences with a pebble lag followed by poorly defined sets of cross-beds suggest deposition by waning flows with an initially erosive phase, probably streams with braided morphologies. Rare episodes of sediment mass-flow deposition (debris-flow deposits; cf., Schultz, 1984) could be inferred from the occurrence of a matrix-supported conglomerate with abundant pebble clasts dispersed in a tuffaceous fine-grained matrix.

That the fluvial system probably had an ephemeral regime is suggested by the occurrence of stacked normally graded beds, whose deposition could result from successive high energy flow events. This interpretation gathers support from the occurrence of limonite-rich horizons with gypsum concretions and the presence of well oxidized flood-plain fines, which are suggestive of semiarid conditions where rainfall and therefore water discharge were confined to short-lived episodes. The location of these horizons (a few metres above the gypsum-bearing Divisadero Largo Formation) suggests that surface evaporation was responsible for drawing saline water upward, where the dissolved solids precipitated as a crust.

The weathered nature of the upper ~45 m of this member in conjunction with little remaining evidence of stratification preclude detailed depositional interpretation of lithofacies 2 (Massive gray tuffaceous sandstone). However, the occurrence of laterally persistent, flat-based massive beds of limited thickness suggests that these accumulations could represent unconfined bedload sheet deposits. This interpretation could also account for the notable lack of large-scale channels or surfaces of lateral accretion.

The occurrence of mottled sandstone beds, abundant gypsum and carbonate-rich mudstone horizons with occasional mud cracks suggests semiarid conditions where episodes of sandstone accumulation alternated with periods of mud flood-plain deposition and little or no sedimentation.

Middle member (180 m)

This member, known as 'Areniscas entrecruzadas' (Cross-bedded sandstones),

comprises a monotonous succession of pale brown, cross-bedded sandstone (plate 4.3).

Lithofacies 4: Large-scale cross-bedded sandstones

Sandstones are cross-bedded, with foresets commonly dipping $\sim 20\text{--}30^\circ$ (fig. 4.1, blow up #4). Sets ranging in thickness from a few decimetres to a few metres are tabular, wedge planar (*sensu* McKee and Wier, 1953) or, less commonly, of trough shape. Individual sets are separated by horizontal or slightly inclined surfaces. Foresets show internally different types of small-scale structures, including inversely graded laminae a few millimetres to 1 cm thick, structureless lenses of medium to coarse sandstone near the base of foresets, normal grading of laminae or strata from coarse to fine sandstone, and parallel laminae or strata lacking grading. Low-relief (high-index) ripples occur on exposed foreset surfaces.

Intervals varying in thickness from a few s to usually up to 1 m of horizontal thin-bedded sandstones alternate with cosets of cross-bedded strata. Sandstone beds are parallel-laminated (locally with crude inverse grading), cross-laminated or massive, and beds are separated by continuous or discontinuous mud drapes.

Rarely, at the base of large-scale cross-beds there are local scours that pinch out in a few metres and are filled with sandstone that fines upward showing a lag of sparse clasts up to ~ 2 cm. Scours are sometimes filled with cross-bedded strata.

Sandstones are tuffaceous, fine- to coarse-grained and well sorted, with abundant subrounded clasts of volcanic rocks and quartz.

Interpretation. This member exhibits many sedimentological features that have been

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Plate 4.3. (a) View of the middle member of the Mariño Formation showing large-scale tabular, planar cross-beds overlain by an ~1 m thick interval of horizontal thin-bedded sandstones, interpreted as dune and interdune deposits, respectively. (b) Detail showing large-scale tabular planar cross-strata overlain by wedge planar cross-bedded strata.

a

b

described in deposits of eolian origin. These include moderate to high angle cross-beds forming wedge-planar or tabular-planar sets, the small-scale internal structures of foresets, the occurrence of horizontal thin-bedded sandstone intervals with mud drapes, ripples with high index, and the well sorted texture of the deposits (e.g., Glennie, 1970; Ahlbrandt and Fryberger, 1982; Collinson and Thompson, 1989). Large and medium-scale cross-beds are interpreted to represent dune facies where the varied type of small-scale structures record the growth and migration of these bed forms. Sandstone lenses (sand toes), inversely graded laminae and normally graded or ungraded strata are produced by avalanching, ripple migration and grain-flow processes (Collinson, 1986; Collinson and Thompson, 1989). Interdune facies are interpreted to be recorded by flat-bedded sandstone units and include minor proportions of water-laid sediments as inferred from the occurrence of local scours filled with sandstone and a basal lag of small pebbles.

No systematic measurements of cross-bedding dip directions were collected, thus precluding recognition of the type(s) of dune. However, the recognition of trough-shaped sets in some stratigraphic intervals suggests sinuous-crested transverse dunes which seem to dominate deserts with complete sand cover (Collinson, 1986). Avalanche face orientations, when measured, show dip directions trending towards the east.

The occurrence of this lithofacies stratigraphically above fluvial sequences (deposited in a semiarid setting) suggests that, sequentially through time, reduced fluvial input favoured the development of an extensive eolian dune field.

Upper member (~550 m)

This member, known as 'Estratos de Mariño' (Mariño strata), comprises a succession of red beds that consist of gray and reddish sandstone and conglomerate bodies alternating with red mudstone or a heterolithic association of mudstone and sandstone. One of the most remarkable features of the upper member is the abundance of sandstone bodies that show variable degrees of channelization and are characterized by the diversity rather than the uniformity in architectural style and sedimentary structures. In a few locations, data on sand-body dimension, continuity of mudstone, and relative positioning of the different types of sedimentary bodies were collected. It should be noted that the good quality and availability of laterally extensive exposures would permit a more detailed facies analysis. Deposits (<10 m thick) of a large-scale cross-bedded sandstone with similar characteristics as described above (lithofacies 4) have been recognized in the basal ~200 m of the upper member in other localities than that of the study profile.

Lithofacies 5: Tabular sandstone sheets

Sandstone sheets are the dominant type of sandstone deposits in the upper member (fig. 4.1, blow up #6; plates 4.4 and 4.5). This lithofacies comprises tabular-sheet-like beds or bed sets laterally persistent for several metres or tens of metres. Individual beds range in thickness from a few centimetres to up to ~1 m. Sandstone sheets occur as (1) isolated beds separated by siltstone and sand-streaked siltstone up to ~1 m thick; (2) repeated alternations of very fine to medium sandstone and siltstone in beds which rarely exceed ~40 cm in thickness (usually 5-20 cm) defining sequences of up to ~1.5 m thick; and (3) stacked

Plate 4.4. (a) Tabular sandstone sheets of the upper member of the Mariño Formation. Beds are vertically stacked or separated by thin siltstone layers. (b) Detail showing parallel-laminated (see arrow) or massive sandstone beds.

a

b

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Plate 4.5. (a) Discordant clastic dikes cross-cutting sandstone and mudstone sheets. (b) Visible in the foreground is a channel-fill conglomerate with a lenticular or multi-storey style of internal geometry. Note the presence of cross-cutting internal erosion surfaces. In the distance are sequences of tabular sandstone sheets which are laterally persistent for tens of metres.

a

b

decimetre-scale beds either amalgamated or separated by siltstone partings which define sandstone bodies of variable thicknesses. Stacked beds are locally arranged in thinning- and fining-upward sequences, which may show at their bases transverse scour fills indicating paleoflows towards the south-southeast. Uni- and multi-storey conglomerate channel fills (lithofacies 6) are intercalated within tabular sandstone sheets.

Sandstones are dominantly medium- and fine grained. Sandstone sheets have planar, nonerosive or low-relief erosive bases which may have mud intraclasts and pebbles that locally line the base of the beds. Sandstones are dominantly parallel-laminated, but there are also massive beds sometimes faintly laminated at the top.

In the lower ~300 m of the Mariño Formation, tabular sandstone sheets make successions up to ~5 m thick. In the upper part of this unit, some successions up to ~15 m thick are dominantly composed of tuffaceous sand sheets consisting of vertically stacked normally graded beds displaying horizontal or low-angle lamination in fine- to coarse-grained beds (fig. 4.1, blow up #10 between ~750 and ~766 m). The base of many beds show concentration of small pebbles (1-2 cm in diameter) above slightly irregular basal surfaces. Tops of beds are locally truncated by scoured surfaces up to 0.2 m of relief which define broad lenticular units filled with a parallel-laminated sandstone, locally with cross-beds and a lag of sparse pebbles. The gradual increase up section of broadly channelized sedimentation units composed of thin conglomerate strata in alternation with sandstone is associated with the development of cut-and-fill structures and the transition to multi-storey channelized successions (lithofacies 6) and extensive conglomerate sheets (lithofacies 8).

Sedimentological observations in other localities show that a common feature in these

successions, particularly in those characterized by repeated alternations of sandstone and siltstone beds, is the occurrence of different types of sole marks on the bases of sandstone beds, such as flute and groove marks, longitudinal scours and load casts. Paleoflow directions of these sole marks on beds at the base of the upper member are southeastward (from N135° to N180°).

Discordant clastic dikes commonly a few centimetres wide and up to 1 m long are sporadically found cross-cutting sandstone and mudstone sheets (plate 4.5a). Sand sheets locally display soft deformation convolution and features that resemble dish dewatering structures. Tops of sandstone beds may contain small carbonate pseudonodules forming thin horizons.

Interpretation. These deposits bear similarities to the thick- and thin-bedded sandstone units described by Turnbridge (1981, 1984) and to other ancient fluvial deposits that have been interpreted as of sheet-flood origin (e.g., Olsen, 1988; Fernández, 1993). The tabular geometry of the sandstone sheets, the thickness of sediment deposited for single events, the presence internally of only first- and second-order bounding surfaces (cf. Miall, 1988), the dominance of horizontal lamination, and the absence of lateral accretion bedding strongly suggest deposition by sheet floods. Where lateral control was available, sand sheets could not be traced into channel-sandstone bodies, favouring the interpretation that they represent sheet-flood deposits (cf. Friend, 1983). Sequences with fining-upward arrangement of bed thickness and grain size suggest deposition from the initial stages, throughout the waning phases of the sheet-flood event (Fernández, 1993). Locally, flood events were

characterized by an erosive initial phase, as suggested by the occurrence of scour fills at the bases of these sequences. In the upper part of the Mariño Formation, the thick successions composed of normally graded and laminated sheet-like beds with tops locally truncated by broad lenticular units filled with a cross-bedded or parallel-laminated sandstone suggest that deposition occurred from recurrent, short-lived, unconfined or broadly confined shallow flows which partially scoured pre-existing sand deposits (e.g., Dreyer, 1993).

Modern examples of ephemeral flood sedimentation show the dominance of upper-flow-regime parallel-laminated sands in the sedimentary assemblage (e.g., McKee et al., 1967; Williams, 1971; Stear, 1985). The predominance of parallel-laminated sandstones in conjunction with other sedimentological features (i.e., oxidized flood-plain mudstones, incipient calcrete horizons, gypsum veins in mudstones interbedded with sand sheets; lithofacies 9) are suggestive of arid or semiarid conditions where sporadic and perhaps violent rainfall favoured episodes of upper-flow-regime discharge. The occurrence of massive beds, dewatering structures and quick-sand injection features suggests the recurrence of sand-laden flow episodes and possibly rapid deposition.

Lithofacies 6: Channel-fill deposits

Channel-fill deposits comprise a wide range of lithologies that can be assembled into two main groups: (a) conglomerate and (b) sand-dominated facies. The latter deposits occur as a well-developed lithofacies in localities situated several kilometres to the west of the Tosca de Mariño profile, such as near the Santo Domingo Camping (south of the Mendoza river), where they occur in the lower ~200 m of the upper member in alternation with

mudstone intervals of up to ~8 m thick. At that locality, some of the multi-storey sandstone successions comprise a wide spectrum of grain size that includes coarser-textured deposits with the characteristics described in (a). In the study profile, laterally restricted outcrops preclude determination of the large-scale internal geometry of sandstone bodies.

(a) *Multi-storey channel-fill conglomerates*. These deposits make up minor channel fills at the top of tabular sandstone sheets as well as multi-storey fills of larger channels attaining thicknesses of up to ~5 m (fig. 4.1, blow up #7 and #12). The thicker conglomerate bodies are laterally persistent for many metres and their bases are invariably erosional. Internally, they are characterized by concave-up erosion surfaces which define channel-like forms that cut into each other both laterally and vertically (plate 4.5b). Although the scale of these forms is highly variable, they are up to ~7-8 m wide and ~2-2.5 m thick. Their bases may contain mud rip-up clasts and large pebbles and they are filled with trough-bedded gravels and pebbly sandstones (Gt and St facies; Miall, 1978) and/or sets of horizontally stratified fine conglomerate and pebbly sandstone. Locally, solitary decimetre-scale sets of planar cross-beds (facies Sp; Miall, 1978) occur above scour surfaces. Conglomerates are medium to fine, moderately to poorly sorted, with a sandstone matrix and clasts of varying size usually up to 10-15 cm long, sometimes reaching between 20 and 25 cm.

Minor gravelly channel fills up to ~1 m thick and a few metres wide are locally characterized by the occurrence of lateral accretion bedding with internal gently dipping surfaces normally draped by a thin sandstone or mudstone drape. Two exposures nearly perpendicular to the channel-axis allowed recognition of cross-stratification between accretion planes (epsilon cross-stratification; Allen, 1965). Paleocurrents indicate eastward paleoflows

and channel migration from south to north.

(b) *Multi-storey and single channel-fill sandstones.* Multi-storey and single sheet-shaped sandstone bodies are characterized by flat, locally concave-up basal erosive surfaces which are incised into or are parallel to mudstone sheets. Single sandstones bodies have variable thicknesses up to ~2 m whereas multi-storey successions are up to ~10 m thick. Similar to the multi-storey conglomerate successions, sandstone bodies show internally concave-up erosion surfaces that define lenticular units a few metres or several metres wide. Multi-storey successions also result from the vertical and lateral superposition of sheet-like and lenticular depositional units, the latter locally displaying thin sandstone sheets that extend outwards from a thicker central part. They are a few tens of metres wide. Sheet-like units may be as thin as a few tens of centimetres, in many cases lacking basal scour surfaces.

Typically, cross-cutting erosion surfaces are draped by mudstone layers, and sheet-like storeys in multi-episodic channel-fill sequences are commonly separated by mudstone layers sometimes with dessication cracks. Individual storeys contain numerous internal erosion surfaces and mudstone veneers. These deposits are characterized by (a) the predominance of a horizontally bedded sandstone filling very large erosional troughs (Sh and Sl facies, Miall, 1978; fig. 4.1, blow up #5 between ~262 and 270 m), and (b) the abundance of scour-and-fill structures and large-scale trough cross-bedding, with trough-shaped sets of a pebbly sandstone and sandstone measuring in vertical transverse section from ~0.5 to up to ~3 m in width and between 0.1 and 0.4 m in height (St and Ss facies, Miall, 1978; fig. 4.1, blow up #12). Sandstone bodies locally comprise sheet-like sedimentation units displaying low-angled lateral accretion bedding. This structure is local in character (i.e., channel depth = 1-1.5 m;

lateral extent of accretion bedding = 7-8 m) and also occurs in thin sandstone sheets isolated between shales (i.e., channel depth = 0.5 m; lateral extent of accretion bedding = 4 m) with accreted strata separated by silt drapes or amalgamated. Paleocurrents suggest eastward paleoflows. Uni-storey sandstone bodies or individual storeys in multi-storey bodies may also comprise fining-upward sequences of a trough cross-bedded and plane-bedded sandstone. These sequences may be capped by a rippled interval and/or a thin veneer of mudstone.

Discrete channel-like forms (~1.5 m thick and a few metres wide) may occur isolated in mudstone. They have well-defined margins, cutbanks and a step-like or concave-up basal scoured surface. They show evidence of scouring and filling with a few internal erosion surfaces separating sets of a cross-bedded fine sandstone.

Interpretation. The occurrence of sheet-shaped sandstone and conglomerate deposits with abundant cross-cutting scour surfaces suggest the existence of repeated episodes of scouring, infilling and switching of shallow channels in mobile channel belts. This lenticular or multi-storey style of internal geometry is common in sandstone and conglomerate strata interpreted to be braided-fluvial deposits (e.g., Rust, 1979; Allen, 1983), which probably formed a network of shallow, interconnected braided streams (braid belts, e.g., Galloway, 1981). The occurrence of multi-storey successions displaying great variability in grain size is thought to indicate variations in stream power and spatially or temporally fluctuating discharge. Transport and deposition of gravels in channels was likely associated with episodes of stronger channelized flows. In contrast, the multi-storey sandstone bodies characterized by smooth, horizontal basal surfaces (sixth-order bounding surface; Miall, 1988), only locally

cut by minor channel scours, might record the initiation of the braided stream system by sheet flows, as suggested by the scarcity of scour surfaces (cf. Olsen, 1989). The interpretation that these deposits may record deposition by stream- and sheet-flow processes is supported by the occurrence of ribbon sandstone bodies with lateral sandstone sheets suggesting that channelization was likely accompanied by extensive sheet-flooding outside the channels and by the abundance of a horizontally bedded sandstone (Sh and Sl facies), both of which suggest the common occurrence of high-energy, shallow flows. That the fluvial system had a regime of flash discharge and fluctuating hydrodynamic conditions is inferred from the abundance of numerous internal scour surfaces, mud drapes, dessication cracks and alternations of upper and lower flow regime bed forms (e.g., fining-upward sequences dominated by a large- to medium-scale trough cross-bedded sandstone which record simultaneous erosion and deposition related to the migration of three-dimensional dunes along channels). Channel migration and switching processes likely predominated over systematic channel lateral migration in view of the local character of the lateral accretion bedding. Some of the relatively unusual features recognized in the sandstone bodies (i.e., abundant plane-bedded facies; lack of scour basal surfaces) are described in other ancient braided deposits interpreted as having been produced under ephemeral stream conditions (e.g., Stear, 1985; Beer and Jordan, 1989; Fernández, 1993).

Channel-like forms with well defined cut banks and isolated within-flood-plain mudstones likely represent straight or slightly sinuous solitary-channel stream deposits that were scoured and filled during successive episodes of flow discharge.

Lithofacies 7: Gravelly sandstone bodies with variable degree of channelization

Parallel stratified gravelly sandstone bodies. These deposits have flat nonerosive or erosive basal surfaces, locally broadly concave upward, and attain thicknesses of usually up to ~2 m. They consist of single or stacked centimetre- to decimetre-scale normally graded beds of a conglomerate with pebble clasts and pebbly coarse sandstone (fig. 4.1, blow up #5 between 270-280 m).

Fining upward sequences. Some sandstone bodies in the middle-upper part of the Mariño Formation consist of stacked fining-upward sequences of variable thickness up to ~1.2 m each, which are bounded by scours ranging from a few cm to ~1 m deep (blow up #8 and #9 between ~590-598 m and ~600-637 m, respectively). Scours are filled with cross-beds of a pebbly sandstone or conglomerate with pebble clasts, overlain by a massive to diffusely horizontally bedded coarse to medium sandstone that may contain horizons of aligned small pebbles. The scoured coarser-grained basal part with cross-beds is not always present.

Tabular conglomeratic sandstone bodies. These deposits are laterally persistent for at least a few tens of metres attaining thicknesses of usually up to ~2 m. They occur in sequences of tabular sandstone sheets (fig. 4.1, blow up #8). Tabular conglomeratic sandstone bodies have flat or low-relief scoured bases and flat tops. Their most distinctive features are (a) the high variability along strike in the vertical arrangement of conglomeratic and sandstone intervals which record changes in the degree of channelization, (b) the presence of abundant local scours of variable size, and (c) the diffuse or clearly defined horizontally bedded stratification exhibited by coarse-grained to pebbly sandstones, sometimes highlighted by the local alignment of pebbles up to ~2 cm in diameter. Local scours of variable size (from a few

decimetres to a few metres in width and up to ~0.5 m thick) disrupt flat-lying sandstone strata. The increase in the abundance of conglomeratic relative to sandstone intervals, either laterally or vertically, defines sequences characterized by the lensoid geometry of conglomeratic lithosomes. Conglomerates are massive or have a crude stratification. These deposits are composed of fine to medium gravels with a sandstone matrix, locally with clasts up to 20 cm in diameter, locally imbricated, indicating paleoflow directions towards the south and east. In other cases, medium conglomerates show internal curved erosion surfaces that bound decimetre-scale normally graded beds which alternate with sandstone layers.

Poorly sorted conglomerate pockets and beds. Beds have sharp, locally erosive bases and thicknesses are commonly up to ~20 cm. Pebbles of variable size (up to ~15 cm) are dispersed in a sandstone matrix, or larger clasts are locally in contact with each other. Clast fabric is disorganized, and ungraded and inverse grading is present in the lower centimetres. The top of beds may be capped with a massive or thinly bedded medium sandstone. These deposits are rare and occur mainly in sequences dominated by tabular sandstone sheets, in the lower ~300 m of the upper member.

Interpretation. Most of the deposits described above share two distinctive characteristics: (a) poorly defined channelized basal contacts (cf. Friend, 1983) and (b) dominance of normally graded beds (without displaying transitions to other sedimentary structures upward) and occurrence of horizontally bedded strata. Together, these features suggest the existence of short-lived flow events, either stream or sheet floods, with flows characterized by rapidly declining flow power (Picard and High, 1973; Turnbridge, 1981).

Events of short-lived higher water discharge are probably responsible for the deposition of the horizontally stratified conglomerates. The flat bases of these beds and their centimetre-scale thicknesses suggest unchannelized or shallow and poorly channelized streams (e.g., Nemec and Steel, 1984; Collinson, 1986). The massive to horizontally bedded sandstone deposits are interpreted to represent deposition from both channelized and unconfined high-energy flows, and similar interpretation is suggested for the tabular conglomeratic deposits with abundant metre-scale scours filled with a massive conglomerate (e.g., Dreyer, 1993). The conglomerate sequences characterized by stacked normally graded beds could represent the deposits of successive unconfined or poorly confined flow episodes.

The poorly sorted conglomerate beds and pockets, characterized by disorganized clast fabric and locally inverse grading, are interpreted to represent sediment-gravity-flow deposits related to debris-flow events (e.g., Schultz, 1984).

Lithofacies 8: Conglomerate sheets and bars

These deposits occur in the uppermost part of the Mariño Formation. Conglomerate sheets have low-relief erosive bases and flat tops and extend laterally over the ~10 m wide exposure. Sheets range in thickness from a few decimetres up to ~1.5 m and alternate with sandstone and siltstone intervals defining packages a few metres thick. They consist mainly of a medium-bedded flat-lying conglomerate often highlighted by intimate association with a massive or a well-laminated sandstone. Each sheet includes a few or several conglomerate beds ranging in thickness from a few centimetres to a few decimetres thick. There are also sheet-like bodies of a massive conglomerate with internal, sometimes concave-upward erosion

surfaces separating individual conglomerate units a few decimetres thick. Fine and medium conglomerates have a clast-supported framework and a sandstone matrix, and locally, larger pebbles may be imbricated.

A multi-storey pattern of channel-like forms with bars comparable in magnitude to that of the channel in which they occur is found in association with conglomerate sheets (blow up #11; plate 4.6a). These macroforms are typically ~1-1.5 m thick and up to ~2.5 m wide, with well defined convex-upward tops (fourth-order bounding surfaces; Miall, 1988) draped by a thin sandstone layer. These sedimentation units may show a basal lag with clasts up to ~20 cm in diameter. Imbricated clasts with their long axes transverse to dip direction, indicate paleoflows towards the east (N110°).

Interpretation. The low-relief scoured bases and sheet-like geometry of these deposits in conjunction with the distinct stratification (flat-lying strata), clast-supported nature of the conglomerate and clast imbrication are features described in conglomerate sequences interpreted as braided stream deposits (e.g., Ramos and Sopeña, 1983; Nemec and Steel, 1984). The interpretation that gravel sheets represent longitudinal bar deposits is supported by the occurrence of channel-like forms and bars with convex-upward morphologies. Well preserved bar forms with sandstone drapes deposited during low-stage flow conditions suggest that lack of dissection was probably due to rapid lateral shifting and abandonment of the individual channels, which might be related to episodic flow events as in modern ephemeral braided streams (e.g., Turnbridge, 1983, Stear, 1985).

Plate 4.6. (a) Convex-upward macroforms (bars) developed in conglomerate deposits. Bar forms are draped by thin sandstone layers (see arrows). (b) Mudstone with carbonate nodules arranged in a vertical pattern interpreted to represent immature paleosols (calcretes). Above it is a sequence of vertically stacked tabular sandstone sheets. Note the occurrence of a local scour filled with a cross-bedded pebbly sandstone.

a

b

Lithofacies 9: Mudstone and heterolithic deposits

This assemblage is developed between sandstone and conglomerate bodies and forms intervals of variable thickness up to ~9 m. It comprises red mudstone and heterolithic interbeds of mudstone-sandstone. Fine-grained centimetre-scale sandstone beds are parallel-laminated, cross-laminated or massive (Facies Fl; Miall, 1978) and are either laterally persistent or pinch out in a few metres. Bases are planar and tops are either planar or wavy due to asymmetric ripples. Siltstones may show parallel or undulatory lamination. These deposits contain small carbonate nodules dispersed or organized into discrete horizons, or they exhibit a pattern of vertically elongated concretions (plate 4.6b). Mudstone intervals may show a mottled texture and sometimes sand-filled cracks. Although rare, rootlets have been observed. Millimetre-scale horizons of gypsum are locally present between intervals of tabular sandstone sheets.

Interpretation. Fine-grained deposits are interpreted to represent flood-plain accumulations formed by settling from sheet floods or overbank vertical accretion. Lamination in siltstones likely developed as a result of deposition from suspension (e.g., Miall, 1996). The intercalated thin sandstone beds may represent splay deposits in areas distal to channel margins. Flood plains were periodically exposed as inferred from the occurrence of desiccation cracks and carbonate nodules. The nodular vertical pattern of concretions developed in mudstones is interpreted to be related to processes of soil formation, comparable in genesis to calcretes (Wright and Tucker, 1991).

Paleoenvironmental characterization

The Mariño Formation comprises a red-bed succession in which fluvial red beds are associated with eolian deposits. Several lines of evidence suggest that the Mariño stratigraphy records deposition in a low-gradient setting where arid or semiarid conditions prevailed. Some of this evidence includes the occurrence of well oxidized flood-plain deposits, incipient calcrete horizons, sandstone fluvial lithofacies that suggest deposition by ephemeral flows, and the thick succession of eolian deposits.

The lithofacies recognized in the lower member of the Mariño Formation allow for the inference that deposition occurred in channelized and perhaps less confined parts of a ephemeral stream system dominated by a network of shallow channels (purple conglomerate and sandstone, lithofacies 1). Sand-laden flow episodes likely triggered by short-lived periods of high-discharge also contributed to the alluvial basin-floor sedimentation (Flat-based gray tuffaceous sandstone, lithofacies 2). The lateral persistence of these beds and the absence of associated channels suggest that these deposits accumulated downslope of channels, at the toe of extensive bajadas that prograde onto an open area or flat plain. Although the present exposure suggests that lithofacies 2 prograded across lithofacies 1, both lithofacies may have coexisted. The interpretation that these sediments occupied a distal position in a drainage system also stems from the occurrence of eolian deposits (lithofacies 4) which sequentially through time prograde across these sediments. The development of extensive dune fields was likely favoured by progressively reduced fluvial input.

The upper member of the Mariño Formation marks the reestablishment of a stream-dominated sedimentation, with a fluvial facies assemblage that suggests deposition by

ephemeral stream processes. Similar to what has been observed in present-day semiarid regions subject to ephemeral flow deposition (e.g., Williams 1971; Stear, 1985), ephemeral flooding was both sheet-like and channelized, giving rise to a complex flow pattern that produced sandstone and conglomerate bodies, each with its own set of characteristic bed forms and geometrical features (lithofacies 5-8). The abundance of parallel-laminated and normally graded sandstones in the sedimentary assemblage (tabular sandstone sheets; lithofacies 5) is considered a distinctive element of a facies model for ephemeral streams (e.g., Miall, 1977; Turnbridge, 1981; Stear, 1983; Dreyer, 1993) and, therefore, supports the above interpretation. The widespread occurrence of horizontally laminated sandstone strongly suggests hydraulic conditions characterized by high flow velocities and shallow depths. In addition, the occurrence of channel sandstone bodies characterized by a rather unusual mix of coarse member architectures (lithofacies 6) alternating with thick flood-plain deposits (considered a relatively unusual feature of a dominantly braided system) could be related to processes of rapid sedimentation and frequent channel avulsion, conditions that occur in arid regions characterized by the occurrence of short-lived episodes of high discharge. Conglomerate deposits with bars lacking evidence of dissection are also some of the sedimentological features that suggest an ephemeral stream deposition.

