ABSTRACT

The Andean Orogen is the type-example of an active Cordilleran style margin with a long-lived retroarc fold-and-thrust belt and foreland basin. Timing of initial shortening and foreland basin development in Argentina is diachronous along-strike, with ages varying by 20–30 Myr. The Neuquén Basin (32°S to 40°S) contains a thick sedimentary sequence ranging in age from late Triassic to Cenozoic, which preserves a record of rift, back arc and foreland basin environments. As much of the primary evidence for initial uplift has been overprinted or covered by younger shortening and volcanic activity, basin strata provide the most complete record of early mountain building. Detailed sedimentology and new maximum depositional ages obtained from detrital zircon U–Pb analyses from the Malargüe fold-and-thrust belt (35°S) record a facies change between the marine evaporites of the Huitrin Formation (ca. 122 Ma) and the fluvial sandstones and conglomerates of the Diamante Formation (ca. 95 Ma). A 25–30 Myr unconformity between the Huitrin and Diamante formations represents the transition from post-rift thermal subsidence to forebulge erosion during initial flexural loading related to crustal shortening and uplift along the magmatic arc to the west by at least 97 ± 2 Ma. This change in basin style is not marked by any significant difference in provenance and detrital zircon signature. A distinct change in detrital zircons, sandstone composition and palaeocurrent direction from west-directed to east-directed occurs instead in the middle Diamante Formation and may reflect the Late Cretaceous transition from forebulge derived sediment in the distal foredeep to proximal foredeep material derived from the thrust belt to the west. This change in palaeoflow represents the migration of the forebulge, and therefore, of the foreland basin system between 80 and 90 Ma in the Malargüe area.

INTRODUCTION

The Andes Mountains formed as a result of the subduction of oceanic plates under South America (e.g. Dewey & Bird, 1970; Allmendinger et al., 1997). Broadly, three different kinematic regimes have been observed in the Andes: (i) back-arc extension as a result of the rate of slab rollback exceeding the margin normal component of ‘absolute’ velocity of the overriding plate; (ii) dominant strike-slip with local transtension to transpression during periods of oblique convergence and (iii) contractional deformation caused by the margin normal component of ‘absolute’ velocity of the overriding plate exceeding the rate of slab rollback (e.g. Schellart, 2008). Such regimes have resulted in different patterns of deformation and exhumation within the fold-and-thrust belt and/or volcanic arc which in turn control subsidence and sedimentation in the associated retroarc basin. One way to reconstruct kinematic regimes and plate behaviour is through investigation of the sedimentary record, which tends to be better preserved than the fold-and-thrust belt. Reconstructing the initiation of crustal shortening, erosion, and foreland basin deposition is essential for understanding the timing and rate of exhumation which can then be used to constrain plate-scale geodynamic models of the Andes.

Timing of the onset of contractional deformation along the Andean margin (Fig. 1) has been a topic of debate since Steinmann (1929) first defined three phases of Andean shortening in Peru: Peruvian during the Late Cretaceous, the Eocene Incaic and the Miocene to recent Quechua phases. Large discrepancies (tens of millions of years) exist in the literature concerning the timing of initial shortening and foreland basin deposition between the central Andes in Bolivia and the Patagonian Andes in Southern Chile and Argentina. In the Bolivian Altiplano, Sempere et al. (1997) suggest that foreland basin subsidence started in the Late Cretaceous (ca. 85 Ma). Foreland basin deposits preserved in the Eastern Cordillera of central Bolivia suggest that flexurally driven subsidence began in the latest Cretaceous–early Cenozoic (Horton & Decelles, 2001; Decelles & Horton, 2003). In Northern and Central
Argentina, well-exposed Cenozoic strata record an active eastward-migrating foreland basin system that initiated in the Palaeocene (Jordan et al., 1983; Carrapa et al., 2011, 2012; Decelles et al., 2011). To the west, in the Atacama Desert of Chile, growth strata in foredeep and wedge-top deposits record Late Cretaceous shortening (Mpodozis et al., 2005; Arriagada et al., 2006). In the same area, inversion structures documented by Martínez et al. (2012, 2013) also record shortening near the Cretaceous–Palaeocene boundary. In the southern Patagonian Andes, Fildani et al. (2003) proposed that the abrupt occurrence of sandstone in the Punta Barra Formation signified the initiation of contractional deformation at 92 ± 1 Ma.