The increase in the abundance of conglomerate sheets and multi-storey channel-fill deposits towards the upper part of the Mariño stratigraphy (lithofacies 6 and 8) defines a gross coarsening upward trend in the succession. Although the details of the alluvial stratigraphy in terms of autogenic and allogenic sedimentary controls cannot be determined with the available data, the overall evolutionary trend of the Mariño stratigraphy from fine-

textured (relatively more distal) deposits to coarser clastic fluvial facies suggests that allogenic influences, especially tectonism, may have played an important role in the development of the observed vertical trend (e.g., Blair and Bilodeau, 1988; Miall, 1996). This topic is addressed in the integration of all data sets (section 5a).

In summary, the Mariño Formation records fluvial-eolian deposition with a fluvial facies assemblage that shares similarities with sandy ephemeral stream deposits interpreted to record deposition in an ephemeral-stream sand-flat complex (e.g., Turnbridge 1984; Fernández, 1993). If the terminal fan model of Kelly and Olsen (1993) is to be applied as a guide to a facies model for the study deposits, Mariño Formation features would suggest that deposition mainly occurred in distal parts of distributary channels, perhaps transitional to the basinal zone of the fan (lower and middle member) and, sequentially through time, in relatively more proximal positions in the drainage system, as suggested by the development of braid-bar complexes in the uppermost part of this formation. The terminal model would require a fairly large alluvial fan system at the foot of the rising Principal Cordillera to account for the occurrence of clast lithologies (see provenance study, section 3) with unambiguous parentage in geological units cropping out in the Principal Cordillera presently at least 60 km distant, which restores palinspastically to ~100-120 km. A modern example is the Bermejo basin of west-central Argentina, in which point-source alluvial fans extend for more than 50 km from apex to toe and may include lithological units located at great distances (>150 km) from the fan (Damanti, 1993). However, drainage patterns and geographic and hydraulic factors could have exerted influences resulting in a significant departure from described models or from other modern ephemeral drainage systems.

(c) La Pilona Formation

Exposures of this unit have been studied in the La Higuera profile located in the western flank of the La Pilona anticline (section III in fig. 1.4a), where this formation reaches a thickness of ~800 m. Lateral continuity of strata is limited because of cover by Quaternary deposits, which in many cases precluded determination of the large-scale geometry of the conglomerate and sandstone bodies.

The basal part of this unit includes the deposits that have been previously mapped as 'Tobas Grises Inferiores' (Lower gray tuffs; Truempy and Lhez, 1937; e.g., De la Motta, 1957). Field mapping for this study found no 'Tobas Grises Inferiores' bounding the base of the La Pilona unit but identified a tuffaceous interval in the middle part of the unit, which is clearly exposed in two of the study sections (Arroyo del Agua Blanca and La Higuera profiles; sections II and III in fig. 1.4a). This interval is considered to be a facies of the La Pilona Formation developed within this formation, in agreement with Kokogian et al. (1987).

In the study section (fig. 4.2), above a strike-slip fault affecting conglomerate strata of the Mariño Formation, there is an approximately 120 m thick interval of rocks of the same formation. Because of poor exposure, the location of the contact with the overlying La Pilona Formation is equivocal. Therefore, the initiation of sedimentation of the La Pilona Formation is placed with the first appearance of a gray medium to fine conglomerate with sandstone layers, which is typical of this unit.

This formation contains a few tephra horizons. In particular, there is a ~1 m thick

Fig. 4.2. La Higuera (a), Quebrada de la Angostura (b) and El Molino (c) profiles, showing details of textures and sedimentary structures of parts of the La Pilon, Tobas Angostura and Río de los Pozos formations.

fallout tephra with abundant biotite crystals known as the 'Horizonte de Ceniza Blanca' (White ash horizon; Truempy and Lhez, 1937), which is laterally traceable for at least ~500 m (plate 2.1). This horizon is situated approximately in the middle part of the La Pilona stratigraphy (sample Md-6; cf. Chapter 2).

Several lithofacies are recognized in this unit and are described next.

Lithofacies 1: Sheets of massive or flat-bedded conglomerate

This lithofacies is defined by a sheet-like flat-bedded or massive conglomerate of variable thickness usually up to ~1.5 m, which is found as single sheets or, more commonly, in vertically stacked successions (fig. 4.2, blow up #a6 and #a7). Single sheets or conglomerate units comprising a few individual sheets occur in the lower part of the La Pilona Formation. These sequences commonly show thin sandstone wedges or layers (up to ~20 cm thick) that permit clear identification of individual sheets. In the uppermost ~200 m of the unit, conglomerate sequences ranging in thickness from ~8 to 35 m show internally multiple planar and sometimes curved erosional surfaces with sandstone wedges or thin siltstone layers sometimes preserved at the top of individual sheets. Beds are massive or horizontally stratified (Gm facies; Miall, 1978) and may show an upward decrease in pebble size and transition into a pebbly sandstone and sandstone. Cross-bedded sets are rare. Sometimes, thin sequences (a few decimetres) of a trough cross-bedded coarse to medium sandstone are intercalated within coarser deposits. Conglomerates are clast-supported with a sandstone matrix, and clasts are subrounded and subangular. Clast size normally ranges from 5 to 15 cm but sometimes reaches 20-25 cm. Clasts are commonly imbricated (a-axis transverse to the dip direction of

the clast) suggesting paleoflow directions to the southeast. Although rare, a multi-storey pattern of channel-like forms of variable dimensions up to ~2.5 m wide and a few decimetres thick occurs in lateral and vertical association with conglomerate sheets in the upper part of the La Pilona Formation. The erosional base of these units may be followed by a lag deposit of coarser grain size than the trough fill. Channel fills consist of fine to medium gravels passing up or laterally into sandy gravels or a trough cross-stratified conglomerate and pebbly sandstone.

Conglomerates are composed of clasts of volcanic rocks, tuffs, granites, and sedimentary and low-grade metamorphic rocks (including graywackes, quartzites, sandstones, and limestones).

Interpretation. This lithofacies is interpreted to result from traction deposition in fluvial channels. This interpretation stems from the prominence of imbricated conglomerates and the distinctive stratification (i.e., massive or crude stratification) that typifies longitudinal-bar deposits (e.g., Hein and Walker, 1977; Ramos and Sopena, 1983). Associated sandstone lenses likely represent waning-flow deposition on bar tops and margins or in channels abandoned following periods of high discharge. The recognition of a multi-storey pattern of channel fills in association with gravel sheets allows for the inference that these deposits likely record deposition in multiple channel courses. This style of internal geometry typifies fluvial deposits interpreted as representing streams of braided morphologies (e.g., Williams and Rust, 1969; Ramos and Sopena, 1983).

Lithofacies 2: Tuffaceous sandstone sheets

This lithofacies is commonly defined by stacked sheet-like beds and lenticular sedimentation units of a thin tuffaceous sandstone and pebbly sandstone ranging in thickness from ~5 to 50 cm (fig. 4.2, blow up #a2; plate 4.7). Sheet-like beds are usually up to ~30 cm thick and they can be traced laterally for several metres. Alternating siltstone and sandstone beds also occur defining sequences a few metres thick. Sandstone beds show flat bases, locally with low-angled erosion surfaces exhibiting less than ~15 cm of relief. Beds are parallel-laminated, thinly stratified or massive. Many beds are normally graded and consist of a pebbly sandstone or matrix-rich fine conglomerate with small pebbles of ~1 cm in diameter (rarely up to ~3 cm) that grade into a sandstone with sometimes a siltstone drape. In the lowermost part of the La Pilona Formation, vertically stacked centimetre- to decimetre-scale normally graded beds make sequences up to ~2.5 m thick (fig. 4.2, blow up #a1). Thin alternations of sandstone and pebbly sandstone or fine conglomerate are frequent. In particular, blueish conglomeratic sandstone bodies of limited thickness (~up to 1.5-2 m) occur in the middle to upper part of the La Pilona Formation which are characterized by alternating thin layers (usually <15 cm thick each) of conglomerate and sandstone (fig. 4.2., blow up #a5). Cross-beds are rare and, when observed, are associated with shallow channel-like forms or scours.

Lenticular sedimentation units bounded by scour surfaces up to ~0.5 m deep truncate underlying flat-bedded strata. Cut-and-fill relief along their basal surfaces is locally observed. These units are commonly filled with a horizontally bedded sandstone, and their bases may contain small pebbles. They record single or multiple episodes of deposition. Surfaces of

Plate 4.7. (a) Stacked sheet-like beds of tuffaceous sandstone and pebbly sandstone of the La Piona Formation. Note the occurrence of lenticular sedimentation units intercalated in this sequence. (b) Detail of normally graded beds.

a

b

lateral accretion of restricted extent are locally observed.

This lithofacies mainly occurs in the lower part of the La Pilona Formation. The weathered nature of these deposits in some stratigraphic intervals between ~60 and ~120 m above the base of the unit precludes determination of individual bed thickness and recognition of internal features. In some cases, sparse lapillus clasts were identified in a massive or faintly stratified sandstone.

Interpretation. The thin-bedded nature, lateral extent, occurrence of normally graded and parallel-laminated sandstone beds and low-relief of scours suggest deposition by shallow, rapid flows. The preferential deposition of transitional to upper-flow-regime beds (laminated sandstones), near absence of overbank fines, occurrence of matrix-rich normally graded beds capped by siltstone drapes, and the presence of massive sandstones suggest that these deposits record a regime of flash discharge and rapid deposition which could be related to flood episodes (e.g., Turnbridge, 1981). Although sediment transport and deposition are inferred to occur from unconfined shallow flows, the occurrence of lenticular sedimentation units with local cut-and-fill structures and evidence of restricted lateral accretion might reflect the existence of poorly defined channels.

Lithofacies 3: Conglomerate and sandstone sheets in alternation

This lithofacies consists of interbedded conglomerate and tuffaceous sandstone. Beds are sheet-like or lenticular and vary markedly in thickness ranging from several centimetres to usually less than ~1.5 m (fig. 4.2, blow up #a3 and #a4; plate 4.8). The interbedding of

Plate 4.8. (a) Conglomerate and sandstone sheets or lenses in alternation. Note the occurrence of sandstone scour-fills and solitary sets of trough cross-bedded strata (near the field assistant; see arrows) in lateral and vertical association with conglomerate. (b) Gravel sheets with crude horizontal stratification interbedded with decimetre-scale sandstone lenses.

a

b

facies is apparently unsystematic. The vertical superposition of gravel and sandstone sheets defines successions up to ~25 m thick, which are bounded by mudstone intervals a few metres thick. Conglomerate sheets and lenses show flat, slightly irregular or curved erosive basal surfaces. Thin gravel sheets (up to ~20 cm thick) may be normally graded, sometimes with an upward transition to pebbly sandstone. Lateral grading of grain size in thin gravel sheets may locally define sandstone-rich layers with sparse pebble clasts. In contrast, thicker conglomeratic units show cut-and-fill basal relief with metre-scale scours and erosion surfaces. The coarser sedimentation units consist of a moderate to poorly sorted clast-supported fine to medium conglomerate with a locally abundant sandstone matrix. Conglomerates are massive or horizontally stratified, and the larger discoidal clasts may be imbricated. Measurement of clast fabric in conglomerate units filling large scours suggest paleoflow directions towards the south-southeast. Cross-stratification, commonly low-angled, is associated only with gravels filling scours with a few decimetres of relief. Clast sizes in thin gravel units are commonly less than ~4 cm in diameter and up to ~10 cm in thicker gravel sheets. Sandstones occur as thin, discontinuous massive layers separating individual gravel units, as horizontally bedded sheet-shaped units in alternation with gravel sheets or as lenticular units filling shallow scours of variable width. Scour-fills and solitary sets of trough cross-bedded strata occur in lateral and vertical association with channelized gravel units. In other cases, broad scours (a few metres wide) that may cut into each other are filled with a low-angle cross-bedded sandstone. A solitary set of cross-bedded strata may occur above gravel sheets.

This lithofacies mainly occurs above sequences dominated by tuffaceous sandstone

sheets between 150 and 350 m above the base of the unit. The relative proportion of conglomerate compared to sandstone sheets varies throughout vertical successions. Intervals of thin gravel and sandstone sheets alternate with thicker, scoured-based conglomerate and sandstone units. As a general trend, there is an increase in the abundance of coarser and thicker conglomeratic units towards the middle part of the La Pilona Formation.

Interpretation. This lithofacies is interpreted to record deposition by sheet-flood and stream-flow processes. Thin gravel sheets, sometimes a few clasts thick, in alternation with sandstone could represent the deposits of shallow, unchannelized or poorly confined flows. This interpretation stems from the lateral extent, limited thickness of conglomerate strata and flat or low-relief basal surfaces lacking evidence of significant erosion. A regime of varying discharge and discontinuous accretion is suggested by the intimate association with sandstone strata whereas rapidly waning flood action or declining discharge is inferred from the occurrence of thin normally graded beds. Thicker gravel sheets with cut-and-fill basal relief and thin drapes of sandstone are interpreted to record deeper, channelized flows in channels or strong channelized flood events, with sandstone representing low-stage deposits preserved on bar tops and margins. The sandstone scour fills could be interpreted as trough fills associated with the migration of sand dunes. The general lack of pronounced incision is interpreted to reflect the existence of broad, relatively shallow channels which often cut into a sandy substrate.

Lithofacies 4: Poorly sorted, matrix-supported to matrix-rich clast-supported conglomerate

Beds of this lithofacies have sharp, nonerosive or locally erosive bases and limited thicknesses (a few tens of centimetres). Conglomerates are poorly sorted, polymodal, with variable clast size distribution (~1-10 cm in diameter). Larger clasts may be in contact defining a framework filled with a poorly sorted sandstone with granules, or pebbles are dispersed in a sandstone matrix, with a gradational spectrum between these two variants. Clast fabric is disorganized, although in some beds the larger tabular clasts assume a preferential horizontal orientation. Crude normal grading is sometimes noted. Top of beds may be capped with a massive or thinly bedded medium sandstone. These deposits occur in the lower to middle part of the La Pilona Formation interbedded in sequences of gravel and sandstone sheets.

Interpretation. Poorly sorted conglomerates either lacking a framework or matrix-rich clast-supported, with a disorganized clast fabric suggest that deposition and transport were related to debris-flow events (e.g., Schultz, 1984). High velocity surges and a relatively modest level of sediment concentration could have favoured the crude normal grading in beds (e.g., Collinson and Thompson, 1989).

In some cases, however, it remains unclear whether or not matrix-rich conglomeratic units with poor to moderate sorting showing transition to a pebbly sandstone rich in small pebbles may represent the product of fast deposition by stream-flows rather than debris-flows.

Lithofacies 5: Fine-grained deposits

This lithofacies comprises the following: (a) Red mudstones or a heterolithic association of thin massive, parallel-laminated or rippled sandstone beds interbedded with mudstone (fig. 4.2, blow up #a1). Mudstone may be nodular in some horizons or contain diffuse carbonate nodules. (b) Greenish and grayish laminated or massive mudstones, sometimes rich in mica. Inter laminations of a very fine sandstone, siltstone and claystone are observed. These deposits first occur ~200 m above the base of the unit as thin accumulations or interstratified with reddish mudstones and contain undetermined organic carbon remains (fig. 4.2, blow up #a4, between ~305 and 310 m). Green mudstone intervals become thicker up section reaching thicknesses of several metres (~5-10 m) in the middle part of the La Piona stratigraphy, although greater thicknesses may be hidden under the Quaternary cover.

Interpretation. The red mudstone and heterolithic association of sandstone and mudstone found in the lower part of the stratigraphy is interpreted to represent flood-plain facies developed by settling from suspension of waning floods or overbank processes. The intercalated thin sandstone beds may represent splay deposits in areas distal to channel margins or very low-energy flood deposits. Flood plains were probably exposed periodically as inferred from the occurrence of red mudstones and incipient carbonate nodules. However, sequentially through time, shallow standing bodies of water or ponds might have occupied depressed areas in the flood-plain, as suggested by the occurrence of laminated greenish mudstones upward in the stratigraphy. The occurrence of lacustrine facies associated with thin gypsum horizons has been mentioned in other studies (e.g., Yrigoyen, 1993a).

Paleoenvironmental characterization

La Pilona Formation is interpreted to record fluvial and clay playa deposition under conditions similar to those that prevailed during sedimentation of the Mariño Formation, i.e., an arid or semi-arid low-gradient setting in which transport and deposition of gravel and sand were largely controlled by a highly variable or seasonal discharge. Some of this evidence includes the occurrence of red mudstones suggesting oxidized flood plains, incipient calcrete nodules, gypsum horizons in green mudstones, and a sandstone fluvial lithofacies dominated by upper-flow-regime parallel-laminated beds (lithofacies 2), a feature characteristic of many modern and ancient ephemeral-flow sands deposited in semi-arid regions (e.g., Turnbridge, 1981a; 1984). Additional indicators of the existence of a regime of varying and/or rapidly waning discharge are given by the frequent vertical transitions between conglomerate and sandstone sheets (lithofacies 3) and the abundance of normally graded beds in sandstone and gravel lithofacies.

The association of sheet-flood, stream-flow, debris-flow, and suspension-dominated flow deposits including thick clay sequences allows for the inference that these deposits accumulated in distal parts of a drainage system, perhaps at the toe of bajadas and transitional to a flat plain, in which water could have ponded and dumped suspended silt and clay at the distal fringes of sheet-flood units or in terminal parts of a flood basin or a playa mud flat. The occurrence of superimposed sheets of sand in the basal part of the La Pilona Formation suggests the existence of repeated episodes of sheet flooding where unconfined flows spread over extensive areas (lithofacies 2).

The La Pilona Formation records a gross coarsening and thickening upward trend in

the lower ~300 m. This trend is defined by the dominance of sandstone sheets towards the base that grade upward into alternating thin units of conglomerate and sandstone, which in turn are overlain by thicker and coarser gravel and sand sheets with evidence of channelization (lithofacies 3). The occurrence of up to ~25 m thick successions dominated by scoured-based sandstone and gravel deposits lacking fine-grained overbank deposits points towards the active nature of the channel courses, suggesting a braid-plain origin for these sequences. The occurrence of channelized deposits capped by sheet-flood units suggests that sheet floods sometimes tap the bed load of the channels. In other cases, episodes of gravel and sand deposition in channels alternated with sequences representing successive sheet-flood events. As suggested by paleocurrent data, gravels, mostly volcanic rocks, were transported from northwestern highland sources.

Sequentially through time, there was progressively less sediment input at this location in the basin as inferred from the occurrence of thinner sand-dominated successions and thicker intervals of mudstone with calcareous concretions (incipient calcretes). This evidence suggests relatively longer periods of subaerial exposure. Quiescent conditions punctuated by episodic flow events that introduced coarser material seemed to dominate in the middle part of the La Pilona Formation. This interpretation stems from the dominance of extensive and thick sheet-like mudstone units (lithofacies 5) which alternate with sandstone and conglomerate deposits of limited thickness. In this flat environment, temporary and semi-permanent? shallow lakes developed as inferred from the occurrence of green laminated mudstones with thin horizons of gypsum.

The presence of a few tephra horizons interbedded with fine-grained deposits provides

further evidence for a low-energy environment which allowed preservation of up to ~ 1 m thick air-fall beds. The occurrence of tephra horizons confirms the existence of extrusive episodes in the Cordilleran region (cf. Pérez and Ramos, 1996) synchronous with sediment deposition.

The thick conglomerate succession that characterizes the uppermost ~200 m of the La Pilona Formation marks the reestablishment of a stream-dominated setting where traction deposition in fluvial channels was the dominant process (lithofacies 1). The shift upward from fine-textured facies to coarser deposition stems from progradation, where increasingly coarse debris derived from the northwest was transported by streams with braided morphologies.

In summary, the La Pilona Formation comprises stream-flow sequences, flood sheets and clay playa deposits which record deposition in an arid or semi-arid setting. Concurrent lines of evidence suggest that the drainage system had a regime of highly variable discharge, probably ephemeral. The facies assemblage bears similarities with other ancient alluvial stream complexes interpreted to record deposition at the distal parts of mixed-load terminal fans (e.g., Kelly and Olsen, 1993). The abundance of extensive sheet-flood units in association with braid-plain deposits suggests that these successions accumulated in distal to medial parts of the 'distributary zone' of Kelly and Olsen's (1993) terminal fan model, where channels die out and flows become progressively less confined down current.

Details of the alluvial stratigraphy in terms of allogenic and autogenic sedimentary controls cannot be determined with the available data. However, the upward coarsening and thickening trend in the lower part of the La Pilona Formation succeeded by a dominantly muddy zone above it, and the development of thick conglomeratic successions in the

uppermost sequence suggest that tectonism played a role in the development of the observed vertical facies trend. This topic is addressed in section 5b where the complete data set is considered.

(d) Tobas Angostura unit

This unit, previously termed 'Tobas Grises Superiores' (Upper gray tuffs; Truempy and Lhez, 1937), comprises volcanogenic deposits characterized by a distinctive gray colour and weathered appearance that are in contrast with the rest of the Neogene succession. These deposits include coarse and fine tuffs, sandstones and siltstones with varying amounts of pyroclastic material, pumice lapilli layers, lesser amounts of tuffaceous conglomerates and, more rarely, agglomerates. Conglomeratic intercalations are more abundant in the basal and upper parts of the unit. This unit rests unconformably upon the La Pilona Formation on a low-relief erosional surface (Yrigoyen et al., 1954) that marks a stratigraphic omission of a few thousand years (cf. Chapter 2).

Lithofacies determinations are based on sedimentological data obtained from the La Higuera profile, in the western flank of the La Pilona anticline (section III in fig. 1.4a), and the Arroyo de la Angostura profile, in the eastern flank of the Tupungato anticline (section V in fig. 1.4c). In the La Higuera section (fig. 4.2, profile a) this unit attains a thickness of 102 m, the uppermost 10 m of which is mostly covered and represents the gradual transition into the overlying Río de los Pozos Formation. The precise location of the unconformity that separates this unit from the underlying La Pilona Formation is rather ambiguous. Therefore,

it is assumed that the base of the unit is marked by a ~10 m thick greenish mudstone interval that becomes progressively more tuffaceous upward and contains intercalations or lenses of an ash-rich sandstone and scattered lapilli (fig. 4.2, blow up #a8). In the Quebrada de la Angostura profile (fig. 4.2, profile b), this unit reaches a thickness of ~72 m. In this section, a distinctive tuffitic, carbonate-rich dark bed bounded by mudstone is taken arbitrarily as marking the gradational contact between the Tobas Angostura and Río de los Pozos formations (fig. 4.2, blow up #b3).

The basal and upper parts of this formation are characterized by the dominance of volcanoclastic deposits, mainly epiclastic volcanic rocks (*sensu* Fisher, 1961) mixed in varying proportions with nonvolcanic fragments. The deposits from which this unit derived its name (Upper gray tuffs) are found ~10-20 m above the base of the unit and consist of volcanoclastic rocks with abundant pyroclastic fragments (i.e., lapilli) or particles (i.e., fine and coarse ash). In this part of the succession, one of the most distinctive features is the common interbedding of different facies, which are described next. Stratigraphic logs showing details of the stratigraphy and sedimentary structures are shown for the deposits of the Quebrada de la Angostura section (profile b).

Lithofacies 1: Horizontally laminated sheet-like beds

This lithofacies comprises lenticular and extensive sheet-like beds that occur as stacked, ~2-30 cm thick tuffitic beds or bed sets separated by siltstone partings or thin mudstone layers (fig. 4.2, blow up #b2). Lenticular sand sheets are laterally persistent for several metres and may show thin sandstone sheets that extend outwards from a thicker

central part (plate 4.9a). Sheet-like beds have planar nonerosive and erosive, locally slightly concave-up, basal surfaces. Tops of beds may be truncated by centimetre- to decimetre-scale intercalations of a monomictic conglomerate composed of lapillus clasts with an ash-rich matrix.

Sandstone is pebbly, medium- to fine-grained, with horizontal lamination, low-angled stratification, normal grading and scour-fill cross-bedding when filling shallow and large scours. These deposits occur in alternation with sequences of a medium to very coarse horizontally bedded sandstone (lithofacies 2).

Interpretation. The sheet-like geometry, great lateral extent and limited thickness of beds in conjunction with almost exclusive occurrence of upper-flow-regime structures, low-relief scours and normally graded beds strongly suggest deposition by shallow, unconfined or broadly confined rapid flows (sheet floods). The dominance of parallel-laminated sands could be related to ephemeral flood sedimentation where sporadic and perhaps violent rainfall, favoured episodes of upper-flow-regime discharge.

Lithofacies 2: Medium to very coarse horizontally bedded sandstone deposits

These deposits make up successions that range in thickness from ~0.5 m to several metres (fig. 4.2, blow up #b1). Sedimentation units consist of a medium to very coarsely grained sandstone with horizontal bedding usually defined by gradational stratal contacts. Strata are a few millimetres or centimetres thick, usually too thick to be called laminae (plate 4.9b and 4.10). Coarse strata are commonly laterally continuous for a few metres while darker

Plate 4.9. (a) Horizontally laminated sheet-like beds and bedsets in the Tobas Angostura unit. Note the occurrence of lenticular sedimentation units with a thicker, scoured-based central part (see arrows) and thinner lateral sheets. (b) Horizontally bedded coarse tuffaceous sandstone with abundant lapilli and pumice out-size clasts.

a

b

Plate 4.10. (a) Detail showing a horizontally bedded tuffaceous sandstone with out-sized lapillus clasts. Note the nature of the stratification, i.e., dominantly horizontal, thick laminae with locally diffuse stratal contacts (see arrow) and a wide range of grain sizes. (b) Debris-flow deposits (near the base of the sequence) interbedded with horizontally bedded sandstone interpreted as hyperconcentrated flood-flow deposits.

a

b

finer-grained beds are more persistent. Out-sized lapillus clasts are abundant. Coarser beds also show abundant lapilli which define continuous horizons. Strata are horizontal except where conformable upon inclined erosion surfaces.

These beds also occur in association with normally graded sequences. These deposits consist of a massive, pebbly, coarse-grained sand or pebble- to cobble-sized gravel (mostly lapilli and scarce volcanic and sedimentary rocks) filling scoured surfaces overlain by horizontally-bedded sandstone. Some graded-stratified deposits grade upward into a rippled sandstone unit (up to ~ 15 cm thick).

Although rare, there are tabular or lenticular decimetre-scale beds (up to ~0.5 m thick) of a matrix-supported conglomerate with sharp, nonerosive bases. These beds are nearly monomictic, composed of pumice and lapillus clasts of variable size up to ~10 cm in diameter dispersed in a coarse tuffaceous sandstone matrix. A poorly sorted matrix-supported conglomerate containing pebbles, cobbles and sometimes boulders of volcanic and sedimentary rocks also occurs interbedded within these sequences, as observed in other localities (plate 4.10b).

Interpretation. These deposits share many similarities with modern and ancient fluvial sequences interpreted to result from hyperconcentrated flood-flow events (e.g., Smith 1986, 1987). This interpretation stems from the recognition of several features, such as the nature of the stratification (i.e., horizontal bedding with diffuse stratal contacts, discontinuous strata), common occurrence of out-sized clasts, lack of internal erosion surfaces within the successions and absence of cross-strata. Based on flume and pipe experiments, Smith (1986)

suggested that the distinctive horizontal bedding could be related to the migration of low-amplitude, long-wavelength, dune-like forms with flows characterized by high suspended-sediment loads that inhibit formation of bed forms that produce cross-stratification (for further discussion, see Smith, 1986). The occurrence of normally graded deposits capped by a rippled interval suggests that, in some cases, the sediments were the product of rapid deposition under conditions of waning flow energy from high-sediment-load floods.

The nearly monomictic matrix-supported conglomerate beds are interpreted as of debris-flow origin. The association debris-flow and hyperconcentrated flood-flow deposits commonly occur in volcanoclastic sequences as a result of lateral transformations in volcanoclastic flowage processes (e.g., Smith, 1986; Haughton, 1993).

Lithofacies 3: Multi-storey channel-fill deposits

These deposits comprise shallow channel-like forms of variable dimensions up to ~2 m wide and ~0.5 m thick that commonly cut into each other. Small hollows (a few decimetres wide) are filled with a poorly sorted clast-supported fine conglomerate above which there is a parallel-laminated or massive sandstone. Conglomerates are composed of extrabasinal pebbles and abundant lapillus clasts of varying size up to ~5 cm, with a framework filled with an ash-rich matrix. Metre-scale channel-like forms exhibit concave-up bases that truncate underlying beds and pass laterally into a flat nonerosive concordant basal surface. They are filled with a horizontally bedded or laminated sandstone. In the uppermost ~20 m of the Tobas Angostura unit (fig. 4.2, blow up #b3), vertical sequences of irregular erosion surfaces separate horizontal or low-angled ($<10^\circ$) cross-beds and, locally, cosets of mutually cross-

cutting troughs that grade upward into a tuffaceous siltstone. These sequences are up to ~1.5-2 m thick.

Interpretation. Multi-storey sandstone bodies characterized by cut-and-fill structures suggest the existence of repeated episodes of scouring, infilling and switching of small channels in mobile channel belts. This lenticular or multi-storey style of internal geometry is recognized in fluvial sandstone bodies interpreted as representing streams of braided morphologies (e.g., Cant, 1978; Allen, 1983). The occurrence of normally graded beds and the dominance of laminated sandstones in multi-storey channel-like forms suggest that these channels were filled by several, upper-flow-regime flows. The vertical sequences of erosion surfaces overlain by horizontally laminated sandstone suggest deposition by shallow flows during periods of high discharge over broad areas.

The fining-upward sequences composed of trough cross-bedded sandstones suggest traction deposition in fluvial channels involving migration of sinuous-crested dunes (cf. Miall, 1977).

Lithofacies 4: Ash beds

Ash horizons are found in units with laterally variable thickness, ranging from a few centimetres to commonly ~15 cm and exceptionally up to ~35 cm (plate 2.2a). These beds are whitish or dark gray and show normal or inverse grading, or they are massive (Fig. 4.2, blow up #b1). They consist of a crystal-rich coarse to fine tuff, with angular constituent grains, commonly with abundant biotite and hornblende crystals.

Interpretation. The characteristics of these deposits allow for the inference that they represent air-fall tephras (e.g., Fisher and Schmincke, 1984). Although most of these beds occur as primary deposits, some of them may have been reworked but not extensively mixed with other sediment.

Lithofacies 5: Fine-grained deposits

Fine-grained intervals include mudstones and tuffaceous mudstones sometimes with interbeds of thin pumice lapilli-rich horizons (a few centimetres thick) and intercalations of massive ash-rich beds (10-20 cm thick). These intervals may contain sparse lapilli. They have variable thicknesses, commonly between 0.5 and 2 m. The thickest interval (up to ~10 m) is found in the basal part of the unit (La Higuerita section) where these deposits consist of a greenish laminated or massive mudstone with lenses of an ash-rich sandstone. Reddish mudstones are interbedded with sandstone and conglomerate strata in the uppermost part of the Tobas Angostura Formation. Rootlets sporadically occur in thin siltstone layers.

Thin carbonate-rich horizons a few millimetres thick are locally present at the top of mudstone intervals. Travertine horizons showing a laminated texture that resembles algally produced stromatolitic laminations occur in the upper part of the Tobas Angostura stratigraphy (fig. 4.2, blow up #b3).

Interpretation. These deposits are interpreted to represent deposits of waning floods and overbank deposits. The intercalated thin sandstone beds may represent splay deposits in areas distal to channel margins or sheet-flood deposits. The restricted green laminated

mudstone facies that marks the base of the unit in the La Higuera profile could be related to lacustrine fine-grained sedimentation. The occurrence of travertine deposits suggests the existence of local springs of carbonated water (Collinson and Thompson, 1989), much like modern springs near the study locality. The incipient carbonate nodules could represent the first stage in the development of mature soils (calcretes).

Paleoenvironmental characterization

The Tobas Angostura unit exhibits a variety of facies that can be ascribed to different processes of transportation and deposition. The volcanoclastic nature of the succession, the abundance of parallel-laminated sheet-like sandstone beds, the frequent vertical transitions of different facies including stream-flow, sheet-flood and hyperconcentrated-flow deposits, debris-flow accumulations and air-fall tephra suggest that sedimentation was the response of a semiarid fluvial system to the pyroclastic volcanism and that deposition occurred in broad extensive areas. The two study profiles show that the initiation of Tobas Angostura deposition was dominated by epiclastic sedimentation followed by sediment sourced from explosive volcanism, as locally recorded by air-fall tephra. Explosive volcanic events are interpreted to have been responsible for creating high sediment fluxes and, in turn, allowing rapid aggradation as evidenced by the occurrence of extensive sheets of pumiceous sandstones intercalated with hyperconcentrated-flow deposits and nearly monomictic debris-flow deposits. This rapid aggradation was in turn favoured by the existence of a drainage system where discharge was likely controlled by seasonal or sporadic rainfall. This interpretation stems from the dominance of a parallel-laminated sandstone facies and the occurrence of

normally graded beds in channel fills, features commonly found in sequences interpreted to be related to flash-flood episodes in semiarid regions (e.g., Turnbridge, 1981; Stear, 1983).

The occurrence of dilute facies (stream-flow deposits) intercalated within sequences dominated by hyperconcentrated-flow deposits probably resulted when aggradation ceased and streams incised channels that were filled with sediment during the next depositional event. Towards the upper part of the Tobas Angostura stratigraphy, the dominance of sandy stream-flow deposits, the occurrence of conglomerates with a greater amount of extrabasinal clasts and the decrease in the amount of pyroclastic material in rocks are indicative of a diminished influx of pyroclastic sediment through time. The shallow scour-and-fill structures found in sandstone sequences likely represent shallow braided streams resulting from flash-flood events in arroyos of the region. Repeated episodes of upper-flow-regime discharge are suggested by the occurrence of vertical sequences of erosion surfaces with parallel-laminated sandstones.

Lateral variations in thickness of this unit define an east-southeast tapering wedge of volcanic and sedimentary rocks (cf. Yrigoyen et al., 1954), which allows for the inference that the eruptive vents were located to the west-northwest of the study area. Sedimentation of this unit between ~8.9 and 8.7 Ma (cf. Chapter 2), overlapped in time with the latest extrusive episodes occurring in the Principal Cordillera, which are recorded by the thick pile of pyroclastic flow deposits, lava rocks and tuffs that crop out in the Aconcagua region (Complejo Volcánico Aconcagua; Ramos et al., 1996a).

(e) Río de los Pozos Formation

This unit is gradational from the Tobas Angostura unit; previously termed 'Serie Amarilla' (Yellow sequence; Truempy and Lhez, 1937), it is composed of yellowish mudstones, tuffaceous mudstones and sandstones, and fine-grained conglomerates. In the Tupungato area, this formation exhibits a variable thickness, from 230 to 400 m, due to the unconformity that separates this unit from the overlying Mogotes Formation (Yrigoyen, 1993a).

Exposures of this unit have been studied in two sections: (a) Quebrada de la Angostura profile, in the eastern flank of the Tupungato anticline (section V in fig. 1.4c) and (b) El Molino profile, in the eastern flank of the La Pilona anticline (section IV in fig. 1.4a). The Quebrada de la Angostura section shows finer-grained deposits when compared with the El Molino section (fig. 4.2, profile b and c respectively). The Quebrada de la Angostura profile is characterized by the occurrence of tuffaceous mudstone with interbeds of sandstone and mudstone ranging in thickness from 0.5 m to ~4 m which alternate with tuffaceous sandstone bodies a few decimetres to several metres thick (lithofacies 1). There are rare conglomerate deposits. The El Molino profile is characterized by intervals of varicoloured tuffaceous mudstones up to ~7 m thick which alternate with massive or crudely bedded conglomerates defining sequences ranging in thickness from ~1.5 to ~13 m (lithofacies 2). In the latter section, the boundary between the Tobas Angostura and Río de los Pozos formations is placed arbitrarily at the first appearance of a fine to medium conglomerate which occurs above a ~5 m thick greenish mudstone interval overlying a highly tuffaceous succession.

Lithofacies 1: Light brown, generally massive tuffaceous sandstone (Quebrada de la Angostura profile)

Generally massive, light brown or yellowish tabular sandstone bodies with flat nonerosive bases (rarely locally erosive) occur as the dominant lithofacies above the tuffitic carbonate-rich dark bed considered to be the uppermost Tobas Angostura strata (fig. 4.2, blow up #b4; plate 4.11a). Sandstones are fine- to medium-grained and tuffaceous (tuffites). Thicknesses of individual beds are often not discernible, in part because of the weathered nature of the exposures. Beds, clearly discernible, are up to ~1.5 m and sandstone sequences persist laterally over the ~60 m wide exposure. Diffuse bed contacts in some intervals allowed recognition of ~10-40 cm thick vertically stacked beds, some of which displayed a faint parallel lamination or thin stratification.

Distinctively, some successions (up to ~10 m thick) show vertical sequences of thin calcareous horizons or crusts spaced several centimetres or decimetres apart. Nodular or concretionary carbonate-rich horizons also do occur (plate 4.11b). Sandstone layers are locally mottled, with patches of colour occurring on the bed surface. Bioturbation is suggested by the occurrence of discontinuous lumps or pockets with different texture (e.g., grain size, compaction) within beds and by the presence of abundant pits in coarse grained sandstones (plate 4.11b). In these sequences, remnant sedimentary structures suggest original deposition as horizontally bedded or scour-fill sandstone (facies Sh, Ss and St?; Miall, 1977). A pebbly sandstone and fine conglomerate sometimes occur intercalated in sandstone sequences above broadly concave-up internal erosion surfaces.

Plate 4.11. (a) Tuffaceous tabular sandstone bodies, generally massive, of the Río de los Pozos Formation. Note diffuse bed contacts about a third of the way down the sequence. (b) Detail of sandstone sequence showing carbonate-rich nodular horizons (bottom) and overlying sandstone with abundant pits thought to be the result of bioturbation.

a

b

Interpretation. The weathered nature of the exposures in conjunction with little remaining evidence of stratification preclude detailed depositional interpretation. However, the lateral persistence, flat bases and, where discernible, limited thickness of sandstone beds (locally thinly bedded or laminated) suggest deposition by sheet-flood processes and sometimes from broadly confined flows. Although massive sandstone beds or intervals could represent the product of rapid deposition, the occurrence of vertical sequences with evidence of pedogenic processes (calcareous crusts, nodules) suggests that deposition of sand was followed by prolonged periods of bioturbation that homogenized the new deposits and obscured depositional surfaces.

Lithofacies 2: Massive or crudely bedded conglomerate (El Molino profile)

Conglomerate deposits are massive or they show a crude horizontal stratification, locally curved following the lower erosional bounding surface (plate 4.12a). Internal erosion surfaces define conglomerate sheets of variable thickness up to ~1.5 m (fig. 4.2, blow up #c1). Clast-supported conglomerates are composed of gravels ranging in diameter from ~5 to ~15 cm. These gravels include volcanic rocks, lapilli clasts from the underlying Tobas Angostura unit, intrusive rocks, and low-grade metamorphic and sedimentary rocks. Conglomerates have a tuffaceous sandstone matrix, locally very abundant. Clasts are locally imbricated (a-axis transverse to the dip direction) suggesting paleoflow directions to the southeast. Conglomerate sequences may locally consist of single beds (<1.5 m thick each) or of vertically stacked decimetre-scale beds showing an upward decrease in pebble size and a transition to a pebbly sandstone and sandstone. Sandstone wedges 5-15 cm thick occur

Plate 4.12. Río de los Pozos Formation. (a) Exposure of alluvial gravel deposit (lithofacies 2) in the El Molino profile. (b) Travertine horizons with a laminated texture (see arrow) in a mudstone sequence in the Quebrada de la Angostura profile.

a

b

intercalated in conglomerate sequences.

Interpretation. This lithofacies is interpreted to result from traction deposition in fluvial channels. This interpretation stems from the distinctive stratification and from the occurrence of imbricated conglomerates, features that typify longitudinal-bar deposits (e.g., Miall, 1977; Ramos and Sopena, 1983). Associated sandstone wedges likely represent waning-flow deposition on bar tops and margins or in channels abandoned following periods of high discharge. The fining-upward beds record the existence of waning flows.

Lithofacies 3: Fine-grained deposits

In the Quebrada de la Angostura section, these deposits consist of yellow and greenish mudstones, either massive or laminated, sometimes with intercalations of centimetre- to decimetre-scale sandstone beds with flat bases and flat tops. Cross-stratification produced by ripples may be present in the silt and thin sand beds and undulating bedding is sometimes observed. These intervals contain thin carbonate-rich layers and travertine horizons showing a laminated texture (plate 4.12b). In the El Molino section, tuffaceous mudstone intervals are highly weathered and some horizons consist of a light-coloured bentonite.

Interpretation. These deposits are interpreted to represent deposits of waning floods and overbank deposits. The intercalated thin sandstone beds might represent splay deposits in areas distal to channel margins or sheet-flood deposits. Like in the Tobas Angostura unit, the occurrence of travertine deposits suggests the existence of local springs of carbonated

water (Collinson and Thompson, 1989), much like modern springs near the study locality. The occurrence of carbonate nodules and crusts could represent the first stage in the development of mature calcitic soils which are in the present day more typical of arid to semiarid, oxidizing climates.

Paleoenvironmental characterization

In the context of the interpretation provided for the Tobas Angostura unit, these deposits would represent the gradual transition from conditions of relatively continuous aggradation where streams flowing in extensive broad plains were choked with volcanic debris (Tobas Angostura unit) to conditions where normal stream flow processes began to prevail. The Río de los Pozos Formation is interpreted to record at the Quebrada de la Angostura locality deposition and transportation of sand from unconfined and perhaps broadly channelized flows (lithofacies 1). The occurrence of vertical sequences of superimposed immature paleosols in conjunction with abundant massive beds and evidence of bioturbation suggest slow sedimentation rates where vertical accretion of sediment likely occurred at a slower rate than downward homogenization by pedogenic processes and biogenic activity.

At the El Molino locality, when the contribution of volcanic detritus waned, stream-flow processes gradually regained precedence as indicated by the shift to a conglomerate deposition in fluvial channels (lithofacies 2).

3. PROVENANCE STUDY

Compositions of sandstones and conglomerates primarily reflect the different tectonic settings of provenance terrains and may provide direct evidence of tectonic activity in the source area (e.g., Johnson et al., 1986; Jordan et al., 1988). The evolutionary trend in clast composition of a sedimentary succession showing the appearance of new lithologies and/or a change in the proportion of the constituent particles is indicative of changes in tectonic setting through time or of erosional modification of the source terrains.

(a) Methodology

To elucidate the petrographic evolution of strata from the Mendoza Precordillera, sandstone samples from the Mariño, La Pilona, Tobas Angostura and Río de los Pozos formations were analyzed using standard point-count techniques. On average, samples collected in the studied profiles are spaced about 60-70 m apart up section. The composition and the relative abundance of clasts were estimated qualitatively in the field to recognize parent lithologies and to establish changes in detritus provenance. The petrology and rough estimates of clast type in conglomerate horizons were taken into account when interpreting the source terrains.

The petrography of 35 sandstone samples was established using thin sections stained for potassium and plagioclase feldspar (Houghton, 1980), 34 of which were point-counted using the Gazzi-Dickinson method (Ingersoll et al., 1984). The use of this methodology

minimizes the extent to which varying degrees of weathering of a single source area results in variations in the sand compositions. Quantitative detrital modes calculated from point counts of thin sections were recalculated to 100% as the sum of Q_m (monocrystalline quartz), Q_p (polycrystalline quartz), P (plagioclase), K (potassium feldspar), L_v (volcanic and metavolcanic lithic fragments), and L_s (sedimentary and metasedimentary lithic fragments), as outlined by Dickinson (1985), and these results were plotted in a ternary 'QFL' diagram (i.e., quartz (total), feldspar (plagioclase and K-feldspar) and lithics (total); e.g., Ingersoll et al., 1984). Summary plots are shown in figure 4.3. The stratigraphic position of the study samples and the vertical trends in the suite of grains are shown in figures 4.4 and 4.5. Point-count data for the studied thin sections are found in table A2 (Appendix).

Criteria for recognizing the different grain types were based on the work of Dickinson (1970, 1985) and Ingersoll et al. (1984). In counting, the most relevant variables were documented as follows. (a) Fine-grained polycrystalline, micro-crystalline and cryptocrystalline quartz were classified as Q_p only when they were considered pure silica. (b) The slightest amount of impurity within a chert grain resulted in this grain being classified as L_s . (c) A single flake of primary mica within a polycrystalline quartz grain made it L_s (metasedimentary lithic). (d) A single feldspar microcryst within a siliceous volcanic fragment made it L_v . It is worth noting that there was also a separate category of unidentified grains which, in most cases, included altered fine-grained and finely crystalline fragments which could be chert or either a volcanic or sedimentary fragment. In most cases, a total of 400 points were counted per slide; however, when considered appropriate, 300 or 500 points were counted.

Fig. 4.3. (a) Composition of the studied samples, according to the classification of Folk, Andrews, and Lewis (1970). Q = monocrystalline and polycrystalline quartz; F = plagioclase and potassium feldspar; L = rock fragments. Numbers refer to samples of the Tosca de Mariño (1-13) and Arroyo del Agua Blanca (14-17) sections, and letters refer to samples of the La Higuerita (a-n) and Quebrada de la Angostura (o-q) sections. (b) Sandstone modal data plotted on QFL tectonic discrimination diagram defined by Dickinson et al., 1983.

Fig. 4.4. Provenance study data, Tosca de Mariño (a) and Arroyo del Agua Blanca (b) sections. L = lower, M = middle, and U = upper member; LP = La Pilona Formation.

Fig. 4.5. Provenance study data, La Higuera (a) and Quebrada de la Angostura (b) sections. TA = Tobas Angostura unit, RP = Río de los Pozos Formation.

Due to the friable nature of the sandstones, the preparation of thin sections involved impregnation procedures with an epoxy resin to preserve the entire sample.

(b) Sandstone petrography: trends and source terrains

Sandstones collected from Mariño and La Pilona formations are feldspathic litharenites and lithic feldsarenites according to Folk et al.'s 1970 classification (fig. 4.3a). Some samples from the La Pilona unit have a matrix representing 10-15% and, therefore, are transitional to wackes. Samples from the Tobas Angostura and Río de los Pozos formations are feldspar-rich tuffitic sandstones, which on the ternary QFL diagram for arenites (Folk et al., 1970) plot in the field of the feldsarenites transitional to lithic feldsarenites (fig. 4.3a). Samples are composed mostly of volcanic-derived detrital clastic grains. The framework grains consist mostly of volcanic lithics and plagioclase whose relative abundance and composition vary throughout the measured profiles; in most samples, they represent, when counted together, at least ~75% of the framework grains. Typically, the groundmass of volcanic rock fragments is altered to some extent (e.g., oxidized, chloritized, sericitized, calcitized). In particular, andesitic and basaltic rock fragments sometimes show a severe oxidation. Monomineralic grains that exceed 5% abundance are largely dominated by plagioclase, with lesser quantities of quartz. Quartz makes up about 6-15% of the framework grains. Potassium feldspar is rare, and when present, represents 1-2%. Pyroxenes, amphiboles and micas, when counted together, represent 1-6% in samples of the Mariño and La Pilona formations and up to ~20% of the framework grains in samples from Tobas Angostura and

Río de los Pozos formations. Opaque minerals make up ~1-3%.

Matrix grains and cement consist respectively of detritus of mixed clay and silt component and of patches of visible crystalline precipitated carbonate. In many samples, the matrix is formed of vermicular stacks of kaolinite, sometimes with fine chloritic material. In sandstones from the eolian member of the Mariño Formation, opal and carbonate patches are found as cements, the former probably resulting from devitrification of glass fragments and shards. The occurrence of iron oxide stains sometimes masks the character of the interstitial material. The most relevant petrographical data from all the studied formations are summarized in table 4.1.

Sandstone modal data plotted on the QFL tectonic discrimination diagram (Dickinson et al., 1983) show that the studied samples fall within a range of fields from undissected to transitional arc (fig. 4.3b). This result suggests that sediment petrography within the basin was controlled by arc tectonics and magmatic activity. These results are in agreement with the fact that the Mesozoic and Cenozoic evolution of the Andes was characterized by the development of a magmatic arc, which was located to the west of the studied sedimentary basin, in the Chilean and Argentine Principal Cordillera (Ramos, 1988; Mpodozis and Ramos, 1990). The occurrence of clast lithologies of unambiguous parentage with the Mesozoic and Tertiary volcanic successions of the High Cordillera provide clear evidence of the source terrains. Miocene volcanic activity overlapped in time with deposition of all the studied stratigraphic units. The volcanic rocks of the Complejo Volcánico Aconcagua in the Principal Cordillera record episodes of volcanic activity and emission of pyroclastic material from ~15.8 to ~8.5 Ma (cf. Ramos et al., 1996a; this study), and the sandstone petrography records the

Table 4.1. Summary of the most distinctive petrographical data for each of the stratigraphic units studied

Framework grains	Mariño Formation ($Qt_{10.3}F_{37}L_{52.5}$)		La Pilona Formation ($Qt_{14.5}F_{26}L_{59.5}$)		Tobas Angostura - Río de los Pozos formations ($Qt_{8.5}F_{70.3}L_{21}$)	
	Description	%	Description	%	Description	%
Quartz (Qm) monocrystalline (Qp) polycrystalline	Qm: commonly far more abundant than Qp; with undulose (<5°) and non-undulose extinction; many grains limpid with embayments, bi-pyramidal shape, vacuoles or conchoidal fractures suggesting volcanic source; sometimes grains strained (tilt walls) Qp: grains usually with 2-4 inequant constituent crystals; scarce chalcidonic rock fragments & even more rarely >8 crystals showing sutured grain contacts	Qm 5-13 Qp 0.5-4	Qm: more abundant than Qp; grains commonly clear, with non-undulose or weak extinction; many grains with features suggesting volcanic origin (e.g., embayments) Qp: includes grains with ~2-4 unstrained crystals & grains with >5 crystals usually elongated with smooth or sutured grain contacts, likely of metamorphic origin (cf. Basu, 1985)	Qm 1.5-20.5 Qp 0-7	Qm: most grains clear, commonly strain-free with conchoidal fractures, abundant embayments, & unitary extinction Qp: almost absent	Qm 2-10 Qp 0-0.5
Feldspar (P) Plagioclase (K) K-feldspar	P: much more abundant than K; grains either fresh or altered to sericite, calcite and kaolinite; altered grains possibly derived from volcanic lithics; typically with twinning; many grains with normal or oscillatory zoning; usually found as subangular broken crystals, with compositions ranging from oligoclase to labradorite, and rarely albite, the latter with exsolution features K: includes altered orthoclase, rare sanidine & microcline; a few perthitic grains	P 18-42 K 0-3	P: far more abundant than K; grains fresh, altered to sericite or chlorite and/or partially replaced by calcite; commonly with twinning and/or zonation; grains have variable compositions, with most grains falling in the oligoclase & andesine fields (4 samples) K: includes orthoclase with minor or severe alteration & microcline	P 9-37 K 0.5-3	P: mostly fresh, unaltered & zoned in samples from Tobas Angostura unit; altered and unaltered grains in Río de los Pozos unit; range from large euhedral compositionally zoned crystals to subangular broken crystals; commonly showing twinning; variable composition but mainly in the oligoclase & albite-oligoclase field K: almost absent	P 49-68 K 0-0.5
Lithic fragments	Lv: dominant, include mesosilicic, silicic & basic volcanics; most abundant are volcanics with pilotaxitic, hyalopilitic & trachytic patterns of microlites, very often with large		Lv: dominant, includes abundant rock fragments with felsitic textures that may contain phenocrysts of quartz, feldspar & opaque minerals (rhyolites); lesser amounts of		Lv: dominant, include rock fragments with felsitic textures, mainly rhyolites (lavas, tuffs) sometimes with quartz, feldspar & hornblende phenocrysts,	

Table 4.1. Concluded.

Framework grains	Mariño Formation ($Qt_{10.3}F_{37}L_{92.3}$) Description %	La Pilona Formation ($Qt_{1.4}F_{28}L_{99.5}$) Description %	Tobas Angostura - Río de los Pozos formations ($Qt_{8.3}F_{70.3}L_{21}$) Description %
Lithic fragments (Lv) Volcanics (Ls-m) Sedimentary, low-grade metamorphic rocks (Lp) Plutonic rocks	(Continued) P phenocrysts & augite as principal mafic; basic volcanics with interstitial, intergranular & ophitic textures with granules of olivine, pyroxene & iron ore; felsitic grains sometimes with quartz and/or feldspar phenocrysts (lavas & tuffs) Lv 35-66 Ls-m 0-5 Lp: low abundance; include quartz & feldspar-rich sandstones, siltstones, polycrystalline quartz grains with rare muscovite and/or chlorite, limestone grains with ooids, peloids or skeletal fragments & chert	(Continued) microlitic, lathwork, & vitric to vitrophiric grains, with the latter including rock fragments with quartz, feldspar crystals, glass shards enclosed in a fine-grained banded ash matrix, & grains with eutaxitic textures Lv 30-67.5 Ls-m 1-17 Lp: include quartz-mica aggregates with quartz showing variable degree of strain (e.g., polygonal textures, subgrains with sutured boundaries), mica-chlorite schists, graywackes, shales or slates, siltstones & limestones	(Continued) vitrophiric grains, glass fragments; microlitic grains often porphyritic (mostly andesites with hornblende); a few grains with lathwork textures Lv 12-20 Ls-m 1-5 Lp: include graywackes, shales or slates, mica-chlorite schists, siltstones & quartz-rich sandstones Lp: very rare
Accessory minerals (M) Mono-mineralic grains (O) Opaques	M: include pyroxene (mainly augite, rarely diopside) & amphibole (hornblende & rare lamprobole) O: include magnetite (most abundant) & minute hematite grains M 2-6.5 O 0.5-3.5	M: hornblende, augite, & scarce biotite O: magnetite most abundant M 0-9.5 O 0.5-5	M: abundant biotite & fresh or slightly altered hornblende; rare apatite O: very large amount of magnetite M 6-20 O 0-1

Notes. Mariño Formation: (1) Coarse-grained sandstone from lower member (Purple conglomerates) is comprised 80% of volcanic lithics, mainly porphyritic andesites with augite as principal mafic constituent. (2) Samples from middle member (colian sandstones) have an ash-rich matrix with abundant glass shards and contain glass rock fragments. (3) Sand-size limestone clasts first occur in the upper member. (4) The relative proportion of acidic to mesosilicic-basic volcanics increases upsection. (5) Augite is more abundant than hornblende throughout but the latter increasingly more abundant upsection.

La Pilona Formation: Stratigraphically upwards, samples show (1) an increase in abundance of Ls-m grains at ~120 m above the base of the unit, (2) an increase in the relative proportion of acidic compared to mesosilicic-basic volcanic rock fragments (Lv grains are almost exclusively of acidic composition at the top of the unit), (3) a general increase in the amount of mono- and polycrystalline quartz grains with a decrease in P grains to approximately equal amounts above the ~300 m level (above the base), (4) hornblende more abundant than augite and augite decreasing, and (5) biotite appearing towards the top.