In the Neuquén Basin (32°S to 40°S; Fig. 2) of Argentina, early scholars (e.g. Keidel, 1925; Groeber, 1929) used the change from marine to fluvial deposition in the Late Cretaceous as evidence for initiation of shortening and foreland basin sedimentation. However, subsequent studies using various methods in different parts of the basin have placed the initiation of foreland basin deposition in the Albian (ca. 115; Howell et al., 2005), Cenomanian (ca. 100 Ma; Cobbold & Rossello, 2003) and Maastrichtian (ca. 70 Ma; Barrio, 1990a). Recent geochronological studies (Tunik et al., 2010; Di Giulio et al., 2012) implementing detrital zircon U–Pb technique support a Late Cretaceous (100 ± 8 Ma) age of foreland basin initiation in the central and southern Neuquén Basin. No study to date has established age constraints in the northern-most Neuquén Basin.

The goal of this study is to determine the timing of foreland basin initiation in the northern part of the Neuquén Basin, using a multidisciplinary approach that integrates sedimentology, petrology and detrital zircon U–Pb geochronology to characterize the stratigraphy, depositional environments, and provenance patterns of Cretaceous to Lower Cenozoic strata in the Malargüe area (35°S). The northern Neuquén Basin is particularly well suited for this study because it marks a major orographic transition characterized by a decrease in topography, crustal thickness and total shortening (Kley & Monaldi, 1998; Giambiagi et al., 2012). The study area lies between the Patagonian Andes and Northwestern Argentinian/Bolivian Andes where significantly different ages of foreland basin strata have been recorded; foreland basin deposition initiated at 92 ± 1 Ma in the Patagonian Andes (Fildani et al., 2003; Romans et al., 2010), whereas foreland basin deposits have been documented at ca. 65 Ma in the Argentinian/Bolivian Andes (Horton & Decelles, 2001; Decelles & Horton, 2003; Carrapa et al., 2011, 2012; Decelles et al., 2011). Understanding the basin evolution and timing of foreland basin initiation in the Malargüe area will fill an important gap in timing constraints along the Cordilleran margin.

**GEOLOGICAL HISTORY**

**Crustal assembly**

The basement of Southern Argentina consists of the Chilenia, Cuyania, Pampian and Patagonia terranes, which were accreted onto the southwestern margin of
Gondwana (Rio de Plata and Amazonia cratons; Fig. 3; Ramos, 2010 and references therein). The Rio de la Plata craton (Fig. 3) is comprised of igneous and metamorphic rocks with zircon U–Pb ages ranging from 2200 to 2000 Ma (Rapela et al., 2007). The Pampian terrane, which contains ca. 1100–1200 Ma basement (Rapela et al., 2007), collided with the Rio de la Plata craton in latest Proterozoic–early Cambrian time (Pampean Orogeny, Ramos, 1988; Rapela et al., 1998; Ramos, 2010). Cambrian to Ordovician subduction along the western Gondwana margin led to development of the Famatinian magmatic arc (Pankhurst et al., 1998; Sato et al., 2003) and the subsequent collision and accretion of the Cuyania terrane (1360–1400 and 1070–1200 Ma; Naipauer et al., 2010) onto the Pampian terrane (Thomas & Astini, 1996, 2003; Ramos, 2004). Following the collision of Cuyania, Silurian-Devonian strata were deposited in a foreland basin (Caminos, 1979; Astini et al., 1995), which was then metamorphosed and deformed during a collision with the Chilenia terrane in the Late Devonian (Ramos, 1999; Willner et al., 2008). After this collision, a magmatic arc and accretionary prism developed along the western and southern margin of Gondwana between 300 and 250 Ma (Hervé et al., 1988; Willner et al., 2005, 2008).