Tobas Angostura and Río de los Pozos Formations: these samples are the most feldspar-rich and are highly tuffaceous. Accessory minerals, mainly biotite, are present in more abundance than in samples from the other stratigraphic units.

effects of the volcanic activity and makes it possible to infer the nature of the arc volcanism. Coeval explosive volcanism resulted in abundant air-fall detritus (glass shards, pumice, lapilli, crystals) in the basin, while erosion of this volcanic suite provided abundant rock fragments, either lavas or tuffs. Other controls on the petrography, in addition to the Miocene arc tectonics and nature of volcanism, are the changing composition of the source area (sequential unroofing), the source area relief and climate, and the depositional environment. The impact that the shift of source area composition has on the petrography is more readily assessed than the other controlling factors, in part because of the limited data set.

Mariño Formation

Samples from this unit have an average modal composition of $Qt_{10.5}F_{37}L_{52.5}$ on the QFL ternary plot (fig. 4.3a), with the framework grains dominated by fragments of volcanic rocks, representing 35-55% of the total grains (fig. 4.4). Plagioclase commonly represents the second constituent in order of abundance, but in a few samples it is present in almost equal amount to or is more abundant than volcanic lithic grains. Most of the plagioclase grains are also probably derived from volcanic source units, and their occurrence is interpreted to result from the disaggregation of coarse phenocryst-bearing volcanic rocks into constituent plagioclase grains. Many quartz grains, the third constituent in order of abundance, are derived from volcanic sources (e.g., grains with bipyramidal shape, embayments). However, a smaller population of grains is likely to indicate recycling of sedimentary source rocks as well as erosion of plutonic sources. This inference is supported by the occurrence of clasts of quartz- and feldspar-rich sedimentary rocks and scarce plutonic rock fragments.

The most distinctive petrographic characteristics of the Mariño Formation are summarized as follows. (1) Coarse-grained sandstones at the base of the unit (Purple conglomerates of the lower member) are nearly monomictic, composed of rounded clasts of volcanic rocks of mesosilicic and basic composition which represent ~80% of the total framework grains (plate 4.13). Andesitic grains composed of plagioclase phenocrysts within a microlitic groundmass are the main grain type. The principal mafic constituent of these rock fragments is clinopyroxene augite (plate 4.14). Some lithic grains contain olivine, pyroxene or iron ore in the interstices between feldspars (basic volcanic types). To a much lesser extent, there are volcanic lithic grains of acidic composition. (2) Sandstones from the eolian member have an ash-rich matrix with abundant glass shards. These samples include volcanic lithic grains that are fragments of volcanic glass. (3) In sandstones from the upper member, the relative proportion of acidic to mesosilicic-basic rock volcanic fragments increases upward, most markedly at ~550-600 m (samples 10-11, fig. 4.4). Whereas sample 10 is composed of 80% mesosilicic-basic fragments of volcanic rocks and 20% acidic rocks (of the lithic volcanics counted), sample 11 is comprised of 65% intermediate-basic volcanic rocks and 35% acidic rocks. The increase in acidic volcanic rocks is accompanied by an increase in the content of tuffs of variable composition. In addition, hornblende-bearing andesites, almost absent in sandstones at the base of the upper member (1-2 grains per thin section), become more abundant upsection. (4) Sand-size carbonate fragments first occur at the base of the upper member (sample 6) and rarely exceed 2% of the total grains. Many of these grains could represent calcrete or multiply reworked grains, but skeletal fragments are clearly identified in sample 7 (~90 m above the base) and upward. They consist of a micritic

Plate 4.13. Lithic-rich sandstone at the base of the Mariño Formation (lower member; Purple conglomerates of Truempy and Lhez, 1937). Volcanic lithic fragments of mesosilicic and basic composition are the dominant grain type. Most abundant are andesitic grains composed of plagioclase phenocrysts within a fine-grained or microlitic groundmass. Microlitic grains show pilotaxitic, hyalopilitic and trachytic patterns of microlites. Note the rounded nature of the clasts. (a) Plane-polarized light. (b) Crossed nicols. The field of view is 5.1 x 3.5 mm.

a

b

Plate 4.14. Lithic-rich sandstone at the base of the Mariño Formation (lower member; Purple conglomerates of Truempy and Lhez, 1937). Note that the principal mafic constituent of the volcanic rock fragments is the clinopyroxene augite. Visible crystalline precipitated carbonate occurs as cement. (a) Plane-polarized light. (b) Crossed nicols. The field of view is 2.6 x 1.8 mm.

a

b

limestone, which may show ooids, peloids and/or skeletal fragments and less commonly grainstones (plate 4.15b). Limestone-bearing conglomerates first occur in the study section ~20 m above the base (plate 4.15a). (5) The dominant accessory minerals are augite and, to a lesser extent, hornblende. In sandstones from the upper member, the amount of hornblende increases gradually upward.

Sandstone petrography in conjunction with paleocurrent data indicate that the detrital constituents of the studied sandstones have a westerly to northwesterly derivation from the Principal Cordillera. All the clasts can be matched with geological units known to be present within this morphostructural province (see regional geology, fig. 4.6). Coarse sandstone and conglomerate at the base of the Mariño Formation (Purple conglomerates) are composed dominantly of volcanic lithic grains that are closely correlated with the volcanic Mesozoic deposits outcropping in the Chilean and Argentine High Andes (between 32°29' and 33°19'). Since the first petrographic study of these rocks (Stelzner, 1885), it has been recognized that the composition of this volcanic suite is mainly andesitic, with abundant porphyritic rocks, in particular augite andesites. The predominantly calc-alkaline arc volcanism of Jurassic-Cretaceous age is represented in the Principal Cordillera by greenish, purple and even darker andesites, augitic andesites, basaltic andesites, hornblende-bearing andesites, olivine basalts and, to a lesser extent, rhyolites, which occur as thick lava piles associated with volcaniclastic and pyroclastic deposits and subvolcanic intrusives (Sanguinetti and Ramos, 1993). These rocks are intruded by Lower Miocene granitic plutons (Aguirre et al., 1974; Munizaga and Vicente, 1982) which were exhumed at about 20 Ma and are presently exposed in the Cordillera del Límite region. The occurrence of rare plutonic fragments and some quartz

Plate 4.15. (a) Photomicrograph of a limestone clast collected from a conglomerate deposit located near the base of the upper member of the Mariño Formation. Plane-polarized light. The field of view is 5.1 x 3.5 mm. (b) Note the presence of sand-size limestone clasts in a feldspar-rich sandstone (feldspathic litharenite according to Folk et al., 1970) collected ~180 m above the base of the upper member. Crossed nicols. The field of view is 2.6 x 1.8 mm.

a

b

Fig. 4.6. Map of the regional geology of the Principal and Frontal cordilleras and the Precordillera (modified from Introcaso and Ramos, 1989).

grains may derive from this source.

The source of limestone clasts and quartz- and feldspar-rich fragments of sedimentary rocks is found in the Jurassic to Early Cretaceous marine deposits cropping out in the Aconcagua region (fig. 4.6). These sequences are comprised of limestones with abundant mollusc fossils, micritic mudstones, sandstones, shales and gypsum levels arranged in several sedimentary cycles (e.g., Polanski, 1972; Yrigoyen, 1979; Legarreta and Gulisano, 1989). Clasts of siliciclastic sedimentary rocks may also have been derived from the continental Cretaceous deposits (Diamante and Cristo Redentor formations) cropping out extensively along the boundary between Chile and Argentina (cf., Yrigoyen, 1979; Cristallini and Ramos, 1996) or from the pre-Jurassic basement (Alto Tupungato Formation). The latter unit most probably provided the rare metasedimentary rock fragments and polycrystalline quartz grains.

Hornblende-bearing andesitic grains, rock fragments of tuffs, glass shards, and clasts of pumicites most likely were derived from the Miocene volcanic succession that rests unconformably upon the Mesozoic volcanic sequences in the Aconcagua region (fig. 4.6, Complejo Volcánico Aconcagua; Ramos et al., 1996a). The Tertiary volcanic succession consists of lavas, welded and unwelded tuffs, volcanic breccias and agglomerates; the most abundant rock type is hornblende-bearing andesite (Ramos and Nullo, 1993). A second possible source for hornblende-bearing andesitic grains could be found in the Mesozoic volcanic deposits. The likelihood that these clasts were derived from the Tertiary sequences stems from the fact that pumicites and tuffs exhibit a similar trend in abundance suggesting that the eroded source terrain was the same. The source of rhyolitic clasts may be found in the Jurassic-Cretaceous volcanic sequences and in the Permo-Triassic sequences of the

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Choiyoi Group. These rocks, which are widely exposed in the Frontal Cordillera (Caminos, 1979), represent the pre-Jurassic basement in the Principal Cordillera that is unconformably overlain by Mesozoic marine sequences (e.g., Ramos and Yrigoyen, 1987; Cegarra et al., 1993).

La Pilona Formation

Sandstones from the La Pilona Formation have an average modal composition of $Qt_{14.5}F_{26}L_{59.5}$ on the QFL ternary plot (fig. 4.3a), with the lithic framework grains dominated by volcanic lithic grains of acidic and, to a lesser extent, mesosilicic and basic composition. In most samples, volcanic lithic grains represent 40-55% of the total framework grains (fig. 4.5) and include rhyolitic and andesitic grains, vitric to vitrophyric rock fragments and, to a much lesser extent, fragments of basic volcanics. Vitrophyric grains include grains with crystals and glass shards enclosed in a fine-grained banded ash matrix, grains with eutaxitic textures, and fragments composed of zeolite aggregates, interpreted as the devitrification product of glass. Feldspars, mostly plagioclase, represent the second or third most abundant type of grain, in some samples as abundant as quartz grains. Lithic sedimentary and metasedimentary grains make up to 17% of the framework grains. They include quartz-mica aggregates with quartz showing variable degrees of strain, mica-chlorite schists, quartz- and feldspar-rich graywackes, micaceous fine-grained rocks (shales or slates) and siltstones. Compared with the Mariño Formation, sandstones of this unit show more abundant clasts of sedimentary rocks, and volcanic lithic grains are dominantly of acidic composition.

Stratigraphically upward, samples of the La Pilona Formation show the following:

(1) The relative proportion of acidic compared to mesosilicic-basic volcanic rock fragments increases. In sandstones from the top of the succession, rhyolitic grains and, to a much lesser extent, fragments of welded tuffs are almost the only type of volcanic lithic grains. (2) The abundance of sedimentary and metasedimentary grains increases. (3) The number of monocrystalline and polycrystalline quartz grains generally increases and the number of plagioclase grains decreases to approximately equal amounts above the ~300 m.

This compositional change is most marked at ~190 m above the base of the formation (sample f, plates 4.16 and 4.17). If sample f is compared with stratigraphically lower samples, there is approximately a three-fold increase in monocrystalline quartz and a seven-fold increase in polycrystalline quartz, the latter likely derived from low-grade metamorphic and sedimentary rocks. Quartz grains with ~5 or more subgrains, in some cases with sutured boundaries, are mostly found in sample f and above it and most probably are of metamorphic origin (cf. Basu, 1985). The abundance of fragments of sedimentary-metasedimentary rocks increases from ~1-5% of total grains to ~17%. Furthermore, volcanic lithic grains are dominated by mesosilicic types of lava in samples below 120 m (e.g., sample e: 65% mesosilicic, 22% acid, 12% basic, 1% fragmentary rocks of andesitic? composition, over 100 grains), whereas felsitic grains of rhyolitic-dacitic composition often with phenocrysts of feldspar and quartz represent 65-70% of the volcanic lithic grains in sample f. The decrease in the abundance of plagioclase from ~35% to ~9% most probably correlates with the marked decrease in content of coarse phenocryst-bearing volcanic rocks (mostly andesites) which yield plagioclase grains as a result of disintegration.

An estimation of the composition of conglomerates in three beds situated at

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Plate 4.16. Sandstone from the La Piona Formation (sample f in fig. 4.5). Note the presence of sedimentary-metasedimentary rock fragments including mica-chlorite schists, quartz-rich aggregates (orthoquartzite?) and feldspar-rich rock fragments; the occurrence of silicic volcanic rock fragments, either lavas or tuffs, with cryptocrystalline or felsitic textures (microcrystalline mosaic of quartz and feldspar), sometimes with a salt-and-pepper appearance; and presence of rock fragments with crystals embedded in a fine-grained ash matrix sometimes with flattened glassy fragments. Mesosilicic to basic volcanic rock fragments often contain phenocrysts within a fine-grained groundmass often partially or totally stained by iron oxides. Visible crystalline precipitated carbonate occurs as cement. (a) Plane-polarized light. (b) Crossed nicols. The field of view is 2.6 x 1.8 mm.

a

b

Plate 4.17. Sandstone from the La Pilona Formation (sample f in fig. 4.5). Note the presence of low-grade metamorphic and sedimentary rock fragments, such as a chlorite-bearing quartz-rich aggregate showing evidence of quartz recrystallization illustrated by subgrain boundaries and alignment of chlorite (bottom), a muscovite-rich pelitic schist (left) and a matrix-rich sandstone, likely a graywacke (upper right); and the occurrence of fragments of silicic volcanic rocks, either lavas or tuffs, sometimes with quartz phenocrysts embedded in a fine-grained matrix (middle to upper right) which originally may have been glassy. Very often, volcanic rock fragments show crypto- to micro-crystalline textures. When viewed under higher magnification, quartz and feldspar are identified in the coarser crystalline grains. Visible crystalline precipitated carbonate occurs as cement. (a) Plane-polarized light. (b) Crossed nicols. The field of view is 1.3 x 0.9 mm.

a

b

approximately ~140, 180 and 185 m indicates that 70-75% of the clasts are dominantly acidic volcanic rocks. The most common rock types include red rhyolites with phenocrysts of quartz and sometimes feldspar, biotite and/or hornblende, dacites with hornblende and whitish tuffs. Subordinately, there are andesitic rocks, some of them with large phenocrysts, and aphanitic dark volcanics. The remaining portion includes leucocratic granites, different types of sedimentary and metasedimentary fragments, such as graywackes, green and whitish quartzites, quartz-rich sandstones with mica, a few phyllites and limestones, green shales and slates, and clasts of milky quartz.

There is a marked increase in the proportion of hornblende and biotite ~650 m above the base of the unit (sample j), with a near absence of biotite in stratigraphically lower samples. Many of these crystals are fresh and some of them are unbroken; most probably, they are derived from primary or reworked volcanic ashes. This interpretation gains support from the occurrence of ash-fall horizons and the highly tuffaceous nature of the rocks in this part of the La Pilona stratigraphy. In addition, sandstone from the uppermost part of the La Pilona Formation (sample l) is more feldspar and biotite-rich, a signature that characterizes the petrographic suite of the overlying Tobas Angostura and Río de los Pozos formations.

The dominance of volcanic-derived detrital grains in sandstones of the La Pilona Formation indicates that the potential source terrains are (1) the Permo-Triassic volcanic succession of the Choiyoi Group, (2) the Jurassic-Cretaceous volcanic complex of the Principal Cordillera, and (3) the Miocene volcanics of the Complejo Volcánico Aconcagua (see regional geology, fig. 4.6). The latter two geological units are mainly composed of mesosilicic volcanic rocks (cf. Sanguinetti and Ramos, 1993; Ramos et al., 1996a) and most

likely provided microlitic and lathwork grains and rock fragments of tuffs.

Visual identification of clasts of rhyolite and granitic rock in conglomerates of the La Pilona Formation at ~140 m and upward and the dominantly acidic composition of sand-size grains above ~190 m allow their correlation with the volcanics of the Choiyoi Group, widely exposed in the Frontal Cordillera (Camino, 1979). The source rock sections consist mostly of a thick pile of rhyolites and associated granites (Kay et al., 1989; Llambías and Sato, 1990), which reaches a thickness of up to 2-4 km in the Cordillera del Tigre (Cortés, 1985). Paleocurrent data indicate a northwestern provenance, which in conjunction with the subrounded nature of the clasts suggesting a relatively long transport allow for inference that the source area is most likely the Cordón del Tigre region. The sedimentary and low-grade metamorphic clasts found in sandstones of the La Pilona Formation can be correlated with the pre-late Paleozoic units present in that region (Cordón del Tigre and Sandalio). The sequences of Lower Paleozoic age described by Koukharsky (1965), Coira (1965) and Amos and Roller (1965) consist of fine-grained rocks of low-metamorphic grade, shales and graywackes which are overlain by a thick Carboniferous succession consisting of sandstones, conglomerates with clasts of metamorphic rocks and mudstones.

La Pilona Formation records events of explosive volcanic activity in the Principal Cordillera as locally recorded by crystal-rich fallout tephras. Air-fall processes may explain the enrichment in biotite, fresh plagioclase and glass shards observed in sandstones from the upper part of the La Pilona Formation.

Tobas Angostura and Río de los Pozos formations

Due to the gradational transition between these two formations and similarities in the modal composition and grain constituents, samples from these two units are described together (plates 4.18 and 4.19). They are tuffites with an average modal composition of $Qt_{8.5}F_{70.5}L_{21}$ on the QFL ternary plot (fig. 4.3a). Some samples show a chloritized clay matrix, which probably resulted from the alteration of volcanic ashes deposited in the basin (plate 4.19). These rocks are dominantly comprised of plagioclase, which represents ~60% of the total framework grains (fig. 4.5). Plagioclase in samples from the Tobas Angostura unit is typically fresh and unaltered and ranges from large, euhedral, compositionally zoned crystals to subangular broken crystals commonly showing twinning (plate 4.18). These features together suggest deposition from air-fall processes with little reworking. Quartz and volcanic lithic grains are present in subequal quantities representing ~10% each. Lithic volcanics consist of rock fragments of rhyolites, hornblende-bearing andesites and tuffs. Accessory minerals represent ~15-20% of the total population of grains, the most abundant of which are unaltered hornblende and biotite. Sedimentary and metasedimentary rock fragments are present in low abundance (1-5%) and are similar to those described in the La Pilona Formation. There are scarce fragments of plutonic rocks containing biotite as a mafic constituent.

Conglomeratic beds at the base of the Tobas Angostura unit in a few localities of the western limb of the La Pilona anticline include clasts of reddish rhyolites with abundant phenocrysts of quartz, whitish tuffs, green acid ignimbrites and volcanic breccias, pumicites, porphyritic rocks of intermediate composition (andesites and trachytes?), quartz- and/or

Plate 4.18. Feldspar-rich tuffite from the Tobas Angostura unit. Plagioclase is typically fresh, unaltered and twinned and ranges from large, euhedral, compositionally zoned to subangular broken crystals. Accessory minerals are hornblende and biotite. Volcanic rock fragments are of silicic composition and include grains with felsitic textures (microcrystalline aggregates of quartz and feldspar, bottom to the right), porphyritic textures (e.g., large grain, bottom left) and vitric to vitrophyric grains which may show alignment of flattened glassy fragments, and phyllosilicates as an alteration product (e.g., bottom). (a) Plane-polarized light. (b) Crossed nicols. The field of view is 2.6 x 1.8 mm.

a

b

Plate 4.19. Feldspar-rich tuffite from the Río de los Pozos Formation. Note the abundance of twinned, sometimes zoned, broken plagioclase crystals; the occurrence of sedimentary and metasedimentary rock fragments, such as a pelitic schist (upper left) and a fine-grained rock composed of quartz, feldspar and mica as main constituents (center); the presence of volcanic rock fragments, sometimes with larger quartz phenocrysts and opaques, some of them resulting from alteration of hornblende. Large biotite and green hornblende are the common accessory minerals. The rock sample has a chloritized clay matrix, which probably resulted from the alteration of volcanic ashes. (a) Plane-polarized light. (b) Crossed nicols. The field of view is 2.6 x 1.8 mm.

a

b

feldspar-rich sandstones, and fine-grained graywackes, sometimes containing clasts of milky quartz. There are also green fine-grained sedimentary or metasedimentary rocks, which include shales, slates, hornfels?, a few of them with quartz veins and small-scale folding, and a few clasts of phyllite.

Conglomerates of the Río de los Pozos Formation in the El Molino section are composed of approximately ~65-70% volcanic rocks of acidic and, to a lesser extent, mesosilicic composition, and the remaining portion includes sedimentary and low-grade metamorphic rocks similar to those described above and massive limestones.

These two units are the most feldspar-rich in the study area. The increase in the abundance of plagioclase relative to volcanic lithic grains and the noticeable increase in biotite, pyroclastic detritus (glass shards, pumicites) and hornblende-bearing andesitic grains, when compared with sandstones of the La Pilona Formation, are thought to be related to volcanic activity in the Principal Cordillera. Air-fall processes in combination with erosion of the volcanic suite provided the basin with this detritus. In fact, the Tobas Angostura unit is correlative with the latest explosive volcanic episodes as inferred from the age of these deposits in the study area (cf. Chapters 2 and 3) and radiometric dates from the Complejo Volcánico Aconcagua (cf. Ramos et al., 1996a).

As with the sandstones of the La Pilona Formation, rhyolite grains and sedimentary and low-grade metamorphic rock fragments are interpreted to be derived from the Choiyoi Group and the pre-late Paleozoic units outcropping in the Frontal Cordillera. Andesitic clasts with large phenocrysts most likely are derived from the Mesozoic volcanic sequences of the Principal Cordillera (V. Ramos, personal communication, 1993) and the fossiliferous

limestone clasts from the marine Mesozoic deposits cropping out in that region.

4. SEDIMENT ACCUMULATION RATE

As the magnetic reversals recorded by the local stratigraphic sections have been calibrated with respect to time (see Chapters 2 and 3), it is possible to estimate the mean rate of sediment accumulation for each. Sediment accumulation rates are derived by plotting sample stratigraphic position versus age. The slope of the derived curve is directly proportional to the sedimentation rate.

Sediment accumulation rate calculated by regression techniques for strata of the Mariño Formation is 0.22 mm/a ($R^2 = 0.99$). The regression plot for the Tosca de Mariño section (fig. 4.7) shows minor variation in the sediment accumulation rate in the short term but is remarkably constant over the long term. It should be noted that part of the apparent short-term variation in sedimentation rates may be an artifact of the sampling procedure.

In the La Higuera section, the La Pilona Formation shows a marked increase in sedimentation rate with time. The mean accumulation rate for the lower part of the unit is 0.17 mm/a. It increases progressively to 0.33 mm/a, 0.77 mm/a, and 0.95 mm/a in the upper part of the formation (fig. 4.8a). Therefore, there is an approximately six-fold increase in sedimentation rate between the lower and upper parts of the La Pilona Formation.

There are not sufficient data points to establish the sedimentation rate corresponding to the lower part of the La Pilona Formation in the Arroyo del Agua Blanca section (fig. 4.8b). However, assuming that most of the long normal chron C5n is preserved, the derived

Fig. 4.7. Comparison of the 'Tosca de Mariño' magnetic polarity stratigraphy (fig. 3.6) with Baksi (1993) time scale. Slope is directly proportional to sedimentation rate.

Fig. 4.8. Comparison of the (a) 'La Higuera' magnetic polarity stratigraphy (fig. 3.17) and (b) 'Arroyo del Agua Blanca' polarity stratigraphy (fig. 3.13) with Baksi (1993) time scale. Slope is directly proportional to sedimentation rate. Note that the 'La Higuera' (HI) section represents a greater portion of the La Pilona Formation than the 'Arroyo del Agua Blanca' (AB) section. In the HI section, the open circle shown at the unconformity (above which lies the Tobas Angostura unit) assumes that no significant portion of the section is missing. Otherwise, the sedimentation rate would be higher. In the AB section, the lower accumulation rate value represents a minimum sedimentation rate because the local section may not record the bottom of the long normal magnetozone.

sedimentation rate is similar to that of the La Higuera section (fig. 4.8a). The sedimentation rate calculated for the lower part of the La Piona Formation in the Arroyo del Agua Blanca section (0.17 mm/a) represents a minimum estimate of the mean accumulation rate because rocks at the base of the La Piona Formation may not record the bottom of chron C5n. As in the La Higuera profile, the Arroyo del Agua Blanca section shows a marked increase in accumulation rate from ~0.17 mm/a to 0.91 mm/a.

For the Quebrada de la Angostura section, a sediment accumulation rate of 0.41 mm/a is estimated assuming that sediments of the Tobas Angostura Formation started accumulating at the beginning of chron 4An (fig. 4.9). This rate represents a minimum estimate of the mean accumulation rate for the section.

As outlined by Jordan et al. (1988), measured rock thickness incorporates four components that create space for deposition: tectonic subsidence, sediment compaction, isostatic lithospheric subsidence due to sediment loading, and change in elevation of the depositional surface. The study of Jordan et al. (1988) showed that relative accumulation rates are not greatly altered by decompaction analysis, as a result of which compaction rates were not manipulated in this study. Even if corrected, the resulting rates would not record the impact that the Pliocene-Pleistocene? Mogotes Formation (as thick as the rest of the Tertiary units), had on the stratal compaction history of the underlying Neogene strata. Similarly, lack of sufficient data precluded correction for isostatic subsidence and elevation change (for details of these corrections, complications and pitfalls, see Jordan et al., 1988). Therefore, interpretation of data is based only on the shape of the 'relative subsidence rate' curves derived from the local sections and not the true tectonic subsidence.

Fig. 4.9. Comparison of the 'Quebrada de La Angostura' magnetic polarity stratigraphy (fig. 3.23) with Baksi (1993) time scale. Slope is directly proportional to sedimentation rate. The accumulation rate value represents a minimum sedimentation rate because the local section may not record the bottom of the normal magnetozone.

5. INTERPRETATION OF DATA

The combined data concerning depositional environment, age and timing of facies changes within units, rates of accumulation and provenance of the detritus (summarized in table 4.2) result in an improved understanding of the history of thrusting in the Principal and Frontal cordilleras. In addition, chronological data provide age constraints on the history of deformation of the southernmost part of the Precordillera thrust belt.

(a) Mariño Formation

Interpretation of the magnetic stratigraphy indicates that deposition of this unit started at ~15.7 Ma and continued to ~12.2 Ma (cf. Chapter 3, figs. 3.6 and 3.28). The beginning of the accumulation of the lower member is marked by fluvial gravels, interpreted to record a phase of thrust activity and exhumation of the andesitic Mesozoic volcanic complex cropping out far to the west, in the Principal Cordillera. Because ages of influx of conglomeratic sequences vary between locations in a basin and none of them provides a precise age for the causative thrusting event (Jordan et al., 1988), it can only be asserted that prior to or at about 15.7 Ma there was an episode of deformation in the High Andes. Whether or not these gravels accumulated in a period of relative tectonic quiescence or were syntectonic indicators of a phase of thrusting will be discussed below.

Deformation that began prior to ~15.7 Ma overlapped in time with episodes of volcanic activity in the Cordilleran region. The noticeable tuffaceous nature of the Mariño

Table 4.2. Summary of most relevant data of the Neogene deposits and interpretation. (1) Based on sandstone petrography complemented with qualitative data from conglomerate deposits; first appearance of diagnostic lithologies in *italic bold*; marked increase in relative abundance in **bold**. Listed are only the most common clast types. (2) Derived from measured rock thicknesses; (*) inferred to be less than for Tobas Angostura unit based on sedimentological evidence (see text). (3) Age is likely as young as or younger than 7.2 Ma (cf. chapter 3). (4) According to terminal fan model of Kelly and Olsen (1993). TA = Tobas Angostura unit; RP = Río de los Pozos Formation; cu = coarsening upward trend; Pz = Paleozoic; P-Tr = Permo-Triassic; Jr = Jurassic; K = Cretaceous; sed. = sedimentary; mm. = metamorphic.

strata above the basal ~20 m (including a few bentonite horizons), also evident in the eolian member, is correlated with the first eruptive episodes recorded by the Complejo Volcánico Aconcagua (Ramos et al., 1996a) in the Argentine Principal Cordillera. The 15.8 ± 0.4 Ma and 13.7 ± 1.8 Ma K-Ar whole-rock dates obtained from the lower section of the volcanic sequences cropping out in the Aconcagua region support this correlation and establish that ~15.7 Ma is a better estimate than the proposed ~13.7 Ma for the beginning of the accumulation of these volcanics (cf. Ramos et al., 1996a). Accordingly, at about 15.7 Ma, tectonogenic sediments were shed into the basin and deposited in a semiarid alluvial plain setting. Most of the lower member (above the basal 20 m) records deposition by clast-poor debris flows which alternated with periods of surface exposure suggesting the dominance of vertical accretion processes. In this environment, local dune fields probably developed. These conditions persisted until ~15.4 Ma.