The final assembly of South America with Patagonia is not well constrained, but there is some consensus about its para-autochthonous character and either its final amalgamation (Ramos, 1984, 2008; Rapalini, 2005; Pankhurst et al., 2006; Rapalini et al., 2013) or intense intraplate deformation (Gregori et al., 2008, 2013) in early Permian time. Gondwanan-Patagonian basement (AP and NPM; Fig. 3) includes an accretionary prism (AP) located on the western margin of South America, which received sediment from the east (Willner et al., 2008), arc batholiths and metamorphic basement assemblages associated with the accretion of the Patagonian terrane along the southern margin of Gondwana in the Permian (peak metamorphism 375–310 Ma; Pankhurst et al., 2006). Widespread magmatism associated with the collision of Patagonia and subsequent orogenic collapse led to the formation of abundant Early Triassic rhyolites and granites known as the Choiyoi igneous complex (Kay et al., 1989;
The Neuquén Basin contains a ca. 5000–7000-m-thick sedimentary succession of Upper Triassic to Miocene strata (Fig. 4; Vergani et al., 1995; Howell et al., 2005). In the north, the basin forms a narrow belt in the Margvargui fossil-and-thrust belt, whereas south of 36°S, the basin expands eastward ca. 250 km, forming a large triangular embayment (Fig. 2). The Neuquén Basin is located in a retroarc position and is characterized by a complex tectonic history that includes: (i) a Late Triassic–Middle Jurassic east–west-trending rift basin along the boundary between the Chilenia, Cuyania, Pampian and Patagonia basement blocks (Fig. 3); (ii) a middle Jurassic accretionary prism, Chilenia and the North Patagonian Massif. The main age range includes Jurassic–Cretaceous plutonic and volcanic arc-related activity associated with the subduction of the Farallon and other oceanic plates beneath South America.

**Stratigraphy of the Neuquén Basin**

The Neuquén Basin contains a ca. 5000–7000-m-thick sedimentary succession of Upper Triassic to Miocene strata (Fig. 4; Vergani et al., 1995; Howell et al., 2005). In the north, the basin forms a narrow belt in the Malargüe fold-and-thrust belt, whereas south of 36°S, the basin expands eastward ca. 250 km, forming a large triangular embayment (Fig. 2). The Neuquén Basin is located in a retroarc position and is characterized by a complex tectonic history that includes: (i) a Late Triassic–Middle Jurassic east–west-trending rift basin along the boundary between the Chilenia, Cuyania, Pampian and Patagonia basement blocks (Fig. 3); (ii) a middle Jurassic accretionary prism, Chilenia and the North Patagonian Massif. The main age range includes Jurassic–Cretaceous plutonic and volcanic arc-related activity associated with the subduction of the Farallon and other oceanic plates beneath South America.

The Neuquén Basin is comprised of the Allen, Loncoche, Jaguil, Roca and Picara formations (Fig. 5; Barrio, 1990a). The dominance of continental and deltaic facies, and of pyroclastic material in the central Andes sector, recorded by Aguirre-Urreta et al. (2011), suggests close proximity to the volcanic arc in the latest Cretaceous. In the central and southern sections of the Neuquén Basin, the Malargüe Group is dominated by marine deposits, comprising mainly shales, limestones and evaporites (Barrio, 1990b). These rocks correspond to the first Atlan-
tic transgression into the basin (Weaver, 1927; Uliana & Dellapé, 1981), and define a regional change in the basin slope associated with eustatic sea level rise (higher elevations to the west and lower to the east; Barrio, 1990a). At the top of the Malargüe Group, in the Malargüe area, a regional unconformity provides evidence for deformation and erosion after ca. 60 Ma; based on cross-cutting relationships between faults and volcanic units, this unconformity reflects Eocene and Miocene–present deformation related to modern Andean uplift (Legarreta et al., 1989; Ramos & Folguera, 2005; Giambiagi et al., 2008).