From ~15.4 to ~14.7 Ma, very limited fluvial input favoured the full establishment of a desert setting with development of extensive dune fields (middle member) indicating that redistribution of sediment by wind was the dominant process for ~700 ka.

The upper member, deposited from ~14.7 to 12.2 Ma, marks the reestablishment of a stream-dominated setting where deposition occurred by different ephemeral stream processes. The lower part of the upper member records sporadic episodes of sheet flooding alternating with periods of sedimentation of overbank fines. To a lesser extent, gravelly streams also contributed to the basin-floor sedimentation. This major change of environment of deposition at ~14.7 Ma is interpreted to have been triggered by a change in basin margin relief far to the west. Fossiliferous limestone clasts first occur in conglomeratic beds placed

stratigraphically ~20 m above the base of the upper member. They are interpreted to be the harbinger of the onset of thrusting and uplift of the Late Jurassic - Early Cretaceous marine sequences outcropping in the Principal Cordillera.

Sequentially in time, there was progressively more intensive activity of the fluvial system in the area, as recorded by vertically stacked sandstone sheets, conglomeratic deposits with variable degrees of channelization, and conglomeratic facies interpreted as low sinuosity braided streams. The provenance study indicates that the period of higher energy flow conditions and/or more continuous discharge, starting at ~13.2 Ma, coincides with an increase in the content of clasts of tuffs and pumicites of variable composition and more abundant hornblende-bearing andesites.

This lithologic assemblage characterizes the Tertiary volcanic pile that rests unconformably upon the Mesozoic marine sequences in the Aconcagua region. According to Cegarra and Ramos (1996), deposition of the basal Tertiary volcanic sequences above a series of folded and thrustured Mesozoic strata occurred shortly after thrusting. Motion of one of the westernmost thrust sheets occurred at approximately 14.3 ± 1 Ma (Ramos, 1996b; Ramos et al., 1996a; Cegarra and Ramos, 1996). Therefore, it is suggested that the increase in content of clasts of pyroclastic rocks and andesites is related to uplift and subsequent erosion of any of the thrust sheets where the Tertiary volcanics unconformably overlie thrustured Mesozoic strata. In addition, the highly tuffaceous nature of the uppermost part of the Mariño succession also suggests contemporary volcanic activity in the Cordilleran region, which could be tied to the extrusive episodes known to occur between ~13.7 and ~11.3 Ma (Ramos et al., 1996a).

A distinctive feature of the sediment accumulation curve for the Mariño Formation is the remarkably constant sedimentation rate in the long term (~3.5 Ma) despite considerable variations in lithology and sedimentary environment through time. Important facies shifts occurred that were not accompanied by changes in the accumulation rate, and some of these changes are interpreted to be related to thrust activity.

Although the conditions necessary to maintain a steady-state accumulation rate seem unlikely in a tectonically active terrane, the same trend is reported for fluvial deposits in other Andean foreland basins. For example, north of the study area at Sierra de Huaco in San Juan province, a 2800 m thick pile of sediments was deposited at a rate of 0.6 mm/a over 4.4 Ma (Johnsson et al., 1984), and further north at Puerta del Corral Quemado, in Catamarca province, Neogene strata accumulated at a fairly constant rate of 0.5 mm/a over ~4.5 Ma (Butler et al., 1984). In these cases, subsidence rates and the rates of sediment supply might have been in a tight balance that sustained the long-term constant rate of sedimentation. In this particular case, it can also be argued that the La Pilona site may have been at the margin of the flexural wavelength associated with structural activity in the Principal Cordillera. Thus, deformation may not have induced a noticeable change in subsidence rate at that point in the development of the basin but may have changed the drainage networks associated with the development of a thrust front (e.g., Damanti et al., 1988).

Because the accumulation history of the Mariño strata does not show changes in sedimentation rates that may highlight the possibility of a tectonically induced subsidence, the presence of distinctive lithologies and changes in provenance provide invaluable information for dating specific tectonic events. The occurrence of andesite-bearing conglomerates at the

base of the unit and fossiliferous limestone clasts in conglomerates of the upper member permits unambiguous identification of the source terrains, although the position of these thrusts cannot be precisely ascertained. Present distribution of rock types shows that some of the thrust sheets that comprise Jurassic limestones are covered by andesitic Mesozoic volcanics. The absence of limestone clasts at the base of the formation and the nearly monomictic composition of the purple conglomerates are taken as indicating that most likely, prior to ~15.7 Ma, there was active thrusting in what is now the western part of the Aconcagua thrust belt where the Mesozoic volcanic complex is widely exposed. The first occurrence of limestone clasts at 14.7 Ma could be tied to motion of one of the westernmost limestone-based thrust sheets presently situated in the western slope of the Aconcagua. It is possible that the Mariño strata record until ~12.2 Ma a history of motion of several thrust sheets.

Thrust activity is also recognized in the stratigraphy by shifts in depositional environment through time. However, the spatial and temporal correlation between thrust emplacement and the observed variations is not readily assessed. The difficulty in interpretation stems from the fact that (1) depositional facies record the influence of interacting surficial controls (e.g., climate, eustatic base level variations, bedrock lithology) that may obscure the tectonic signal (Jordan et al., 1988; Jordan, 1995), (2) lack of data from other locations in the basin do not allow for inference of which of the possible controls is responsible for the vertical trend of facies, and (3) the depositional response to thrust loading is potentially diverse (cf., Wiltschko and Dorr, 1983; Heller et al., 1988; Burbank et al., 1988; Flemings and Jordan, 1990). Nevertheless, the Mariño strata show an evolutionary trend from

fine-textured deposits to coarser clastic fluvial facies that suggest a tectonically driven environmental deposition (e.g., Miall, 1981; Blair and Bilodeau, 1988).

Several studies have interpreted thrust motions by analyzing lateral facies changes and variations through time. Orogenic events in thrust belts have been dated by the age of conglomeratic sequences in the adjacent foreland basin (e.g., Wiltschko and Dorr, 1983; Burbank et al., 1988), supporting the traditional interpretation that gravels are syntectonic indicators of tectonic activity. In contrast, studies based on the mechanics of subsidence and models of basin filling (e.g., Heller et al., 1988, Paola, 1988) suggest that during times of tectonism and uplift in the thrust belt, coarse alluvial deposits accumulate only in the most proximal part of the basin and grade rapidly into fine-grained deposits toward the distal part of the basin. In this interpretation, the initiation of a phase of thrust activity is linked with a rapid subsidence in excess of sediment supply which favours deposition of the coarsest sediment proximal to the thrust (Paola, 1988). As deformation proceeds and the relief in the thrust belt increases, there is greater erosion and therefore greater sediment supply to the foreland basin (Heller et al., 1988). The theoretical model developed by Flemings and Jordan (1990) and Jordan and Flemings (1990) shows the complex stratigraphy that results from episodic thrust deformation in a nonmarine foreland basin. The results suggest that during continued thrusting and during quiescence, sedimentary facies prograde in the direction of thrust movement. These progradations are punctuated by retrogradations at the onset of thrust cycles, in agreement with Blair and Bilodeau (1988) and Heller et al. (1988), which coincide with the onset of formation of an erosional unconformity in distal parts of the basin.

The evolutionary trend in facies of the Mariño strata suggests a tentative correlation

with the models that stress the importance of thrust emplacement as the mechanism for generating retrograding facies (e.g., Heller et al., 1988; Flemings and Jordan, 1990). As previously pointed out, the beginning of the accumulation of the Mariño strata is marked by fine conglomerates (Purple conglomerates) interpreted to have recorded a phase of thrust activity prior to ~15.7 Ma. These deposits probably accumulated during a period of tectonic quiescence, rather than being syntectonic indicators of thrusting (fig. 4.10). This interpretation is compatible with (1) the observation by Cegarra and Ramos (1996) that there was a phase of thrust activity in the western part of the Aconcagua thrust belt prior to 15.7 Ma, whose onset is thought to have been initiated at ~20 Ma; (2) the provenance study consistent with derivation from the western part of the belt; and (3) the limited stratal thickness of the deposits (~20 m) and grain size, which allow for inference of declining tectonic activity in the thrust belt.

It is hypothesized that the onset of an important phase of thrust activity in the Principal Cordillera coincided approximately with the beginning of deposition of the eolian member at about 15.4 Ma (fig. 4.11). The subsidence induced by uplift and deformation of the Mesozoic marine sequences could explain the change in depositional setting and texture of facies observed between the lower and middle members. The trapping of detritus near the thrust front (e.g., Paola, 1988) is thought to have resulted in limited fluvial input in more distal parts of the basin, which in turn favoured the development of extensive dune fields. Following the initial rapid subsidence and as deformation proceeded, the increasing relief in the thrust belt resulted in a greater sediment supply. The shift in depositional facies at about 14.7 Ma from an eolian environment to a stream-dominated setting could represent the stage

Fig. 4.10. Schematic diagram (modified from Jordan, 1995) showing a tentative interpretation for the tectonically derived Purple conglomerates of the Mariño Formation. These deposits are thought to record a phase of thrust activity prior to ~16 Ma and are interpreted to be deposited during a period of tectonic quiescence in the thrust belt. The horizontal scale is approximate and is based on palinspastic reconstructions (Cegarra and Ramos, 1996).

Fig. 4.11. Schematic diagram showing a tentative interpretation of the thrust history in the Principal Cordillera as recorded by the Mariño Formation. In this interpretation, the onset of an important phase of deformation in the Aconcagua fold-and-thrust belt is linked with a major shift in depositional facies at 15.4 Ma, coincident with the beginning of accumulation of the eolian member (for explanation, see text). The occurrence of fossiliferous limestone clasts in conglomerates at the base of the upper member (~14.7 Ma) is taken as indicating that the thrust terrain is the Mesozoic marine deposits cropping out in the Aconcagua region. It should be noted that thrusting continued until after ~12.2 Ma, the age of the youngest Mariño dated horizon, probably until ~8 Ma (Ramos et al., 1990, 1991, 1996a).

of active thrusting shown by the model of Jordan and Flemings (1990) where the sediment flux overwhelmed the subsidence adjacent to the thrust, and facies prograde farther out across the basin. The vertical trend of increasingly coarse debris in the upper member suggests alluvial fan progradation. Assuming this interpretation is valid, the occurrence of limestone clasts in conglomeratic beds at ~14.7 Ma implies a thrust event that started at about 15.4 Ma and is marked by a retrograding facies (eolian succession) at the study site (see fig. 4.11).

Flemings and Jordan (1990) stress the importance of the erosional unconformity that develops at the onset of thrust cycles in distal parts of the basin. Although the lower and upper contacts of the eolian member appear to be concordant, the nature of these surfaces is rather ambiguous, and some workers suggested that the middle member is bounded at its base and top by unconformities (e.g., Truempy and Lhez, 1937). Magnetostratigraphic results suggest that if there were a hiatus bounding the lower and/or upper boundaries, it would span a short period without significant erosion. It should be mentioned that it is possible that the unconformity occurs at the conformable-appearing contact between the Purple conglomerates and the upper ~45 m of the lower member.

To conclude, it is surmised that accumulation of eolian strata occurred under an increased subsidence where tectonism played an important role in the preservation of these sediments. It is worth mentioning that eolian sandstones of approximately similar age to that of the Mariño Formation are known to occur north of the study area, in the Precordillera of San Juan; these deposits (~20 to 14 Ma) are thought to be coeval with thrusting in the western part of the Precordillera thrust belt and are considered to represent the first stage of fill of the Bermejo foreland basin (Jordan et al., 1993 a,b). Milana et al. (1993) proposed a

tentative correlation between the ~700 m thick syn-tectonic eolian succession cropping out in the Central Precordillera (Pachaco Formation) and the Mariño eolian strata, suggesting that there was a regional climatic control for the development of these accumulations. However, this correlation hypothesis needs to be tested since the eolian sequence described by Milana et al. (1993) has yet to be dated. Because evidence of arid conditions is found throughout the entire Mariño Formation, the interpretation favoured here is that the evolutionary trend within the Mariño stratigraphy is controlled mainly by tectonism rather than solely autogenic factors.

(b) La Pilona Formation.

Deposition of the La Pilona Formation started at or slightly before ~11.7 Ma and continued to ~9 Ma (cf. Chapter 3). The association of sheet-flood, stream-flow, debris-flow, and suspension-dominated-flow deposits including thick playa clay sequences allows for inference that these deposits accumulated in a low-gradient alluvial setting, probably at the very toe of extensive bajadas that prograde onto a flat flood basin or playa mud flat. Concurrent lines of evidence suggest that these deposits accumulated in a region characterized by arid or semi-arid climatic conditions and that the drainage system had a regime of highly variable discharge, probably ephemeral.

From ~11.7 Ma to ~10.0 Ma, the La Pilona Formation records an overall gradual transition from relatively fine-textured fluvial deposits to coarser fluvial lithofacies defining a gross coarsening and thickening upward trend. Vertically stacked sheet-flood sandstone units and siltstones at the base of the section are replaced upward by thick sequences of

alternating gravel and sandstone sheets, which are a result of mixed-load deposition in channels and sheet floods. These sequences are interpreted to record sedimentation in a distal braid - plain setting bordering a playa lake where shallow braided channels transported and deposited gravels, mostly volcanic rocks (Choiyoi Group), from northwestern highland sources. The present exposure reveals that, sequentially through time, there was less sediment input at this location in the basin. The occurrence of thinner sand-dominated successions and thicker intervals of mudstones with evidence of soil formation (incipient calcretes) suggests the existence of relatively longer periods of subaerial exposure. The dominance of quiescent conditions increases progressively up section between ~9.8 and 9.3 Ma as inferred from the occurrence of extensive and thick sheet-like mudstone units. In this flat environment, temporary and semi-permanent? shallow lakes developed, as inferred from the occurrence of green laminated mudstones with thin horizons of gypsum. Following ~9.3 Ma, there was a return to higher energy conditions related to stream-flow processes. The shift upward from fine-textured facies to coarser deposition stems from progradation, where increasingly coarse debris derived from the northwest was transported by streams with braided morphologies.

A remarkable feature of the sediment accumulation curve for the La Pilona Formation is the dramatic increase in sedimentation rate with time, most marked at ~9.7 Ma. The accumulation rate is low (0.17 mm/a) for the basal part, suddenly increasing 2 to 6 times (to 0.33-0.95 mm/a) in the uppermost part of the unit (La Higuerita section, fig. 4.8a). Based on the shape of the accumulation rate curve, the abrupt increase in rate at ~9.7 Ma could be interpreted as likely indicating the time of thrust emplacement. The fact that two sites record a simultaneous accumulation increase also favours tectonic subsidence.

Based on the subsidence history, timing of changes of detritus provenance and facies shifts (summarized in fig. 4.12), the favoured interpretation is as follows. The La Pilona strata record between 11.7 and 10.0 Ma a coarsening upward trend that includes a shift to channelized conglomeratic deposition at approximately 10.9 Ma. By that time, there is also an important compositional change, marked by an abrupt increase in the content of sedimentary, low-grade metamorphic clasts and acidic volcanics of the Choiyoi Group whose sources are found in the Frontal Cordillera. The first thick conglomerate composed of clasts of these lithologies occurs at ~10.9 Ma whereas a sandstone collected ~20 m from below this horizon does not show these lithologies. The concurrent shifts in facies and provenance suggest that just prior to or at approximately 10.9 Ma there was a phase of thrust activity in the Frontal Cordillera (fig. 4.13).

At approximately 9.7 Ma, the La Pilona strata record a marked increase in subsidence rates, most likely related to a significant phase of deformation. This interpretation is strengthened by the observed vertical trend in facies, which agrees with models of foreland basin sedimentation that relate the onset of tectonic episodes with the initiation of a phase of fine-clastic deposition (e.g., Blair and Bilodeau, 1988; Heller et al., 1988; Flemings and Jordan, 1990). In particular, the onset of a period of increasing subsidence seemed to have occurred somewhat prior to ~9.7 Ma, as inferred from the occurrence of progressively thicker red and green (lacustrine) mudstone intervals and thinner intercalations of sandstone deposits. The continued increase in accumulation rates that accompany progradation of the

LA PILONA FORMATION

La Higuera section

Fig. 4.12. Summary of most relevant data of the La Pilon Formation, which includes sediment accumulation rates (complementary data are provided in fig., 4.8), timing of changes of detritus provenance and facies shifts.

Fig. 4.13. Schematic diagram showing a tentative interpretation of the thrust history recorded by the La Pilona Formation (~11.7 - 9.0 Ma). This unit is thought to record the initiation of a phase of deformation in the Frontal Cordillera at about 10.9 Ma and a phase of important tectonic significance at approximately 9.7 Ma. For explanation, see text.

conglomeratic fluvial deposits in the upper part of the La Pilona Formation could be related to the beginning of the time that proximal subsidence ceased to trap sediment, which instead was passed to distal sites.

Within the context of this interpretation, it is possible that immediately prior to ~10.9 Ma, deformation of the Frontal Cordillera was just beginning. At this time, it is probable that there were only local exposures of the deeper structural levels and/or that the site of deposition was far from the tectonic load that defines the flexural wavelength of the basin. The subrounded nature of the conglomerates and their comparatively finer grain size when compared with the uppermost conglomeratic sequences suggest at least a long transport from the thrust-bounded terrain. Thrust emplacement at ~9.7 Ma may have involved a more generalized thrust activity along the same fault or motion of a new thrust that stepped basinward.

There is no field evidence for the development of an erosional unconformity at around 9.7 Ma, which would be expected at the onset of a new thrust cycle (cf., Jordan and Flemings, 1990). However, its existence cannot be ruled out, and it may have developed within a fine-grained interval where an unconformity would not be readily recognized. The correlation of the local magnetic column (La Higuera section) to the magnetic polarity time scale (MPTS) does not highlight the existence of a hiatus in the local section. This correlation was guided by the relative thicknesses of normal and reversed zones and by ^{40}Ar - ^{39}Ar dates of interbedded ashes (cf. Chapters 2 and 3). However, in the discussing of the age assignments of the polarity zones for the La Higuera section (cf. Chapter 3, section 4d), it was mentioned that the isotopic age of sample Md-6 (10.42 ± 0.20 Ma) was in disagreement with the age assignment

provided by the MPTS for the younger age of the long normal chron C5n (9.7 Ma). If the isotopic date of this sample were the faithful record of the age of the ash horizon, this would imply the existence of an unconformity spanning a few thousand years, which was developed when fine-grained sedimentation began to be more significant. If this were the case, the conglomerates from the lower part of the La Pilona unit would have accumulated at a faster rate than predicted.

(c) Tobas Angostura and Río de Los Pozos formations

Deposition of these units at the study localities occurred between ~8.9 and ~8.3 Ma, where an unconformity separates the Río de los Pozos Formation from the overlying Mogotes Formation, although most probably the Río de los Pozos Formation is at least as young as ~7.2 (cf. Chapter 3). The Tobas Angostura unit exhibits facies assemblages consistent with deposition on a low-relief alluvial setting where sedimentation was the response of a semiarid fluvial system to the pyroclastic volcanism. Periodic and explosive eruptive episodes, recorded by air-fall tephra, produced large volumes of fragmented volcanic debris which subsequently overwhelmed gravelly braided streams. Streams could sometimes accommodate the initial influx of volcaniclastic material, but they rapidly became choked as deposition switched to unchannelized flows, leading to the emplacement of extensive sheets of pumiceous sandstones and hyperconcentrated flood-flow deposits. Sequentially through time, as the contribution of volcanic detritus waned, drainage of hill slope with different lithologies gradually regained precedence. The provenance study documents the presence of clasts derived from the Principal and Frontal cordilleras. The coarsening and deepening upward trend of the Río de

los Pozos Formation is consistent with progressive incision and enlargement of channel networks.

Deposition of the Tobas Angostura unit between ~ 8.9 and 8.7 Ma is correlative with the latest eruptive episodes that included voluminous pyroclastic eruptions occurring in the Aconcagua region (Ramos et al., 1996a). Lateral variations in thickness of this unit (i.e., east-southeastward tapering apron) in conjunction with radiometric dates from the upper section of the Complejo Volcánico Aconcagua (cf., Ramos et al., 1991, 1996a) support this correlation (fig., 4.14).

There are not enough data points to reconstruct the subsidence history of the basin during accumulation of the Tobas Angostura and Río de los Pozos units. However, it is worth mentioning that the magnitude of the derived accumulation rate for the Tobas Angostura unit (a minimum estimate of ~ 0.4 mm/a) may primarily reflect the influence of large volcanism-induced sediment loads, rather than tectonism. Tobas Angostura deposition resulted from volcanic events that provided large sediment loads probably in excess of geomorphic thresholds allowing rapid aggradation. The degree to which such short-term aggradation is recorded in these strata is difficult to quantify, but lack of evidence of widespread incision suggests that volcanism-induced depositional episodes may have occurred so frequently that erosion was never initiated. Widespread aggradation may have been terminated in coincidence with the cessation of volcanic activity, but in addition, factors such as climatic change, regional uplift in the High Andes and/or Frontal Cordillera or changes in base level may have contributed to the reestablishment of an environment dominated by normal fluvial processes. Existing data are insufficient to evaluate these alternatives.

Fig. 4.14. Schematic block diagram (modified from Jordan et al., 1983) showing the correlation established in this study between deposition of the Tobas Angostura unit and the largest and latest eruptive episodes in the Aconcagua region.

6. SUMMARY OF CONCLUSIONS

The chronostratigraphy and sedimentological data from the synorogenic strata outcropping in the southernmost extension of the Precordillera, in Mendoza Province, reveal details of Miocene and Pliocene tectonic activity, volcanism and deposition. Neogene strata accumulated in response to deformation and exhumation in the Principal and Frontal cordilleras of the Andes.

(1) The Mariño Formation, whose deposition spans ~15.7-12.2 Ma, records two phases of thrusting in the Principal Cordillera. The earlier phase occurred prior to ~15.7 Ma and is tied to deformation and exhumation of the andesitic Mesozoic volcanic sequences. The second phase of thrusting is interpreted to have occurred at about 15.5 Ma and involved deformation of the Mesozoic marine sequences of the Aconcagua fold-and-thrust belt.

(2) The first episode of thrusting is recorded by the 20 m thick fine conglomeratic sequences of the lower member (Purple conglomerates) of the Mariño Formation. It is hypothesized that these strata accumulated during a period of declining tectonic activity and that their deposition is linked with thrust activity in what is now the western part of the Aconcagua fold-and-thrust belt. These deposits could record the earlier phase of thrusting in the Aconcagua thrust belt, whose onset is known to have occurred at about 20 Ma (Cegarra and Ramos, 1996).

(3) The onset of the second phase of thrusting that involved deformation of the marine Mesozoic sequences is tied with the beginning of the accumulation of the middle member at about 15.4 Ma. These deposits are interpreted as a major retrograding facies developed as a result of a thrust-induced subsidence. The beginning of deposition of the upper member at 14.7 Ma marking the reestablishment of a stream-dominated setting is linked with active deformation in the thrust belt that led to progradation of increasingly coarse debris. This interpretation supports the models that emphasize the importance of thrust emplacement as a mechanism for generating retrograding facies (e.g., Heller et al., 1988; Jordan and Flemings 1990).

(4) The sequence of terrain unroofing recorded by the Mariño strata documents an eastward migration of the deformation front. The thrust sheets that comprise Jurassic limestone are situated in what is now the central part of the Aconcagua fold-and-thrust belt. Motion on one of the westernmost faults occurred at 14.5 ± 1.0 Ma (Cegarra and Ramos, 1996), which allows for inference that the Mariño strata accumulated in response to exhumation of one of these thrust sheets. It is possible that the coarsening upward trend of the Mariño strata paralleled in part the eastward progression of thrusting of the thrust sheets that comprise Jurassic limestone, although there are not sufficient data to prove this statement.

(5) Deformation that began at 15.5 Ma overlaps in time with episodes of volcanic activity in the Cordilleran region, as inferred from the occurrence of pyroclastic-rich horizons and the tuffaceous nature of the Mariño strata. The beginning of deposition of this unit is temporally

coincident with the eastward shifting of the subduction-related magmatic foci to the Argentine Principal Cordillera. The volcanic and pyroclastic rocks of the Complejo Volcánico Aconcagua (Ramos et al., 1996a) record ages spanning the interval ~ 15.8 - 8.5 Ma, whereas in the adjacent Chilean region, the Farellones Formation records an important period of magmatic activity between ~ 20.4 and ~ 16.6 Ma (Ramos et al., 1991, 1996a; Rivano et al., 1990).

(6) The erosional unconformity spanning less than 1 Ma that separates the Mariño strata from the La Pilona unit most probably is related to a phase of deformation and exhumation in the Cordilleran region, but insufficient data preclude more detailed interpretation.

(7) The La Pilona Formation, whose deposition spans ~ 11.7 - 9.0 Ma, is thought to record the initiation of a phase of deformation in the Frontal Cordillera just prior to 10.9 Ma and a phase of important tectonic significance at approximately 9.7 Ma. Provenance and paleocurrent data are consistent with clast derivation from northwestern highland sources, most probably from the Cordón and Cuchillas del Tigre, which provided volcanic detritus from the Choiyoi Group and low-grade metamorphic and sedimentary rocks of lower Paleozoic age. Available data preclude determination of whether these two tectonic events involved episodic movement of the same fault or a shift in the locus of active thrusting involving motion on other faults.

(8) The onset of a phase of uplift in the Frontal Cordillera is thought to have occurred just prior to 10.9 Ma, based on a shift from fine-textured deposits to conglomeratic fluvial

deposition and a significant compositional change. The interval represented by these conglomerates could be interpreted as the time when deformation of the Frontal Cordillera was just beginning, when there were probably only local exposures of the deeper structural levels and/or the source area was a considerable distant from the site of deposition. The abrupt increase in subsidence rates at about 9.7 Ma in conjunction with the shift to finer-grained facies recording clay-playa and lacustrine deposition suggest that these changes were due to thrust emplacement which may have involved a basinward shift in the locus of uplift. The shift from fine-grained to conglomeratic deposition in the upper part of the La Pilona unit stems from progradation as active thrusting progressed.

(9) Exhumation and displacement of the Frontal Cordillera overlapped in time with deformation in the Principal Cordillera in which the latest phase of deformation occurred at approximately 8.0 Ma (Ramos, 1996b; Cegarra and Ramos, 1996).

(10) Deposition of the La Pilona Formation was partially synchronous with extrusive episodes in the Cordilleran region, as sometimes recorded by air-fall tephra. Sedimentation of this unit overlapped in time with accumulation of most of the upper 1500 m thick section of the Complejo Volcánico Aconcagua, whose age has been estimated to span the time interval ~11.1-9.6 Ma (cf. Ramos et al., 1996a). However, the top of this section is considered to be of younger age based on the age of the Tobas Angostura unit as outlined in the next conclusion.

(11) Deposition of the Tobas Angostura tuffs between ~ 8.9 and 8.7 Ma is correlative with the largest and latest extrusive episodes occurring in the Principal Cordillera. Radiometric dates of volcanic rocks and sills from the top of the Aconcagua mountain are in good agreement with this correlation (cf. Ramos et al., 1996a).

(12) Deposition of the Tobas Angostura unit may primarily reflect the influence of large volcanism-induced sediment loads, rather than tectonism. Deposition of the overlying Río de los Pozos Formation, which probably is at least as young as ~ 7.2 Ma, records the reestablishment of normal fluvial processes. The provenance study documents the presence of clasts derived from the Frontal and Principal cordilleras.

(13) The coarse deposits of the Mogotes Formation accumulated at the study site in response to thrusting of the Frontal Cordillera, along the eastern flank of the Cordón del Plata. Deposition of the Mogotes Formation at the study site probably started at ~ 3.0 Ma (cf. Yrigoyen, 1994; Yrigoyen et al., 1995) and ceased prior to ~ 0.4 Ma, according to temporal constraints provided by air-fall horizons interbedded in Quaternary terraces (this study), and most probably before or at ~ 1 Ma. This implies that tectonic activity in the Frontal Cordillera shifted through time to the south, from the Cordón del Tigre to the Cordón del Plata.