METHODS

This study reports new sedimentological and zircon U–Pb data from the Agrio, Huitrín and Diamante formations and the Malargüe Group, which are used to determine depositional environments, stratigraphic relations, provenance and maximum depositional ages throughout the Cretaceous and early Cenozoic. Three areas were targeted for this study, based on previous geological mapping in the region (Nullo et al., 2005), from west to east: the Los Angeles (LA), Bardas Blancas (BB) and Malargüe West (MW; Fig. 6). Fieldwork included a detailed description of outcrops and measurements (dm scale) of stratigraphic sections beginning at least 200 m below the contact between the 136–128 Ma Agrio Formation (Lazo et al., 2005; Archuby et al. 2011) and 128–125 Ma Huitrín Formation (Veiga et al. 2005) and continuing through the 98–76 Ma Diamante Formation (Leanza et al., 2000, 2004), and the 72–56 Ma Malargüe Group (Barrio, 1990b; Aguirre-Urreta et al., 2011) where present (Figs 6 and 7). The studied sections span an E–W distance of ca. 50 km and a N–S distance of ca. 45 km (Fig. 6).

Twenty-nine standard petrographic thin sections from the Los Angeles, Bardas Blancas and Malargüe West sections were stained for Ca- and K-feldspars and point-counted (450 counts per slide) according to the Gazzi–Dickinson method (G–D) (Gazzi, 1966; Dickinson, 1970; Ingersoll et al., 1984) to assess bedrock provenance of the Cretaceous–lower Cenozoic deposits of the study area. Petrographic parameters are listed in Appendix S1(A), and recalculated modal data, as well as member averages and standard distributions, are listed in Table 1. Ternary diagrams with total quartz-feldspar-lithic (QFL) and quartz-plagioclase-potassium feldspar (QPK) are shown in Fig. 9 (Dickinson et al., 1983). None of the samples
counted had more than 100 lithic fragments to allow for a separate quantitative analysis of the lithic fraction, however, qualitative descriptions of the lithic fragments are provided based on the available data.

To better constrain the maximum depositional age and provenance of the Upper Mesozoic–Lower Cenozoic sedimentary units in the Malargüe fold-and-thrust belt, systematic sampling of key formations for U–Pb detrital zircon analysis was undertaken. Samples were processed and analysed following the procedures outlined in Gehrels et al. (2006, 2008). The analytical data and a more detailed methodology are reported in Appendix S2(A,B).

The age data are displayed in relative age-probability diagrams using the routines in Isoplot (Fig. 10; Ludwig, 2008). Maximum depositional ages were calculated by taking the youngest age components, generally three or more grains (only two for the Huitrín Formation were available), that overlap in age within error and calculating a mean and standard deviation taking into account the original uncertainty in the grain age based on recommendations from Dickinson & Gehrels (2009).

RESULTS AND INTERPRETATIONS

Facies analysis, palaeocurrents and depositional environments

Sedimentological descriptions and interpretations are based on three detailed stratigraphic sections (Fig. 7a–c). Lithofacies identified in the measured stratigraphic
sections are based on the lithofacies codes of Miall (1978)
with some modifications (Table 2) and carbonate facies
description are based on the carbonate naming scheme of
Dunham (1962) as modified by Embry & Klovan (1971)
(Table 2). Palaeocurrent directions were determined for
the Diamante Formation using limbs of trough cross-
strata (method I of Decelles et al., 1983).

Agrio Formation

Description: The upper 200 m of the Agrio Formation was
measured in the Los Angeles (LA), Bardas Blancas (BB)
and Malargüe West (MW) areas to give context and assess
facies changes in the overlying Huitrín and Diamante for-
mations (Fig. 5). In all three measured sections, the Agrio
Formation consists of thinly laminated to thinly bedded,
calcareous mudstone (Fcl; see Table 2 for all lithofacies
descriptions) to fossiliferous (containing bivalves and
ammonites) packstone-grainstone (Lf) arranged in 5–10-
m-thick coarsening upwards packages (Fig. 8a). Mud-
stone throughout all three sections is thinly laminated
(Fcl) and ranges from black and organic rich with sparse
calcareous nodules (Fig. 8a), generally cored by ammon-
ite fossils, to light grey-white and carbonate rich (Lf)
with abundant bivalves that outcrop in resistant tabular, lat-
erally continuous limestone beds (Fig. 8b). Ammonites are
common, and marine reptiles and other open marine
fauna have also been documented within the laminated
siltstone and shale in the BB area (Spalletti et al. 2001;
Lazo et al., 2005). Over the 200 m of measured section,
the facies undergo a transition from shale-dominated to
limestone-dominated, becoming lighter in colour, and
contain more abundant bivalve fossils. Discontinuous
0.5–1-m-thick layers of medium-grained, massive sand-
stone (Sm) are present in the southernmost (BB; Fig. 6)
area; four separate Sm bodies are located about 50 m
below the contact with the overlying Huitrín Formation
(Fig. 7b).