(14) Motion of the La Pilona and Tupungato structures in the Precordillera started after the beginning of the accumulation of the Mogotes Formation at ~ 3.0 Ma and ceased prior to ~ 0.4 Ma. The occurrence of growth geometries in strata of the Mogotes Formation (Chiaramonte,

1996) suggests that deformation was partially synchronous with deposition of this unit.

(15) Most of the hydrocarbons in the Cuyo Basin are hosted in oil fields largely controlled by structural traps, such as La Pilona and Tupungato anticlines (Moratello, 1993). The study of ages of the Neogene sequences suggests that appropriate conditions for the beginning of oil generation (maturation) in the Triassic Cacheuta Formation, the main source rock, occurred certainly after deposition of the Mariño Formation (~12 Ma) and most likely after deposition of a number of younger units. Prospective closures developed during east-west contraction in the Late Pliocene - Early Pleistocene? prior to ~0.4 Ma, the age of the air-fall horizon interbedded with Quaternary gravels in the La Pilona area.

(16) The eastern border of the Precordillera is still active, as inferred from earthquake focal mechanisms, intense compressive deformation and neotectonic activity (Triep, 1987; Ramos, 1990; Bastías et al., 1993, Kozłowski et al., 1993). In some localities of the study site, Quaternary deposits unconformably overlying the La Pilona anticline are disrupted by normal faults with centimetre-scale throws.

(17) The tephra horizons preserved in Quaternary terraces of the Cacheuta-Tupungato region attest to the existence of explosive eruptions that occurred at approximately 0.4-0.3 Ma (this study). Although lack of sufficient data precludes determination of the volcanic source, it is well known that volcanic activity occurred south of 33°15' in the Quaternary (Hildreth and Moorbath, 1988; Guerstein, 1993; Sruoga et al., 1993). There are several possible sources,

such as the Maipo volcano (0.5 Ma; Stern et al., 1984), the Marmolejo - San José volcanic complex and the Tupungatito volcano, among others (cf. Sruoga et al., 1993).

CHAPTER 5

TECTONIC SYNTHESIS

The nonmarine Cenozoic deposits of the Mendoza Precordillera which are exposed adjacent to the Cordón del Plata in the Frontal Cordillera, Argentina, record the eastward migration of the Andean deformation front since the middle Miocene. Magnetostratigraphy and sedimentological data from these strata constrain the sequence and timing of deformation in the Principal and Frontal cordilleras. The preceding interpretations (cf. Chapter 4) will be summarized with reference to the Neogene tectonic evolution of this segment of the Andes (33°S), with additional information provided from other studies (e.g., Jordan et al., 1983; Ramos, 1988, 1993, 1996b; Mpodozis and Ramos, 1990; Cegarra and Ramos, 1996).

Cenozoic strata accumulated in the Huarpes depression situated adjacent to the eastern slope of the Cordón del Plata, at the southern end of the Precordillera. The Precordillera fold-and-thrust belt, which defines the eastern flank of the Andes, dies out as a physiographic unit at 33°S. At this latitude, the feature that characterizes the subduction geometry beneath the Andean Cordillera is a southward transition in the dip of the slab from nearly horizontal to ~30°; this appears to occur as a smooth flexure (Cahill and Isacks, 1992). Associated with the change in the geometry of the subducted plate, there are changes in the large-scale tectonic features of the orogen north and south of 33°S (Jordan et al., 1983; fig. 5.1). The lack of an active volcanic arc, the development of the Precordillera as a thin-skinned thrust belt, and the basement block uplift of the Pampeanas Range in the flat-subduction

Fig. 5.1. Schematic block diagrams showing subducted plate segments and Neogene morphotectonic units (after Jordan et al., 1983).

segment (28°S to 33°S) are linked with the shallowing of the subducted plate during the Late Miocene and later (Jordan et al., 1983; Jordan and Allmendinger, 1986; Ramos, 1988; Kay et al., 1988; Kay and Abbruzzi, 1992). South of 33°15'S, there is Quaternary arc volcanism, extensive back-arc volcanism (between 35°S and 37°S, Bermúdez et al., 1993) and a narrower region of foreland deformation (Jordan et al., 1983; Ramos and Nullo, 1993).

Neogene strata began to accumulate adjacent to the Cordón del Plata at about 15.7 Ma, in response to thrusting in the Principal Cordillera and coincident with the eastward shifting of the subduction-related magmatic foci to the Argentine side. Geochemical and geochronological data from the volcanic rocks in the region suggest that the subducted-plate angle began to lessen by ~16 Ma, long before the magmatism ceased, and that thickening of the continental crust related to the tectonic stacking of the Andean Cordillera evolved since approximately that time (Kay et al., 1988, 1991; Ramos et al., 1991). It has been suggested that there is a causal relationship between the subduction of the aseismic Juan Fernandez Ridge and the flattening of the subduction zone (Barazangi and Isacks, 1976), which according to Pilger (1984) began at about 15 Ma.

There is evidence of an earlier Miocene phase of deformation in the Principal Cordillera which is linked with the beginning of a period of higher convergence rates at about 25 Ma related to a major plate reorganization in the east-central Pacific (Ramos, 1988; Ramos et al., 1991). Geological and chronological data suggest that the onset of deformation in what is now the western part of the Aconcagua fold-and-thrust belt occurred at approximately 21-20 Ma (Ramos et al., 1991; Cegarra et al. 1993; Ramos, 1996b). In this study, it is hypothesized that the phase of thrusting recorded by the basal conglomerates of the Mariño

Formation (Purple conglomerates) prior to ~15.7 Ma might represent the last stages of this phase of deformation (cf. Chapter 4). It involved exhumation and displacement of the andesitic Mesozoic volcanic sequences, presently widely exposed in the adjacent Chilean region and along the Argentine boundary.

The important phase of deformation that started at about 16-15 Ma controlled the development of thin-skinned shortening of the Mesozoic marine sequences in the Principal Cordillera (Ramos et al., 1991; Cegarra et al., 1993; Ramos, 1996b; this study). This deformation involved motion on faults in what is presently known as the central part of the Aconcagua fold-and-thrust belt. The Mariño strata record thrusting which started at ~15.5 Ma and which was still continuing at ~12.0, the age of the youngest dated Mariño deposits. Thrusting may have involved the motion of several thrust sheets comprising Jurassic limestone which deformed in a forward-breaking sequence (cf., Cegarra and Ramos, 1996). An event of tectonic significance occurred at ~12.0 Ma, as recorded by the erosional unconformity that separates these strata from the overlying La Pilona Formation. However, lack of sufficient temporal constraints in conjunction with the complexity of the thrust history (cf. Cegarra and Ramos, 1996) preclude more detailed interpretation. According to Ramos (pers. comm., 1997), this unconformity is also recorded by the synorogenic deposits that crop out in the western slope of the Frontal Cordillera, to the west of the Cordón del Portillo, where a marked angular unconformity separates the Tunuyán conglomerates from the Papal Formation (cf. Yrigoyen, 1993a).

The last phase of deformation in the Aconcagua fold-and-thrust belt at about 9-8 Ma coincides with the latest episodes of volcanic activity in the Cordilleran region (cf. Ramos and

Yrigoyen, 1987; Ramos et al., 1990, 1991, 1996a; Cegarra and Ramos, 1996; Ramos, 1996b). These extrusive episodes included voluminous pyroclastic eruptions, as recorded in distal parts of the basin by the pyroclastic-rich Tobas Angostura deposits, dated at 8.9-8.7 Ma (this study). During that time, the deformation front of the Principal Cordillera migrated to its present position. The timing of deformation is constrained by the proximal synorogenic deposits of the Vacas basin (Pérez and Ramos, 1996; Santa María Conglomerates of Schiller, 1912) that record the eastward progression of deformation in the Principal Cordillera. These deposits, which contain in their uppermost part a pyroclastic horizon dated at 8.1 ± 0.6 Ma (Ramos et al., 1990; Pérez and Ramos, 1996), are overridden by Jurassic limestones in the Cerro Penitentes situated in the present thrust front.

Exhumation and displacement of the Frontal Cordillera occurred prior to the latest episode of uplift in the Principal Cordillera and is recorded by the La Pilona Formation in the Cacheuta region. This unit, deposited from ~11.7 to ~9.0 Ma, records the onset of a phase of deformation in the Frontal Cordillera just prior to ~10.9 Ma and an important phase of uplift and deformation at approximately 9.7 Ma. The onset of thrusting is marked by the appearance of volcanic detritus from the Choiyoi Group and of Lower Paleozoic low-grade metamorphic and sedimentary rocks derived from highland sources to the northwest, most probably from the Cordón and Cuchillas del Tigre, where part of the shortening of the Frontal Cordillera was taken-up by back thrusts (Cortés, 1993). A probable scenario just prior to ~10.9 Ma is that there were only local exposures of deeper structural levels and/or that the thrust-bounded terrain was considerably distant from the site of deposition. The 9.7 Ma phase of uplift may represent a more generalized thrust activity along the same fault or a shift in the

locus of tectonic activity involving a basinward migration of deformation. The unconformity at ~9 Ma that separates La Piona strata from the Tobas Angostura unit involved a short hiatus that developed as a result of tectonic uplift in the Frontal Cordillera and/or in the Principal Cordillera.

Tectonic activity in the Frontal Cordillera shifted through time to the south, along the eastern slope of the Cordón del Plata. The coarse deposits of the Mogotes Formation accumulated at the flank of the thrust belt in response to motion on the Las Carreras fault system (Espolón de la Carrera; Polanski, 1958), an imbricate structure with east vergent thrusts that in the northern end is segmented and forms an en echelon array (Cortés, 1993). Deposition of this unit probably started at ~3 Ma at the study site (cf., Yrigoyen, 1994; Irigoyen et al., 1995) and ceased prior to ~0.4 Ma, as indicated by radiometric data from ash fall beds interbedded with Quaternary deposits (cf. Chapter 2). The basal unconformity ascribed by some authors to the Diaguitic Phase (e.g., Yrigoyen, 1993a) could have developed as a result of diachronous pulses of deformation along the Cordón del Plata starting after ~6 Ma as inferred by the estimated age of the youngest Río de los Pozos strata which unconformably underlie the Mogotes conglomerates (cf. Irigoyen et al., 1995).

It is possible that deformation along the Frontal Cordillera first started in the north where an ancient suture zone (the Paleozoic suture zone of Chilenia and Precordillera terranes; Ramos et al., 1984) may have acted as a zone of weakness favouring displacement during the Andean compression. Evidence of a Pleistocene tectonic reactivation along the irregular border of the Chilenia terrane is found in the Uspallata Valley (Cortés, 1991).

The Neogene strata of the Mendoza Precordillera has been deformed in the Late

Pliocene (and Early Pleistocene?) into fault-propagation folds and emergent east-verging thrusts. The occurrence of growth geometries on strata of the Mogotes Formation in the study region (Chiaramonte, 1996) suggests that deformation was partially synchronous with deposition of the unit. Motion of the La Pilona and Tupungato structures ceased prior to ~ 0.4 Ma according to geochronologic data, most probably at about 1 Ma (cf. Chapter 2). Based on field mapping, well data and seismic reflection profiles, Sarewitz (1988) interpreted La Pilona structure as a duplex anticline (fig. 5.2; plate 5.1). However, the interpretation of an imbricate wedge overlying a ramp and bounded above by west-verging faults is in disagreement with field observations which indicated a preponderance of east- rather than west-verging structures (see plates 5.2-5.4). According to Sarewitz (1988), seismic reflection profiles show a kinklike geometry at the base of the western flank of the anticline and an obvious zone of horizontal detachment beneath the core and eastern flank of the anticline, which suggests the existence of a fault-bend fold.

The present location of the thrust front near the city of Mendoza indicates that deformation has progressed eastward in the last ~ 3 Ma. Seismic sections show that the orogenic front is composed of a set of imbricated overthrusts (Bettini, 1980; Figueroa and Ferraris, 1989). Evidence of neotectonic activity is found in Quaternary terrace gravels in the Arroyo del Agua Blanca valley in the Cacheuta region (plate 5.5) and farther east in the outer foothills of the Mendoza Precordillera, where Quaternary faulting activity is expressed as fault-line scarps that disrupt alluvial fans and terraces (Bastías et al., 1993).

In summary, Tertiary synorogenic deposits record the eastward migration of the orogenic thrust front since the Early Miocene. In the Principal Cordillera, thrust activity is

Fig. 5.2. Balanced cross-section through La Pilona anticline (from Sarewitz, 1988).

Plate 5.1. The gypsum- and anhydrite-bearing Divisadero Largo Formation (withish exposure, see arrow) is considered by Sarewitz (1988) to be the upper west-verging detachment surface of the La Pilona anticline (see fig. 5.2). Evidence of bedding-shear displacements is found in gypsum horizons of this unit, above which lie unconformably the Purple Conglomerates of the Mariño Formation. This exposure is located north of the Mendoza river, in the Camping Santo Domingo area.

Plate 5.2. (a) Bedding-shear displacement, top to the east, is inferred from the occurrence of fractures (see arrows) in Triassic sedimentary rocks. These fractures are filled with gypsum probably derived from the overlying gypsum- and anhydrite-rich Divisadero Largo Formation. (b) East-verging fault-propagation fold observed in the Triassic rocks that crop out south of the Cacheuta peak, near the Arroyo de las Minas riverbed.

a

b

Plate 5.3. In the study area, the Upper Eocene? Divisadero Largo Formation is deformed into east-verging fault-propagation folds (a syncline is shown in photograph a) and east-verging thrusts (photograph b).

a

b

Plate 5.4. (a) In the study area, the Mariño Formation is deformed into broad synclines (as shown) and anticlines. (b) An east-verging duplex anticline (see arrow) observed in rocks of the Mariño Formation.

a

b

Plate 5.5. Evidence of neotectonic activity is observed in Quaternary terrace gravels in the Arroyo del Agua Blanca valley. These deposits are disrupted by faults with centimetre-scale throws.

known to have occurred in the western part of the Aconcagua fold-and-thrust belt at ~20 Ma (Ramos et al., 1990; Cegarra and Ramos, 1996). Motion on the central thrusts began at about 15.5 Ma; deformation progressed eastward and ceased by ~8 Ma (Ramos et al., 1990; Cegarra and Ramos, 1996). Thrust activity in the Principal Cordillera is recorded by the proximal synorogenic deposits of the Vacas basin (Pérez and Ramos, 1996) and, east of the High Cordillera in the Uspallata and Cacheuta regions, by distal synorogenic deposits that contain several tuff horizons attesting to Cordilleran volcanic activity at that time. Episodes of deformation in the Principal Cordillera were partially synchronous with thrust activity to the north, at the latitude of La Ramada thrust belt in southern San Juan province. Based on the study of the Cenozoic nonmarine strata of the Manantiales basin, Jordan et al. (1996) conclude that the La Ramada thrust belt was active during the interval spanning ~19-10 Ma, and possibly as recently as 8-5 Ma. The earlier phase of thrusting at about or before 19 Ma involved deformation in the western part of the belt as in the Aconcagua region. At about 16 Ma, there was an important phase of deformation which approximately coincides in time with the beginning of active thrusting in the central part of the Aconcagua thrust belt, after which deformation shifted to the east and thrusting continued after 10.5 Ma, the age of the youngest Manantiales dated horizon.

The earlier phase of thrusting in the Principal Cordillera at the latitude of the La Ramada belt coincides in time with the onset of deformation of the San Juan Precordillera where motion of east-verging thrust faults started at about 20 Ma and continued up to the present (Jordan et al., 1993b). To the south, in Mendoza province, thrust activity in the Aconcagua thrust belt partially overlapped in time with exhumation and displacement of the

Frontal Cordillera, which started at about ~11 Ma. Thus, deformation in the Principal and Frontal Cordilleras probably progressed simultaneously between about 11 and 8 Ma. At the latitude of the study site, the last phase of uplift of the Frontal Cordillera occurred at about 3.5-3.0 Ma, after which deformation shifted eastward to the Precordillera and continued up to the present. Figure 5.3 shows schematically the migration of the orogenic front since Miocene times and distribution of the synorogenic deposits that record deformation of the Principal and Frontal cordilleras.

Based on present topography and gravity data, it has been estimated that the Andean shortening between the trench and the foreland at the latitude of the La Ramada thrust belt (30° - 32° S) is about 150 km, whereas at the latitude of the Aconcagua fold-and-thrust belt (33° S) the shortening is approximately 130 km (Introcaso et al., 1992). Allmendinger et al. (1990) found that most of the Cenozoic shortening at 30° S is concentrated along the Precordillera and accounts for 60-75% (90-127 km) of the total 150-170 km of shortening. Cristallini et al. (1995) estimated that at this latitude, shortening in the Principal Cordillera is no more than 18 km. At 33° S, the greatest shortening is mainly restricted to the Principal Cordillera, since the Frontal Cordillera was uplifted as a single block without significant shortening, and inversion of the Cuyo basin involved even less contraction (Ramos et al., 1996b; Cegarra and Ramos, 1996). The Principal Cordillera records a 55-60 km shortening including the Chilean side of the Aconcagua fold-and-thrust belt (Kozlowski et al., 1993). Figure 5.4 depicts an structural cross-section of the Principal and Frontal cordilleras and the Cuyo basin at the latitude of the Tupungato anticline (33° 20'S).

Fig. 5.3. Migration of the orogenic front since Early Miocene to the present in northern Mendoza province. Map modified from Kozłowski et al., 1993.

Fig. 5.4. Structural cross-section of the Principal and Frontal cordilleras and the Cuyo basin at 33°20'S latitude (from Kozlowski, 1993)

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APPENDIX

Table A1. Paleomagnetic results

Table A1.1. I alcomagetic results

Specimen	J _{NRM} (10 ⁻⁴ A/M)		Susc. (10 ⁻³ SI)	Characteristic component		Virtual geomagnetic paleopole			Characteristic component		Min- eral	Po- lar- ity	Viscous component		
	D (°)			I (°)	E (°)	Lat (°N)	Lon (°E)	dm (°)	dp (°)	T _b range (°C)			Min	Max	
										Min					Max
TOSCA DE MARINÓ SECTION															
Divisadero Largo Formation															
Site 00. S/D bed: 150/10															
01A	1.42		9.63	4.3	-24.4	2.7					100	560-580	N	20	100
03A	0.98		9.87	45.0	-38.6	2.1					100	560?-580	N	20	100
Mean				23.0	-33.1	37.1		64.5	350.5	-	-				
Site 02. S/D bed: 150/10															
02A	0.77		3.38	176.9	54.10	2.6					250	600-620	H	20	250
03A	0.48		3.35	86.0	83.9	2.6					200	600-620	H	20	200
Mean				166.6	71.7	36.4		-64.7	308.2	-	-				
Site 04. S/D bed: 155/8															
01B	0.47		7.91	192.3	31.8	6.3					300	660-680	M-H	100	250
03A	0.25		8.08	212.9	65.5	8.3					300	660-680	M-H	20	250
Mean				199.0	49.0	35.9		-73.5	195.0	-	-				
Marinó Formation															
Site 05. S/D bed: 150/8															
01A	0.01		0.10	-	-	-									
02B	0.01		0.12	-	-	-									
Site 06. S/D bed: 155/20															
01A	2.65		16.31	144.4	-9.6	2.4					250	580-600?	?		
02A	2.49		14.64	144.9	-27.6	1.8					200	580-600	?	A	200
03A	1.33		9.02	131.4	-15.0	3.2					200	580-600?	?	A	
Mean				140.1	-17.4	18.1		-33.4	61.5	18.7	9.7				
Site 07. S/D bed: 155/20															
01A*	1.27		8.97	231.9	56.3	29.4					620	660-680	H	R	20
02A*	0.77		7.90	210.2	44.9	25.4					≥530	600-640	H	R	100
Mean				219.7	51.1	17.7		-56.7	210.0	-	-				
Site 08. S/D bed: 150/15															
01A	0.91		7.97	36.8	-71.3	20.6					350	580-600	?	N	20
02A	0.75		6.00	43.6	-53.1	21.8					300	560-580	?	N	20
03A*	0.67		29.90	86.7	-52.2	21.7					400	560-580	?	N	100
Mean				59.2	-60.8	27.3		43.2	49.3	41.5	31.7				

Table A1. Continued

Table A1. Continued														
Specimen	J _{NRM} (10 ⁻⁴ A/M)	Susc. (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic pole			Characteristic component		Min- eral	Po- lar- ity	Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dp (°)	T _b range (°C)				Min	Max
									Min	Max				
TOSCA DE MARIÑO SECTION, continued														
Mariño Formation, continued														
Site 09, S/D bed: 150/15														
02A	0.91	12.83	22.9	-49.5	4.6				100	560-580	?	N		
03A	1.02	11.17	52.9	-71.9	11.8				350	560-580	?	N	20	300
Mean			32.5	-61.4	26.2	62.9	50.3	-						
Site 10, S/D bed: 145/18														
01A	0.20	0.85	50.8	-27.5	3.9				20	640?-660	H	N	20	200
02A ^b	0.75	4.24	59.2	-21.2	8.5				20	>550	H	N		
03B	0.29	2.21	10.5	-41.4	8.3				20	>500-580	H	N		
Mean			42.5	-31.7	37.9	48.7	8.9	42.6	23.9					
Site 11, S/D bed: 150/15														
01A	1.56	16.19	31.9	-45.1	2.1				100	500(580)	?	N		
02A	2.04	16.44	18.5	-45.3	1.1				100	450-500	?	N		
03A	1.26	13.43	-	-	-				-	-	-	-		
Mean			25.2	-45.4	9.4	67.4	11.6	-						
Site 12, S/D bed: 150/14														
01B	1.46	7.79	12.4	-31.8	2.9				20	660-680	H	N		
02B	2.12	5.76	19.5	-61.8	3.4				100	~620	H	N		
03A ^b	1.25	10.03	350.7	-57.8	2.9				20-100	>550	H	N		
Mean			8.0	-51.1	28.7	83.1	12.0	39.0	26.4					
Site 13, S/D bed: 145/20														
01B	1.97	29.85	0.4	-38.5-	3.2				100	350?	?	N		
02A	1.89	21.89	15.7	-55.1	3.1				100	~450	?	N		
03A	2.20	26.77	23.5	-18.6	5.1				300-350	560	?	N		
Mean			13.7	-37.8	32.4	73.0	340.0	38.2	22.6					
Site 14, S/D bed: 180/40														
01A	1.98	27.62	53.7	-33.3	2.7				20	~530	?	N		
02A	2.40	25.20	20.3	-30.9	1.2				100	450-500	?	N		
03B	2.29	28.25	48.6	-62.4	1.1				20-100	>530-550	?	N		
Mean			39.4	-43.2	35.3	54.9	18.2	43.9	27.2					

Table A1. Continued

J _{NRM} (10 ⁻⁴)		Susc. (10 ⁻³ SI)	Characteristic component		Virtual geomagnetic paleopole			Characteristic component		Min- eral	Po- lar- ity	Viscous component	
Specimen	A/M		D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dm (°)	dp (°)			Min T _b range (°C)	Max
TOSCA DE MARIÑO SECTION, continued													
Mariño Formation, continued													
Site 15, S/D bed: 215/30													
01A	2.79	27.67	60.6	-66.0	1.4				20	400-450	?	N	
02A	1.88	21.13	35.4	-60.9	0.9				20	450(560?)	?	N	
03A	2.12	21.90	67.7	-74.9	1.4				20	400-450	?	N	
Mean			51.6	-67.8	14.8	48.8	62.3	24.8	20.8				
Site 16, S/D bed: 210/25													
01B	0.99	23.33	283.6	-35.5	2.8				100?	>550	H	N	
02A	2.11	21.05	-	-	-								
03A*	1.54	15.56	337.9	-40.8	13.0				100	600-620	H	N	
Mean			309.7	-41.4	42.3	45.2	199.3	-					
Site 17, S/D bed: 225/45													
01A*	3.77	25.53	37.5	83.9	2.1				400-450	500-530	?	R	100
02A*	1.51	20.14	212.4	34.4	37.3				400-450	530-550	?	R	20
03B	1.35	22.29	-	-	-				-	-	-	-	20
Mean			211.6	65.2	61.7	-62.2	241.0	-					20
Site 18, S/D bed: 170/41													
01A	2.03	14.89	47.0	-13.5	2.0				20	500-530	?	N	
02A	2.32	21.57	38.9	-20.6	1.2				20	400-450	?	N	
02B	2.01	16.60	36.8	-17.1	1.3				20	530-560	?	N	
03A	2.28	19.46	43.0	-25.8	1.6				100	400-450	?	N	
Mean			41.4	-19.3	7.7	45.5	359.2	8.0	4.2				
Site 19, S/D bed: 175/20													
01A	4.55	26.77	44.8	-43.4	1.0				100	450-500	?	N	
02A	3.95	30.54	19.8	-29.4	2.4				20	450-500	?	N	
03B	4.20	32.00	56.4	-47.3	1.3				20	450-500	?	N	
Mean			38.7	-41.1	26.8	54.8	15.3	32.6	19.8				
Site 20, S/D bed: 145/22													
01B	1.81	12.82	356.4	-64.3	3.1				100	500-530	?	N	
02A	1.91	14.64	24.3	-36.0	2.8				100	580-600	?	N	
03B	1.06	13.69	21.5	-29.1	5.1				100	500-530	?	N	
Mean			17.5	-43.5	32.8	72.9	358.6	40.9	25.5				

Table A1. Continued

Table A1. Continued															
Specimen	J _{NRM} A/M	Susc. (10 ⁻³ SI)	Characteristic component		Virtual geomagnetic paleopole				Characteristic component		Polar- ity	Viscous component			
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dm (°)	dp (°)	Min		Max	T _b range (°C)	Min	Max
TOSCA DE MARINÓ SECTION, continued															
Marino Formation, continued															
Site 21. S/D bed: 150/10															
03A	0.45	6.08	339.2	4.6	21.6					200	530-560	M	A	20	200
04A ^c	0.47	6.43	-	-	-					-	-	-	A		
04B ^c	0.46	5.97	-	-	-					-	-	-	A		
Mean			339.2	4.6	21.6	49.5	257.7	-	-						
Site 22. S/D bed: 147/25															
01A	0.21	4.59	203.1	12.1	6.0					200-250	≥630	H	R	20	200
02A ^b	0.18	-	251.9	17.5	5.2					200-250	>550	H	R	20	200
03A	0.17	4.99	215.3	30.4	6.4					200	≥600	H	R	20	200
04A	0.34	2.95	256.1	56.9	9.3					250	660-680	H	R	20	200
Mean			218.3	34.6	29.2	53.1	8.17	33.5	19.3						
Site 23. S/D bed: 145/25															
01A ^a	2.09	15.63	167.4	47.0	18.5					≥500	600-620	H?	R	100	500-530
02A ^a	1.61	16.80	220.8	65.0	40.3					≥500	≥580	?	R	100	500-560
03A ^a	2.43	14.41	184.4	79.6	23.3					>500	~580	?	R	100	400
			186.9	65.8	32.2	-74.1	273.8	52.7	43.0						
Site 24. S/D bed: 145/25															
01A	1.66	12.71	40.6	20.6	1.6					100	450-500	?	A		
02A	1.69	14.58	188.6	-80.1	1.1					20	560?-580	?	A		
03A ^c	1.45	15.03								400?	580?	?	R?	20	300
Site 25. S/D bed: 180/25															
01A	0.84	8.97	8.3	-34.1	5.9					300	530-560	?	N	20	200
03A	0.73	9.98	53.8	-33.1	4.8					250	>530-580	?	N	20	200
Mean			31.1	-35.7	37.5	59.2	3.1	-	-						
Site 26. S/D bed: 170/25															
02A	0.51	6.77	197.0	13.3	4.6					200	>560-600	M	R	20	200
03A	0.06	6.44	203.9	33.4	12.5					350	560?-580	M	R	20	300
04A	0.18	8.52	227.3	15.5	11.8					200	580-600	M	R	20	200
Mean			209.5	21.2	29.1	-55.0	168.5	30.6	16.1						
Site 27. S/D bed: 160/28															
01A	0.28	6.17	192.8	59.9	11.5					200	560-580	?	R	20	200
02A	0.78	8.81	28.6	-36.5	2.2					100	530-560	?	N		368
03B	0.39	6.18	244.3	46.4	14.1					200-250	560-580	?	A		368

Table A1. Continued

Specimen	J _{NRM} (10 ⁻⁴ A/M)	Susc. (10 ⁻³ SI)	Characteristic component		Virtual geomagnetic paleopole			Characteristic component		Min-eral	Po-lar-ity	Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dm (°)	dp (°)			T _b range (°C)	Min
TOSCA DE MARINO SECTION, continued													
Mariño Formation, continued													
Site 28, S/D bed: 170/30													
01A	1.82	16.52	338.5	-57.0	4.7				300				
02B ^b	1.42	16.92	3.0	-56.1	19.1				250?				
03A	1.38	16.61	48.2	-30.3	7.1				250				
Mean			350.9	-57.1	13.5								
Site 29, S/D bed: 170/32													
01B	2.11	14.05	33.0	-13.2	2.2				100				
02B ^b	1.53	13.54	21.9	-20.5	-				100				
03A	1.25	11.27	1.3	-36.8	1.5				100				
Mean			19.9	-23.9	29.3								
Site 30, S/D bed: 180/30													
01A	0.48	7.73	218.7	60.8	5.3				250				
02A	0.42	7.20	182.1	55.8	13.2				200				
Mean			199.1	59.6	19.5								
Site 31, S/D bed: 165/40													
01A	1.43	10.87	60.5	-41.7	4.5				250				
02A	1.34	11.08	29.0	-53.8	10.5				200				
03A	1.23	12.31	-	-	-				-				
Mean			46.6	-48.8	24.1								
Site 32, S/D bed: 168/22													
01B ^b	0.22	4.69	253.7	18.0	14.3				200?				
02A	0.27	5.04	233.3	7.2	14.2				200?				
03A	0.20	3.86	188.8	37.6	3.8				200				
Mean			213.7	24.0	50.4								
Site A33, S/D bed: 180/24													
01A	1.93	14.35	25.4	-46.7	1.1				100				
02B	2.33	-	41.1	-48.1	0.6				100				
03B	2.34	15.64	46.4	-43.3	1.0				100				
Mean			37.8	-46.4	12.2								
Site B33, S/D bed: 160/20													
01B	0.51	9.03	208.7	57.5	8.5				200				
03A	0.67	9.62	-	58.9	-								
04A	0.50	10.33	178.2	59.1	8.4				200-250				
Mean			193.7	59.1	15.9								

Table A1. Continued

Table A1. Continued																
Specimen	J _{NRM} (10 ⁻⁴ A/M)	Susc. (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole				Min- eral	Po- lar- ity	Characteristic component			Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dm (°)	dp (°)			T _b range (°C)		Min	Max	
												Min	Max			
TOSCA DE MARINO SECTION, continued																
Marino Formation, continued																
Site 34, S/D bed: 163/20																
01A	0.11	10.19	205.0	67.3	3.2							?	350?	R	20	200
02A	0.13	11.96	169.1	56.3	6.1							H?	>580	R		
03A	0.24	8.04	215.9	41.1	8.3							H?	>550?	R		
Mean			198.4	56.7	30.3	-74.4	221.5	43.9	31.8							
Site 35, S/D bed: 160/20																
01B	0.42	5.42	34.2	-40.7	3.0							M?-H	≥560	N		
02A ^b	0.30	5.58	30.4	-40.7	6.0							M?-H	>550	N		
03A	0.37	5.70	23.5	-34.1	6.4							M?-H	≥560	N		
Mean			29.1	-38.5	8.7	61.8	4.6	10.4	6.2							
Site 36, S/D bed: 160/20																
01A	0.92	12.61	16.1	-36.0	4.4							?	≥530	N	20	200
02B	0.73	13.32	26.3	-40.9	3.3							?	530-560	N	20	200-250
03A ^b	0.72	12.63	97.0	-51.4	20.8							?	>550	N?		
Mean			21.0	-38.6	9.3	68.2	354.7	-	-							
Site 37, S/D bed: 135/15																
01A	0.19	10.82	194.6	52.7	5.5							M	≥560	R	20	200
02A ^b	0.27	6.22	210.0	69.4	8.3							M	>550	R		
03A	0.24	6.02	243.4	71.2	8.0							M	≥560	R	20	200-250
Mean			210.8	65.8	22.3	-62.5	242.9	36.3	29.6							
Site 38, S/D bed: 145/12																
01A	0.46	4.44	68.4	-41.9	4.1							M?-H	>560	N		
01B	0.44	3.82	52.6	-71.5	3.4							M?-H	≥580	N	20	200
Mean			63.7	-56.9	30.6	38.3	44.9	-	-							
02A	0.11	6.85	223.4	44.5	4.0							M?-H	>550	R		
03A	0.15	6.90	159.1	70.6	8.8							M?-H	>560	R		
Mean			204.1	61.9	40.2	-68.9	233.3	-	-							
Site 39, S/D bed: 127/5																
01A	0.65	9.20	201.0	54.8	3.0							H	600-620	R	20	200
02A ^b	0.41	7.95	214.4	61.7	14.0							H	>550	R		
04A	0.37	8.48	233.8	78.7	1.9							H	600-620?	R	20	200-250
Mean			211.1	65.4	21.0	-62.4	241.9	34.1	27.7							

Table A1. Continued

Specimen	J _{NRM} (10 ⁻⁴ A/M)	Susc (10 ⁻³ Sf)	Characteristic component		Virtual geomagnetic paleopole				Characteristic component		Min- eral	Po- lar- ity	Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dm (°)	dp (°)	Min T _b range (°C)			Max T _b range (°C)	Min
ARROYO DEL AGUA BLANCA SECTION														
Marino Formation														
Site 40, S/D bed: 135/13														
01B	1.09	10.37	10.3	-41.8	3.7					250	M?-H	N		
02A ^b	0.81	12.98	31.6	-51.1	8.8					20-250	H	N	>550	
03A	0.89	10.09	44.1	-45.7	3.9					20-250	H	N	660-680	
Mean			28.2	-47.1	19.9	65.4	17.1	25.7	16.6					
Site 41, S/D bed: 140/15														
00A	0.58	8.21	184.5	54.4	3.2					200-250	H	R	>550	100-200
03A	3.21	6.01	190.1	17.9	1.0					100-200	H	R	660-680	
Mean			187.9	36.1	36.7	-75.1	141.2	-	-					
Site 42, S/D bed: 120/14														
01A ^b	1.08	6.52	307.9	-23.0	3.0					20?	M-H	N	>550	
02A	0.52	10.14	300.4	-16.0	8.6					200-250	M-H	N	580-600	200
03A	0.67	6.27	339.3	-29.5	2.6					250	M-H	N	580-600?	200
Mean			315.0	-23.7	31.1	44.1	216.8	33.2	17.7					
Site 43, S/D bed: 124/15														
01A ^b	0.69	11.77	207.4	44.0	26.9					200?	M	R	>550	
03A	0.63	12.51	225.3	57.6	6.2					200	M	R	560-580?	200
04B	0.81	12.34	211.6	44.3	5.7					200	M	R	560-580	200
Mean			213.8	48.8	15.0	-61.1	203.5	19.6	13.0					
Site 44, S/D bed: 140/12														
02A	0.59	10.09	16.7	-31.1	3.5					100	?	N	580-600	
03A ^b	0.36	11.76	319.1	-21.3	8.3					250?	?	N	>550	
04A	0.37	10.92	351.0	-46.8	7.5					250	?	N	530-560	200
Mean			5.3	-39.6	25.2	78.4	316.0	-	-					
Site 45, S/D bed: 130/15														
01B ^b	0.44	6.03	202.1	14.8	2.8					200-250	H	R	>550	200
02A	0.07	7.93	195.7	15.3	4.3					200	H	R	620-640	200
03A	0.08	7.55	166.0	24.8	7.6					200-250	H	R	>560	
Mean			188.5	18.9	29.5	-65.3	131.3	30.7	16.0					

Table A1. Continued

Specimen	J _{NRM} (10 ⁻⁴ A/M)	Susc (10 ⁻³ SI)	Characteristic component		Virtual geomagnetic paleopole			Characteristic component		Min- eral	Po- lar- ity	Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dp (°)	Min T _N range (°C)			Max	Min T _N range (°C)
ARROYO DEL AGUA BLANCA SECTION, continued													
La Pilona Formation													
Site 46, S/D bed: 138/19													
02A ^b	1.12	8.89	20.4	-48.2	8.8				200?	>550	H		
03A	1.46	10.49	25.8	-64.0	2.6				200	660-680	H	20	200
04A	1.07	14.36	22.1	-52.4	7.9				200-250	>560	H		
Mean			22.3	-54.8	12.7	71.4	34.2	18.0	12.8				
Site A47, S/D bed: 138/19													
00A	1.32	6.82	349.8	-50.1	3.6				20	>580	H		
02A ^b	1.22	7.30	340.9	-34.1	4.6				20	>550	H?		
Mean			344.5	-42.3	16.8	73.9	229.3	-	-				
Site B47, S/D bed: 138/19													
01A	1.45	6.75	8.2	-54.5	2.1				20	640-660	H		
03A	1.68	7.25	359.2	-56.8	1.5				20	>560	H		
04A ^b	1.49	8.53	14.8	-55.2	2.6				20	>550	H		
Mean			7.5	-55.7	6.9	83.1	49.6	9.9	7.1				
Site A48, S/D bed: 140/20													
00A	0.66	2.80	11.2	-36.7	2.2				20	>580	H		
01A ^b	0.59	1.83	346.3	-13.6	3.6				20	>600	H		
02A	0.37	2.33	2.7	-30.2	4.7				100	>580	H	20	100
Mean			359.2	-27.2	25.5	71.4	288.6	27.8	15.1				
Site B48, S/D bed: 140/20													
01A ^b	1.10	3.02	20.7	-42.3	1.9				20-100	>550	H		
02A	1.36	3.34	24.7	-40.9	1.8				100	640-660	H		
03A	0.67	2.55	11.8	-40.0	0.8				20	>560	H		
Mean			19.0	-41.2	7.8	70.8	356.2	9.5	5.8				
Site 49, S/D bed: 140/20													
02A	0.69	8.82	16.7	-53.9	4.4				250	>560	M?-H	20	200
02B ^b	0.53	9.14	9.1	-39.4	7.6				250?	>550	M?-H		
04A	1.04	11.66	354.8	-30.5	4.2				250	>550	M?-H		
Mean			5.5	-41.6	22.2	79.6	320.4	27.2	16.6				

Table A1. Continued

Specimen	J _{NRM} (10 ⁻⁴)	Susc (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component		Min- eral	Po- lar- ity	Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	dm (°)	dp (°)	Min			Max	T _b range (°C)
ARROYO DEL AGUA BLANCA SECTION, continued														
La Pilona Formation, continued														
Site 50, S/D bed: 140/20														
01A	3.15	8.89	95.0	-74.3	1.4							N	20	200
02A _b	2.98	9.92	35.6	-45.9	5.3							N		
03A	2.45	7.98	43.1	-50.3	1.5							N		
Mean			47.8	-58.5	32.1	51.4	44.0	47.7	35.3					
Site 51, S/D bed: 115/20														
01A	2.68	5.17	246.1	44.8	0.8							R	20	530?
02A	1.45	9.39	220.6	59.6	1.0							R		
03A ^b	1.34	8.58	206.3	49.6	4.0							R		
Mean			225.3	52.6	23.2	-52.4	213.8	32.1	22.1					
Site 52, S/D bed: 120/19														
02B	0.65	1.90	228.7	36.1	3.3							R		
03B	0.48	1.88	206.8	57.8	2.5							R		
04A ^b	0.46	1.30	261.3	53.2	8.3							R		
Mean			232.6	50.9	31.5	-46.0	214.0	42.6	28.8					
Site 53, S/D bed: 141/20														
01A	0.32	3.14	226.8	42.9	2.0							R		
02A	0.27	2.99	262.4	8.5	3.4							A		
03A	0.29	3.10	86.8	-20.3	3.1							A	20	200
Site 54, S/D bed: 141/20														
01A ^b	0.46	3.57	26.3	-23.0	16.4							N		
02A ^b	0.39	2.97	41.0	-30.2	15.9							N		
03A	0.18	3.42	286.0	-21.5	4.9							N?	20	200?
Mean			33.4	-26.8	15.0	54.3	356.9	-	-					

Table A1. Continued

Table A1. Continued

Specimen	J _{NRM} (10 ⁻⁴) A/M	Susc (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component		Po- lar- ity	Viscous component	
			D(°)	I(°)	E(°)	Lat	Lon	dm	dp	T _b range (°C) Min		Max	Min
LA HIGUERITA SECTION													
Marino Formation													
Site 01. S/D bed: 137/38													
00A	0.86	9.36	300.9	6.1	1.2					200	450-500	?	A
01A	1.07	9.17	295.8	8.1	0.6					200	450-500	?	A
Mean			298.3	7.1	5.4	21.3	220.3	-	-				
Site 02. S/D bed: 137/38													
00A	3.49	10.98	186.6	-22.7	0.4					100	>550-630	?	A
01A ^c	2.24	8.78								200?	>500-530	?	R?
02A	2.52	8.90	104.5	39.6	1.0					200	550-630	?	A
Site 03. S/D bed: 105/35													
00A	0.64	13.28	204.8	37.3	6.1					200	500-550	M?	R
01A	0.39	7.84	160.2	39.4	1.1					200	500-550	M?	R
02A ^a	0.22	10.41	79.5	36.2	25.9					200?	>560	M?	R?
Mean			182.8	40.5	34.6	-79.7	125.7	-	-				200
Site 04. S/D bed: 105/35													
00A ^b	2.01	15.26	332.2	-51.2	12.5					20-100	550-600	?	N
01A	2.52	18.25	358.4	-43.1	2.7					100	>550-630	?	N
02A ^b	2.32	13.93	7.9	-31.9	1.9					20-100	>550-600	?	N
Mean			354.7	-43.0	25.4	80.7	260.0	31.5	19.5				
La Pilona Formation													
Site 00. S/D bed: 120/25													
01A	1.57	13.26	205.8 ^d	65.9 ^d	7.5					400?	500	?	R
			101.0	0.1	11.5					500	560		
02A ^a	1.50	14.83	223.2	77.6	19.8					400?	>500-550	?	R
Mean			211.7	71.9	12.7	-57.8	258.1	-	-				
Site 01. S/D bed: 145/32													
01B ^b	1.29	12.5	186.6	39.0	14.4					250?	>550	H	R
02A	1.48	8.76	235.1	46.3	1.7					~500	>640	H	R
03B	0.96	13.0	185.7	33.2	11.3					250	>630	H	R
Mean			200.2	41.7	34.4	-70.0	179.0	42.1	25.8				200-400
													450-500
													250

Table A1. Continued

Specimen	J _{NRM} A/M (10 ⁻⁴)	Susc (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component		Min- eral	Po- lar- ity	Viscous component	
			D(°)	I(°)	E(°)	Lat	Lon	dm	dp	T _b range (°C) Min			Max	T _b range(°C) Min
LA HIGUERITA SECTION, continued														
La Pilona Formation, continued														
Site 02. S/D bed: 152/26														
0201A	0.31	8.45	186.5	47.6	4.7					250?	>550	R		
0202B	0.24	8.46	200.9	48.1	7.9					250	530?-560	R	20	200
0203A	0.70	14.1	174.5	53.8	18.0					250	500?-530	R	20	200
Mean			187.3	50.30	13.8	-83.1	186.0	18.6	12.5					
Site 03. S/D bed: 140/25														
01A	0.25	8.62	179.2	62.90	7.9					200	560-580	R	20	200
02A	0.29	9.04	246.5	62.4	6.5					200-250	>550	R		
03A*	0.32	9.04	179.0	78.9	3.6					200-250	>550	R		
Mean			206.5	71.4	26.7	-60.7	260.2	46.8	40.9					
Site 04. S/D bed: 166/25														
01B	0.25	6.95	175.3	77.6	4.9					200	560-580	R	20	200
02A ^b	0.28	6.54	221.1	65.2	15.4					200?	>550	R		
03A	0.25	5.43	228.5	67.2	2.5					200-250	>550	R	20	200
Mean			214.8	71.1	16.3	-57.0	254.5	28.4	24.8					
Site 05. S/D bed: 135/29														
01A ^b	0.31	6.56	35.3	61.3	18.1					200?	>550	A		
02A	0.49	8.01	151.0	47.1	11.4					200	560-580	R	20	200
03B	0.27	8.24	152.1	45.3	16.5					200-250	>530	R	20	200-250
Mean			151.5	46.2	1.9	-64.9	26.1	-	-					
Site 06. S/D bed: 135/25														
01A	0.95	12.53	124.2	14.9	17.2					350?	530-550?	A		
02A	1.24	13.17	4.0	-36.4	2.3					100	530-560?	N		
03B	0.31	8.46	171.4	50.5	1.9					250	>580	R		
Site 07. S/D bed: 155/20														
01A ^b	0.87	8.25	345.6	-58.5	5.8					20	>550	N		
02B	1.24	9.91	347.1	-19.9	2.2					20	>630	N		
03A	1.32	6.63	9.5	-49.2	2.4					20	660-680	N		
Mean			353.6	-43.1	35.0	80.2	254.2	43.5	27.0					

Table A1. Continued

Table A1. Continued																
Specimen	J _{NRM} A/M (10 ⁻⁴)	Susc (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component			Po- lar- ity	Viscous component			
			D(°)	I(°)	E(°)	Lat	Lon	dm	dp	Min	Max		Min	Max		
LA HIGUERITA SECTION, continued																
La Pilona Formation, continued																
Site 08, S/D bed: 150/43																
01A	1.68	6.56	61.0	-17.4	1.7					20	620-640	H	N			
02A	1.53	9.25	14.3	-20.0	7.6					20	>550	H	N			
03A	1.97	11.04	11.8	-7.4	1.3					20	≥630	H	N			
Mean			28.4	-16.1	44.1	53.8	343.8	45.4	23.4							
Site 09, S/D bed: 140/20																
01A	1.49	6.66	330.2	-48.1	5.5					20	>630	H	N			
02A ^b	1.48	7.77	22.1	-39.7	7.3					20?	>550	H	N			
03A	3.02	5.95	12.3	-60.6	1.1					300	660-680	H	N	100		250
Mean			2.0	-51.8	32.8	88.2	0.1	44.8	30.6							
Site 10, S/D bed: 147/28																
01A ^b	1.22	6.80	2.8	-46.9	0.6					20	>550	H	N			
1002A	1.13	7.35	16.0	-41.9	2.0					20	>630	H	N			
1003A	1.17	7.42	13.9	-46.1	2.5					100	>640	H	N			
Mean			11.1	-45.1	8.6	78.4	349.8	10.9	6.9							
Site 11, S/D bed: 155/32																
01A ^b	0.81	3.47	65.5	-27.6	2.5					200-250	>550	M-H	N			
02B	0.98	4.13	72.5	-31.6	2.6					200-250	620-640	M-H	N	20		200
03A	0.62	3.24	67.6	-21.9	1.9					200-250	620-640	M-H	N			
Mean			68.4	-26.9	8.6	25.6	21.6	9.4	5.1							
Site 12, S/D bed: 152/27																
02A	0.26	1.98	1.1	-52.1	1.6					200-250	660-680	M-H	N			
03A	0.41	1.68	12.2	-41.5	1.3					200-250	660-680	M-H	N	20		200
Mean			7.2	-46.9	13.0	82.1	344.1	-	-							
Site 13, S/D bed: 152/29																
01A	0.41	2.31	32.9	-44.4	3.4					100	660-680	M?-H	N			
02A ^b	0.47	2.81	355.2	-44.6	1.3					20-100	>550	M?-H	N			
03B	0.41	2.76	14.8	-36.9	3.2					20	660-680	M?-H	N			
Mean			14.3	-42.9	22.1	75.1	351.6	27.4	17.0							

Table A1. Continued

Table A1. Continued													
Specimen	J _{NRM} (10 ⁻⁴)	Susc (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component		Po- lar- ity	Viscous component	
			D(°)	I(°)	E(°)	Lat	Lon	dm	dp	T _b range (°C)		Max	Min
LA HIGUERITA SECTION, continued													
La Pilona Formation, continued													
Site 14, S/D bed: 152/22													
02B ^b	0.34	2.80	351.7	-58.9	1.2					20	>550	M?-H	N
03A	0.43	3.16	23.2	-57.5	3.3					20	600-620	M?-H	N
04B	1.02	4.67	19.2	-61.6	1.7					100	640-660	M?-H	N
Mean			11.4	-60.1	13.7	78.0	65.1	20.8	15.7				
Site 15, S/D bed: 150/27													
03A	1.09	5.55	8.3	-40.5	2.5					100	660-680	M-H	N
04A	0.81	8.74	9.5	-45.3	4.5					20	660-680	M-H	N
Mean			8.8	-42.9	4.8	78.7	336.6	-	-				
Site 16, S/D bed: 153/25													
01A	0.74	3.99	31.1	-41.3	3.2					20	>550	M-H	N
02A	0.33	1.88	21.4	-5.6	1.7					250	640-660	M-H	N
Mean			25.5	-23.5	36.7	58.7	345.1	-	-				
Site 17, S/D bed: 160/30													
01B	0.41	2.18	339.1	-21.7	2.1					200-250	660-680	M-H	N
02B	0.24	-	21.3	-3.0	1.2					200-250	>630	M-H	N
03A ^b	0.24	1.46	13.0	-28.0	0.6					200-250?	>550	M-H	N
Mean			4.9	-18.5	39.6	65.9	302.9	41.2	21.4				
Site 18, S/D bed: 154/30													
01A	0.29	4.24	332.2	-57.5	5.8					250	640-660	M-H	N
02B	0.35	-	43.4	-64.0	3.8					250	>630	M-H	N
03A ^b	0.31	3.86	7.1	-65.2	9.7					250?	>550	M-H	N
Mean			4.8	-65.3	26.0	75.1	98.0	42.2	34.2				
Site 19, S/D bed: 145/32													
01A	1.26	15.7	68.9	-70.8	12.2					300?	580-600	?	N
02A	1.05	-	41.1	35.4-	1.4					20	450?	?	A
03A	0.83	10.00	4.6	-37.3	3.3					20	580-600?	?	N
Mean			22.1	-57.6	46.7	71.3	43.1	-	-				
Site 20, S/D bed: 135/30													
02A	0.56	-	176.0	44.7	2.0					200	>550	M-H	R
03A	0.70	9.27	176.1	35	4.3					200	660-680	M-H	R
Mean			176.0	39.8	9.7	-79.0	91.4	-	-				

Table A1. Continued

Table A1. Continued																	
Specimen	J_{NRM} (10^{-4})		Susc (10^{-3} SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component			Po- lar- ity	Viscous component			
	A/M			D(°)	I(°)	E(°)	Lat	Lon	dm	dp	T_b range (°C)	Min		Max	Min	Max	
LA HIGUERITA SECTION, continued																	
La Pilona Formation, continued																	
Site 21, S/D bed: 170/42																	
01B	0.37		-	186.7	40.2	3.5						200	>550	R	20	200	
02A	0.47		5.47	197.0	41.3	3.5						200	580-600	R	20	200	
03A ^b	2.63		4.10	191.8	35.1	2.6						200?	>550	R			
Mean				191.8	38.9	7.9	-74.8	157.1	9.4	5.6							
Site A22, S/D bed: 175/25																	
01A	0.68		-	-	-	-						20	>580	A			
02A	1.10		8.44	129.3	-11.2	2.6											
Site B22, S/D bed: 198/37																	
00A	0.84		2.58	236.7	48.2	2.0						100-200	>630	M-H			
01A ^b	0.60		2.10	218.4	38.3	3.5						100-200	>600	M-H	R		
04A	2.68		3.13	178.1	38.2	1.6						100-200	660-680	M-H	R		
Mean				210.0	44.2	35.8	-63.1	193.4	45.0	28.2							
Site 23, S/D bed: 150/32																	
01B	0.71		4.26	200.0	46.7	3.9						200	640-660	M-H	R	20	200
02A	0.51		4.75	200.6	37.3	3.7						200	580-600	M-H	R	20	200
03A	0.76		4.38	208.6	39.5	3.2						200	640-660	M-H	R		
Mean				203.2	41.2	9.3	-67.2	181.9	11.3	6.9							
Site 24, S/D bed: 170/25																	
01B	0.32		-	1.1	-44.2	1.8						20	>630	H	N		
02A	0.44		0.98	5.8	-42.2	1.9						20	660-680	H	N		
03A ^b	0.29		1.16	2.4	-35.7	2.4						20?	>550	H	N		
Mean				3.1	-40.7	7.3	79.8	307.2	8.9	5.4							
Site 25, S/D bed: 145/42																	
01A	0.66		-	29.3	-53.1	4.3						200	>550	H	N	20	100-200
02A ^b	0.59		5.28	30.1	-38.4	2.9						200?	>550	H	N		
03A	0.99		5.98	49.0	-27.9	3.2						200	660-680	H	N	20	200
Mean				37.2	-40.2	23.8	55.8	13.2	28.7	17.3							

Table A1. Continued

Table A1. Continued																
Specimen	J_{NRM} (10 ⁻⁴)	Susc (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component			Min- eral	Po- lar- ity	Viscous component		
			A/M	D(°)	I(°)	E(°)	Lat	Lon	dm	dp	T _b range (°C)			Min	Max	
											Min					Max
LA HIGUERITA SECTION, continued																
La Pilona Formation, continued																
Site 26, S/D bed: 150/40																
01A	0.44	3.78	196.9	51.5	4.4					200-250	>550	H	R	20	200	
02A	0.22	1.41	191.8	24.7	4.1					200	620-640	H	R	20	200	
03A ^b	0.07	0.79	207.7	25.3	7.8					200?	>550	H	R			
Mean			199.0	33.9	26.1	-67.7	165.1	29.9	17.1							
Site A27, S/D bed: 140/40																
01A	2.10	5.94	225.0	18.8	1.9					100	640-660	H	R			
02A	2.62	3.24	184.8	34.1	2.9					200	620-640	H	R			
Mean			206.3	27.9	38.7	-60.0	169.7	-	-							
Site B27, S/D bed: 140/40																
01A ^b	2.40	4.98	219.2	49.7	1.6					200?	>550	H	R			
02A	2.28	6.51	229.8	45.8	0.9					200	>630	H	R	20	200	
03B	2.09	5.50	217.2	18.3	1.0					200	660-680	H	R	20	200	
Mean			221.5	38.2	27.8	-51.5	194.0	33.0	19.5							
Site 28, S/D bed: 140/20																
01A ^b	0.63	4.64	141.5	69.4	4.6					200?	>550	H	R			
02A	0.64	4.89	215.6	72.6	2.2					200	>600	H	R	20	200	
03B	0.67	6.46	203.9	54.3	2.3					200	>600	H	R	20	200?	
Mean			190.3	68.4	27.3	-70.0	271.9	46.2	38.9							
Site 29, S/D bed: 170/25																
02A	0.20	-	235.0	33.4	6.3					200	350(450)	?	R	20	200	
03B	1.98	5.26	218.2	42.7	2.0					100-200	640-660	H	R			
04A	1.37	6.63	258.5	-50.9	3.6					20-250	>580	H	A			
Mean			227.1	38.3	16.1	-47.0	197.7	-	-							
Site 30, S/D bed: 152/27																
01A ^b	0.43	4.92	265.4	14.7	3.8					200?	>550	M?-H	A			
02A	0.71	10.37	207.9	51.5	3.2					200-250	>630	M?-H	R	20	200	
03A	0.57	6.63	206.7	43.1	2.8					200	660-680	M?-H	R	20	200	
Mean			207.2	47.3	8.4	-66.2	196.9	-	-							

Table A1. Continued

Table A1. Continued																
Specimen	J_{NRM} (10^{-4})		Susc (10^{-3} SI)	Characteristic component			Virtual geomagnetic paleopole			Characteristic component			Min- eral	Po- lar- ity	Viscous component	
	A/M			D(°)	I(°)	E(°)	Lat	Lon	dm	dp	Min	T _b range (°C)			Max	Min
LA HIGUERITA SECTION, continued																
La Pilona Formation, continued																
Site 31, S/D bed: 142/25																
01A	1.19		3.62	186.1	25.9	1.3					200	660-680	H	R		
02A ^B	0.97		4.02	202.9	46.4	1.1					200?	>550	H	R		
03A	1.94		3.78	242.2	66.9	1.0					200	660-680	H	R		
Mean				193.4	36.4	24.4	-72.5	157.0	-	-						
Tobas La Angostura Formation																
Site 32, S/D bed: 130/24																
01A	0.77		6.63	0.4	-45.5	2.9					20	560-580	H?	N		
02A	1.28		5.16	38.0	-47.4	1.4					100	600(620)	H	N		
03A ^b	0.95		5.86	10.8	-41.5	5.1					20?	>550	H	N		
Mean				5.8	-43.6	8.5	75.1	2.1	27.1	17.3						
Site 33, S/D bed: 130/20																
01A	0.48		5.06	301.5	-55.9	4.0					100	450?	?	N		
02B	0.49		4.92	335.4	-48.4	6.1					20-100	450?	?	N		
03A	0.37		4.47	325.2	-26.6	11.0					20-100	450?	?	N		
Mean				322.2	-44.4	29.7	56.6	202.8	37.4	23.5						
Site 34, S/D bed: 120/20																
01A	2.29		3.85	345.3	-44.4	1.9					20	620-640	M-H	N		
02A	1.72		3.34	16.4	-66.4	3.4					20	620-640	M-H	N		
03A ^b	1.68		3.24	25.9	-68.0	2.1					20	>550	M-H	N		
Mean				3.9	-60.9	26.7	80.6	92.9	40.9	31.3						
Site 35, S/D bed: 125/20																
01A	0.70		3.73	22.2	-68.8	7.9					20	>560?	M?	N		
02B	0.77		3.89	87.7	-57.3	7.5					20	>550	M?	N		
03A	1.25		4.24	56.1	-68.6	7.2					200-250	580-600	M?	N		
Mean				60.2	-67.3	23.3	43.3	61.0	38.9	32.4						
Site 36, S/D bed: 109/18																
01A	1.41		3.65	66.8	-45.9	3.9	32.9	34.1	-	-	20	>580	H	N		

Table A1. Continued

Specimen	J _{NRM} (10 ⁻⁴ A/M)	Susc. (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole				Characteristic component		Min- eral	Po- lar- ity	Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	Dm (°)	Dp (°)	Min T _g range (°C)	Max T _g range (°C)			Min T _h range (°C)	Max
Tobas Angostura unit															
Site 01, S/D bed: 47/12															
02A	1.06	7.22	355.4	-33.2	3.8						20	≥600	M?	N	
03A	0.88	7.72	358.9	-32.2	1.0						20	500-550	M?	N	
04A	1.05	7.38	328.2	-53.1	3.4						20	530-560	M?	N	
Mean			349.6	-40.2	26.1	76.2	246.9	31.5	19.0						
Site 02, S/D bed: 70/14															
01A	1.29	3.53	28.6	-52.2	2.3						20	620-640	H-M?	N	
02A ^b	1.57	4.09	349.8	-38.8	3.7						20	>550	H-M?	N	
03A	1.20	3.72	2.6	-43.1	4.1						20	>550	H-M?	N	
Mean			5.3	-45.8	23.5	82.5	330.1	30.0	19.1						
Río de los Pozos Formation															
Site 03, S/D bed: 48/12															
01A	1.57	3.06	11.5	-61.0	4.1						20	600-620	H-M?	N	
02A	1.75	3.37	9.2	-45.80	2.1						20	>550	H-M?	N	
03A ^b	1.63	3.66	10.6	-62.10	7.1						20	>550	H-M?	N	
Mean			10.3	-56.30	14.0	80.8	47.5	20.2	14.5						
Site 04, S/D bed: 47/12															
01A	0.30	2.54	7.0	-56.1	13.0						20	580-600	M?	N	
02A ^b	0.53	2.81	14.2	-53.7	9.8						20	>550	M?	N	
03B	0.31	2.43	9.1	-39.4	12.3						20	>550?	M?	N	
Mean			10.1	-49.8	14.2	81.1	6.7	19.0	12.7						
Site 05, S/D bed: 52/9															
01A ^c	0.13	2.70											?	A?	
02A ^c	0.21	3.37											?	N?	
03A	0.13	2.88	108.1	-63.7	3.3						20	350?	?	A?	
Site 06, S/D bed: 70/9															
01A	0.36	2.80	195.1	53.6	2.3						200	640-660	H-M?	R	20
02A ^b	0.23	1.93	170.1	35.4	6.7						200?	>550	H-M?	R	
03B	0.19	1.81	171.3	38.2	4.8						200-250	>550	H-M?	R	
Mean			177.2	42.9	21.3	-81.4	93.9	26.4	16.3						200

QUEBRADA DE LA ANGOSTURA SECTION

Table A1. Continued

Specimen	J _{NRM} (10 ⁻⁴) A/M	Susc. (10 ⁻³ SI)	Characteristic component			Virtual geomagnetic paleopole				Characteristic component		Min- eral	Po- lar- ity	Viscous component	
			D (°)	I (°)	E (°)	Lat (°N)	Lon (°E)	Dm (°)	Dp (°)	Min T _b range (°C)	Max T _b range (°C)			Min T _b range (°C)	Max
QUEBRADA DE LA ANGOSTURA SECTION, continued															
Río de los Pozos Formation , continued															
Site 07. S/D bed: 75/9															
01B	0.33	4.87	165.2	5.5	2.5					200	620-640	H-M?	R?	20	200
02B ^b	0.36	5.58	168.5	63.9	6.2					20(200?)	>550	H-M?	R		
03C	0.60	4.95	181.7	58.5	5.4					20(200?)	>550	H-M?	R		
Mean			175.7	61.4	8.3	-80.1	309.9	-	-						
Site 08. S/D bed: 75/9															
01B	0.26	2.45	177.9	45.2	2.6					200	640-660	H-M?	R	20	200
02C ^b	0.25	2.37	187.0	33.2	4.7					200?	>550	H-M?	R		
03A	0.19	2.38	173.9	55.1	4.3					200?-250	>550	H-M?	R		
Mean			180.4	44.6	18.4	-83.1	114.1	23.2	14.6						
EL MOLINO SECTION															
Tobas Angostura unit															
Site 01. S/D bed: 335/8															
01A	0.31	3.00	339.5	-60.7	13.7					20	>580	M?	N		
02A ^b	0.20	3.28	22.4	-67.6	12.9					20?	>550	M?	N		
Mean			358.1	-65.7	19.4	75.1	115.7	-	-						
Site 02. S/D bed: 320/8															
01A ^b	1.67	1.67	345.4	-55.4	1.3					20	>550	H-M?	N		
02A	2.07	1.72	338.6	-28.6	1.3					20	660-680	H-M?	N		
03A	1.55	1.30	320.8	-65.2	3.6					20	>550	H-M?	N		
Mean			336.7	-50.2	31.6	70.2	200.7	42.4	28.4						
Río de los Pozos Formation															
Site 03. S/D bed: 320/8															
01A ^b	0.36	2.32	332.7	-53.5	7.2					20	>550	H-M?	N		
02A ^b	0.31	1.63	326.4	-68.9	8.3					20	>550	H-M?	N		
03A	0.41	2.29	315.6	-65.9	5.4					20	640?-660	H-M?	N		
Mean			325.9	-62.90	13.8	61.4	167.5	21.9	17.2						
Site 04. S/D bed: 310/15															
01A	0.18	1.12	348.6	-45.1	2.3					20	640?-660	H-M?	N		
02A ^b	0.19	2.27	336.6	-53.0	5.5					20	>550	H-M?	N		
03A ^b	0.69	3.69	19.1	-34.5	10.9					20	>550	H-M?	N		
Mean			356.9	-45.6	28.6	83.4	266.1	36.3	23.1						

Table A1. Concluded

NOTES. Specimen: In the text, the first two digits indicate the site number, the next two digits indicate the sample number, and the final letter designates the specimen (e.g., 0101A). J_{NRM} = the intensity of natural remanent magnetization. Susc. = the magnetic susceptibility. In the characteristic component, directional data are corrected for tilting: D = declination, I = inclination, E = error. E is the error associated with Kirsvinck analysis (1980), or when shown in parentheses, it is the error associated with Fisher statistics (1953) in cases where Kirsvinck analysis was not used; for site means, E is α^{95} (circle of 95% confidence about the mean) when three specimens were averaged or θ (the angle between the specimen directions) when two specimens were averaged. For the virtual geomagnetic paleopole, Lat, Lon = latitude and longitude of the mean paleopole; dm, dp = semiaxes of the oval of 95% confidence about the mean paleopole. T_b range = blocking temperature range.

^a Mean specimen direction calculated using Fisherian statistics (1953) because directions were too scattered to employ Kirsvinck analysis (1980).

^b Specimen processed only at 500, 550 (600) because companion samples from same site were very stably magnetized.

^c Data from this specimen were too scattered for a meaningful estimate of the magnetization direction.

^d Direction used in calculation of mean direction per site.

Tosca de Mariño Profile, sample 1

FG	100	100	100	100	Tot	%
Qm	0	1	0	1	2	0.50
Qp	0	0	0	0	0	0.00
Pla	19	16	10	13	58	14.50
Ksp	0	0	0	0	0	0.00
Lv	78	75	83	83	319	79.75
Ls	0	2	3	0	5	1.25
Acc	0	1	1	1	3	0.75
Opa	0	2	2	0	4	1.00
Und	3	3	1	2	9	2.25

Tosca de Mariño Profile, sample 3

FG	100	100	100	100	Tot	%
Qm	13	12	15	9	49	12.25
Qp	2	1	0	2	5	1.25
Pla	30	31	40	46	147	36.75
Ksp	2	2	1	1	0	1.50
Lv	49	50	34	39	172	43.00
Ls	0	1	1	0	2	0.50
Acc	2	3	5	3	13	3.25
Opa	0	0	2	0	2	0.5
Und	2	0	2	0	4	1.00

Tosca de Mariño Profile, sample 2

FG	100	100	100	100	Tot	%
Qm	11	12	13	16	52	13.00
Qp	6	1	3	4	14	3.50
Pla	43	44	37	36	160	40.00
Ksp	0	0	0	0	0	0.00
Lv	33	33	37	35	138	34.50
Ls	0	0	0	0	0	0.00
Acc	1	3	3	2	9	2.25
Opa	2	5	3	4	14	3.5
Und	4	2	4	3	13	3.25

Tosca de Mariño Profile, sample 4

FG	100	100	100	100	100	Tot	%
Qm	14	14	7	3	11	49	9.80
Qp	2	5	2	4	2	15	3.00
Pla	24	26	32	35	36	153	30.60
Ksp	1	1	0	2	1	5	1.00
Lv	49	43	44	47	38	221	44.20
Ls	0	2	2	0	0	4	0.80
Acc	5	6	7	7	7	32	6.40
Opa	3	0	4	2	4	13	2.60
Und	2	3	2	0	1	8	1.60

Table A2. Continued

Tosca de Mariño Profile, sample 5

FG	100	100	100	100	Tot	%
Qm	6	7	8	8	29	7.25
Qp	0	0	1	1	2	0.50
Pla	46	40	43	39	168	42.00
Ksp	1	0	0	1	2	0.50
Lv	36	41	37	44	158	39.50
Ls	1	0	0	0	1	0.25
Acc	6	6	7	3	21	5.25
Opa	1	2	2	2	7	1.40
Und	3	4	2	2	11	2.75

Tosca de Mariño Profile, sample 7

FG	100	100	100	100	Tot	%
Qm	8	7	7	6	28	7.00
Qp	0	1	3	1	5	1.25
Pla	33	30	38	29	130	32.50
Ksp	0	4	1	1	6	1.50
Lv	53	52	46	53	204	51.00
Ls	0	1	0	1	2	0.50
Acc	3	2	0	4	9	2.25
Opa	2	1	1	2	6	1.50
Und	0	2	0	2	4	1.00
Carb.	1	0	3	1	5	1.25

Tosca de Mariño Profile, sample 6

FG	100	100	100	100	100	Tot	%
Qm	16	8	7	7	7	45	9.00
Qp	2	2	2	3	0	9	1.80
Pla	42	37	41	37	49	206	41.20
Ksp	2	2	3	3	4	14	2.80
Lv	31	35	37	38	36	177	35.40
Ls	0	3	1	3	0	7	1.40
Acc	2	6	4	3	1	16	3.20
Opa	3	2	3	3	2	13	2.60
Und	2	4	2	2	1	11	2.20
Carb.	0	1	0	1	0	2	0.40

Tosca de Mariño Profile, sample 8

FG	100	100	100	100	Tot	%
Qm	8	8	7	6	29	7.25
Qp	4	2	3	2	11	2.75
Pla	40	44	30	29	143	35.75
Ksp	0	1	3	0	4	1.00
Lv	41	32	48	48	169	42.25
Ls	2	2	0	0	4	1.00
Acc	1	9	4	7	21	5.25
Opa	1	0	2	1	4	1.00
Und	2	2	1	3	8	2.00
Carb.	1	0	2	4	7	1.75

Table A2. Continued

Tosca de Mariño Profile, sample 9

FG	100	100	100	100	Tot	%
Qm	11	10	10	11	42	10.50
Qp	1	0	1	2	3	0.75
Pla	41	40	33	27	141	35.25
Ksp	0	3	0	0	3	0.75
Lv	32	36	51	51	170	42.50
Ls	3	0	1	3	7	1.75
Acc	6	9	3	3	21	5.25
Opa	5	0	0	3	8	2.00
Und	1	2	1	0	4	1.00
Carb.	0	0	0	0	0	0.00

Tosca de Mariño Profile, sample 11

FG	100	100	100	100	Tot	%
Qm	4	11	7	6	28	7.00
Qp	1	1	0	1	3	0.75
Pla	23	27	24	28	102	25.50
Ksp	3	2	2	2	9	2.25
Lv	58	45	58	53	214	53.50
Ls	1	4	3	2	10	2.50
Acc	3	4	3	6	16	4.00
Opa	1	3	0	1	5	1.25
Und	1	1	1	0	3	0.75
Carb.	5	2	2	1	10	2.50

Tosca de Mariño Profile, sample 10

FG	100	100	100	100	Tot	%
Qm	10	12	9	8	39	9.75
Qp	6	4	4	3	17	4.25
Pla	29	28	27	23	107	26.75
Ksp	1	1	3	2	7	1.75
Lv	45	50	50	56	201	50.25
Ls	0	1	0	1	2	0.50
Acc	4	2	5	4	15	3.75
Opa	2	1	1	1	5	1.25
Und	1	0	0	2	3	0.75
Carb.	2	1	1	0	4	1.00

Tosca de Mariño Profile, sample 12

FG	100	100	100	Tot	%
Qm	8	7	9	24	8.00
Qp	1	1	0	2	0.66
Pla	41	41	36	118	39.33
Ksp	0	0	1	1	0.33
Lv	42	43	45	130	43.33
Ls	0	0	0	0	0.00
Acc	4	5	4	13	4.33
Opa	3	2	2	7	2.33
Und	1	0	2	3	1.00
Carb.	0	1	1	2	0.66

Table A2. Continued

Tosca de Mariño Profile, sample 13

FG	100	100	100	100	Tot	%
Qm	8	4	5	6	23	5.75
Qp	1	0	0	1	2	0.50
Pla	25	27	32	28	112	28.00
Ksp	0	0	0	0	0	0.00
Lv	53	58	57	56	224	56.00
Ls	4	1	1	2	8	2.00
Acc	4	5	2	3	14	3.50
Opa	0	2	2	1	5	1.25
Und	3	2	1	2	8	2.00
Carb.	2	1	0	1	4	1.00

Arroyo del Agua Blanca Profile, sample 15

FG	100	100	100	100	Tot	%
Qm	6	5	7	11	29	7.25
Qp	2	0	2	3	7	1.75
Pla	37	28	24	24	113	28.25
Ksp	2	1	0	2	5	1.25
Lv	42	55	60	47	204	51.00
Ls	3	6	0	2	11	2.75
Acc	6	4	3	8	21	5.25
Opa	1	1	1	0	3	0.75
Und	0	0	3	2	5	1.25
Carb.	1	0	0	1	2	0.50

Arroyo del Agua Blanca Profile, sample 14

FG	100	100	100	100	Tot	%
Qm	5	6	10	7	28	7.00
Qp	0	0	0	0	0	0.00
Pla	17	16	21	18	72	18.00
Ksp	0	0	0	0	0	0.00
Lv	69	69	60	66	264	66.00
Ls	3	5	3	4	15	3.75
Acc	1	2	1	1	5	1.25
Opa	2	0	1	2	5	1.25
Und	2	2	3	2	9	2.25
Carb.	1	0	1	0	2	0.50

Arroyo del Agua Blanca Profile, sample 16

FG	100	100	100	100	Tot	%
Qm	2	6	5	7	20	5.00
Qp	2	0	0	3	5	1.25
Pla	34	28	25	29	116	29.00
Ksp	0	0	1	0	1	0.25
Lv	50	59	60	53	222	55.50
Ls	1	1	0	0	2	0.50
Acc	6	4	8	4	22	5.50
Opa	3	1	1	3	8	2.00
Und	2	0	0	1	3	0.75
Carb.	0	1	0	0	1	0.25

Table A2. Continued

Arroyo del Agua Blanca Profile, sample 17

FG	100	100	100	100	Tot	%
Qm	1	1	0	1	3	0.75
Qp	0	1	1	0	2	0.50
Pla	17	17	24	20	79	19.75
Ksp	0	0	0	0	0	0.00
Lv	69	70	64	67	270	67.50
Ls	1	0	2	1	4	1.00
Acc	5	7	3	6	21	5.25
Opa	7	3	5	5	20	5.00
Ind	0	1	1	0	2	0.50
Carb	0	0	0	0	0	0.00

La Higuera Profile, sample b

FG	100	100	100	100	100	Tot	%
Qm	6	7	9	6	11	39	7.80
Qp	0	1	0	0	1	2	0.40
Pla	40	38	35	40	33	186	37.20
Ksp	1	1	0	0	0	2	0.40
Lv	34	43	45	37	40	199	39.80
Ls	3	6	4	3	8	24	4.8
Acc	10	0	4	9	5	28	5.60
Opa	4	1	3	2	1	11	2.20
Ind	2	3	0	3	1	9	1.80
Carb	0	0	0	0	0	0	0.00

La Higuera Profile, sample a

FG	100	100	100	100	Tot	%
Qm	1	8	8	6	23	5.75
Qp	0	0	0	0	0	0.00
Pla	36	33	29	34	132	33.00
Ksp	1	2	3	0	6	1.50
Lv	48	43	49	53	193	48.25
Ls	2	2	1	1	6	1.50
Acc	6	7	7	5	25	6.25
Opa	3	2	2	0	7	1.75
Ind	2	2	0	1	5	1.25
Carb	1	1	1	0	3	0.75

La Higuera Profile, sample c

FG	100	100	100	Tot	%
Qm	3	6	4	13	4.33
Qp	1	0	0	1	0.33
Pla	39	45	44	128	42.67
Ksp	0	2	0	2	0.67
Lv	46	37	42	125	41.67
Ls	3	4	2	9	3.00
Acc	4	5	5	14	4.67
Opa	2	0	1	3	1.00
Ind	2	1	2	5	1.66
Carb	0	0	0	0	0.00

La Higuera Profile, sample d

FG	100	100	100	100	Tot	%
Qm	7	4	7	9	27	6.75
Qp	1	0	3	0	4	1.00
Pla	24	32	27	32	115	28.75
Ksp	2	1	3	3	9	2.25
Lv	49	45	48	42	184	46.00
Ls	2	0	0	1	3	0.75
Acc	11	11	7	8	37	9.25
Opa	3	5	4	4	16	4.00
Ind	2	1	0	1	4	1.00
Carb	0	0	1	0	1	0.25

La Higuera Profile, sample f

FG	100	100	100	100	100	Tot	%
Qm	22	27	21	14	18	102	20.40
Qp	7	6	7	11	4	35	7.00
Pla	9	6	10	7	15	47	9.40
Ksp	2	3	1	0	1	7	1.40
Lv	41	36	32	43	32	184	36.80
Ls	14	14	20	15	23	86	17.20
Acc	1	4	3	5	3	16	3.20
Opa	3	1	3	2	2	11	2.20
Ind	1	3	3	3	2	12	2.40
Carb	0	0	0	0	0	0	0.00

La Higuera Profile, sample e

FG	100	100	100	100	Tot	%
Qm	14	7	11	6	38	9.50
Qp	4	4	0	1	9	2.25
Pla	35	42	48	44	169	42.25
Ksp	1	1	0	0	2	0.50
Lv	28	29	29	36	122	30.50
Ls	5	3	3	5	16	4.00
Acc	8	8	5	5	26	6.50
Opa	1	3	2	1	7	1.75
Ind	4	3	2	2	11	2.75
Carb	0	0	0	0	0	0.00

La Higuera Profile, sample g

FG	100	100	100	100	Tot	%
Qm	6	6	5	5	22	5.50
Qp	4	2	1	3	10	2.50
Pla	26	21	22	24	93	23.25
Ksp	3	2	1	1	7	1.75
Lv	42	43	52	48	185	46.25
Ls	13	20	16	16	65	16.25
Acc	3	1	0	1	5	1.25
Opa	1	0	1	1	3	0.75
Ind	2	5	2	1	10	2.50
Carb	0	0	0	0	0	0.00

Table A2. Continued

La Higuera Profile, sample h

FG	100	100	100	100	Tot	%
Qm	11	9	10	15	45	11.25
Qp	2	4	2	0	8	2.00
Pla	9	16	7	12	44	11.00
Ksp	3	2	3	3	11	2.75
Lv	51	57	49	51	208	52.00
Ls	17	6	21	12	56	14.00
Acc	4	3	3	3	13	3.25
Opa	2	2	3	2	9	2.25
Ind	1	1	2	2	6	1.50
Carb	0	0	0	0	0	0.00

La Higuera Profile, sample j

FG	100	100	100	100	100	Tot	%
Qm	13	10	6	4	15	48	9.60
Qp	1	4	3	0	1	9.0	1.80
Pla	21	20	17	21	22	101	20.20
Ksp	1	1	1	2	1	6	1.20
Lv	53	56	57	48	45	259	51.80
Ls	2	0	3	10	2	17	3.40
Acc	7	7	11	13	10	48	9.60
Opa	1	1	1	1	3	7	1.40
Ind	1	1	1	1	1	5	1.00
Carb	0	0	0	0	0	0	0

La Higuera Profile, sample i

FG	100	100	100	100	Tot	%
Qm	11	20	9	11	51	12.75
Qp	3	2	1	1	7	1.75
Pla	10	12	14	20	56	14.00
Ksp	1	0	0	0	1	0.50
Lv	58	49	55	47	209	52.25
Ls	11	8	14	16	49	12.25
Acc	1	2	2	2	7	1.75
Opa	2	2	0	2	6	1.50
Ind	3	5	5	1	14	3.50
Carb	0	0	0	0	0	0.00

La Higuera Profile, sample k

FG	100	100	100	100	Tot	%
Qm	9	15	14	13	51	12.75
Qp	3	2	2	2	9	2.25
Pla	10	9	6	10	35	8.75
Ksp	2	2	1	1	6	1.50
Lv	66	62	68	64	260	65.00
Ls	6	9	8	7	30	7.50
Acc	0	0	0	0	0	0.00
Opa	1	0	0	1	2	0.50
Ind	3	1	1	2	7	1.75
Carb	0	0	0	0	0	0.00

Table A2. Continued

La Higuera Profile, sample l

FG	100	100	100	Tot	%
Qm	23	16	18	57	19.0
Qp	7	5	2	14.0	4.7
Pla	22	29	28	79	26.4
Ksp	2	2	3	7	2.33
Lv	31	32	30	93	31.00
Ls	4	8	6	18	6.00
Acc	6	4	6	16.0	5.33
Opa	4	2	3	9	3.00
Ind	1	2	4	7	2.33
Carb	0	0	0	0	0.00

La Higuera Profile, sample n

FG	100	100	100	100	Tot	%
Qm	8	5	10	6	29	7.25
Qp	0	0	0	0	0	0.00
Pla	59	55	56	66	236	59.00
Ksp	0	0	0	0	0	0.00
Lv	9	18	14	7	48	12.00
Ls	2	1	1	1	5	1.25
Acc	20	21	19	18	78	19.50
Opa	2	0	0	1	3	0.75
Ind	0	0	0	1	1	0.25
Carb	0	0	0	0	0	0.00

La Higuera Profile, sample m

FG	100	100	100	100	Tot	%
Qm	9	9	10	12	40	10.00
Qp	0	0	1	0	1	0.25
Pla	59	62	50	60	231	57.75
Ksp	0	0	0	0	0	0.00
Lv	15	12	13	8	48	12.00
Ls	4	5	8	3	20	5.00
Acc	12	10	16	15	53	13.25
Opa	0	1	1	1	3	0.75
Ind	1	1	1	1	4	1.00
Carb	0	0	0	0	0	0.00

La Angostura Profile, sample o

FG	100	100	100	100	Tot	%
Qm	10	7	10	9	36	9.00
Qp	0	0	0	0	0	0.00
Pla	45	54	49	49	197	49.25
Ksp	0	3	0	1	4	1.00
Lv	23	21	17	20	81	20.25
Ls	4	1	3	3	11	2.75
Acc	17	12	21	17	67	16.75
Opa	0	0	0	1	0	0.25
Ind	1	2	0	1	4	1.00
Carb	0	0	0	0	0	0.00

La Angostura Profile, sample p

FG	100	100	100	100	Tot	%
Qm	10	5	7	7	29	7.25
Qp	1	0	0	0	1	0.25
Pla	65	69	70	68	272	68.00
Ksp	0	0	0	1	1	0.25
Lv	16	17	15	16	64	16.00
Ls	2	2	0	1	5	1.25
Acc	5	6	7	6	24	6.00
Opa	1	1	1	1	4	1.00
Ind	0	0	0	0	0	0.00
Carb	0	0	0	0	0	0.00

La Angostura Profile, sample q

FG	100	100	100	100	Tot	%
Qm	2	3	1	2	8	2.00
Qp	0	0	0	0	0	0.00
Pla	62	69	67	66	264	66.00
Ksp	0	0	0	0	0	0.00
Lv	14	9	18	14	55	13.75
Ls	7	2	3	4	16	4.00
Acc	13	15	8	12	48	12.00
Opa	2	2	1	1	6	1.50
Ind	0	0	2	1	3	0.75
Carb	0	0	0	0	0	0.00

NOTES: FG = framework grains; Qm = monocrystalline quartz grains; Qp = polycrystalline quartz grains; Pla = plagioclase feldspar grains; Ksp = potassium feldspar grains; Lv = lithic volcanic grains (include volcanic glass fragments, glass shards, and pyroclastic rock fragments); Ls = lithic sedimentary and metasedimentary grains; Acc = accesory minerals (mainly amphibole, pyroxene and mica); Opa = opaque minerals; Und = undifferentiated grains; Carb = carbonate grains. Each column is the categorization of 100 counted grains where Tot = total number of grains counted. In Tosca de Mariño and Arroyo del Agua Blanca profiles, samples 1-16 = Mariño Formation and sample 17 = La Pilona Formation; in La Higuerita profile, samples a-b = Mariño formation, samples c-l = La Pilona formation, and samples m-n = Tobas Angostura unit; in Quebrada de la Angostura profile, samples o-q = Tobas Angostura and Río de los Pozos formations. Samples from the Mariño formation have an average modal composition of $Qt_{10.5}F_{37}L_{52.5}$ (sample 1, a coarse-grained sandstone, is excluded from this calculation because its modal composition is interpreted to reflect the compositional dependance of grain size); samples from the La Pilona formation: $Qt_{14.5}F_{26}L_{59.5}$; samples from Tobas Angostura and Río de los Pozos formations: $Qt_{8.5}F_{70.5}L_{21}$. Quantitative detrital modes calculated from point counts of thin sections were recalculated to 100% as the sum of Qt (Qm + Qp), F (Pla + Ksp) and L (Lv + Ls), as outlined by Dickinson (1985).