Interpretation: Laminated siltstone and shale (Fcl) are
interpreted to have been deposited by suspension settling
in open marine water where waves did not interact with
the seabed (Burchette & Wright, 1992). The association
of black laminated shale (Fcl) with well-preserved fossils
suggests deposition in a low energy open marine environ-
ment. The dark colour indicates high organic content that
is typical of deposition in a poorly oxygenated or com-
pletely anoxic environment (Droste, 1990). The presence
of ammonites indicates normal marine salinity and oxy-
genation in the upper part of the water column (Batt,
1993).

The upper ca. 1–2 m of each coarsening upward pack-
age contain fossiliferous limestone beds, which range in
texture from wackestone to grainstone. The increase in
concentration of large (>2 mm) fossil fragments, and a
decline in micritic matrix up section may indicate sort-
ing by bottom currents (Aigner, 1985; Faulkner, 1988), or
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</tbody>
</table>

relatively rapid deposition (high primary productivity) of large invertebrate faunas (Schlager, 1991). Overall, the measured sections comprise repeating shallowing upwards cycles in a shallowing upwards sequence (Fig. 7a–c). Based on facies and fossils within these areas the Agrio Formation is interpreted to have been deposited in an open marine environment.

**Huitrín Formation**

**Description:** The thickness of the Huitrín Formation is highly variable, ranging from <50 m in the BB area to 200 m in the LA area (Figs 6 and 7). The contact between the Agrio Formation and the overlying Huitrín Formation is covered in all three areas, but appears to be conformable based on uniform bedding dips and general continuity along-strike. Also, biostratigraphic (Lazo & Damborea, 2011) and maximum depositional ages from this study (discussed later) suggest that there is no significant time gap between the two formations.

The contact between the Agrio and Huitrín formations in the BB and LA areas (Fig. 6) is marked by a change from white, thinly laminated calcareous shale (Fsl; see Table 2 for lithofacies codes) below a covered 2–5-m-thick interval, to brown, red and yellow rippled siltstone (Sr) and mudstone. The basal Huitrín Formation in the LA and MW areas contains ca. 10 m package of massive green, yellow and red siltstone (Fsml) and mudstone with minor bedded gypsum nodules. The overlying ca. 20 m of this formation transitions from laminated mudstone (Fsml) with minor gypsum, to dominantly gypsum with minor mudstone, and finally 60 m of massive gypsum (Em) with locally preserved thin laminations in both the LA and MW areas, and 5 m of massive gypsum in the BB area (Fig. 8c,d). Thinly laminated gypsum beds (El) are wavy and irregular, with minor interbedded siltstone, shale and micrite (Fsml and Fcl). A 2-m-thick section of thinly laminated, sandy limestone (Fsl) near the top of the Huitrín Formation (12 m below the contact with the overlying Diamante Formation; Fig. 7a) was sampled for detrital zircon U–Pb analysis in the LA area.

In the BB and MW areas (Fig. 7b,c), the massive gypsum gradually transitions to thinly laminated gypsum (El) that in turn is overlain by a 10-m-thick interval of resistant, tabular, laterally continuous, thin medium-bedded, fossiliferous (bivalves; Lf) and intraclastic (Ls), and peloidal (Lp) wacke-grainstone, which is locally dolomitic, and contains minor interbedded gypsum (laminated and nodular). In general, the gypsum layers appear to be thickest in the westernmost section (LA, 80 m) and thin to the south (5 m) and east (40 m), where it is potentially replaced by limestone and laminated siltstone and mudstone.

**Interpretation:** The association of calcareous shale (Fcl) and both laminated (El) and massive (Em) evaporate facies suggest deposition in a low energy, restricted environment (James & Kendall, 1992). Thinly laminated siltstone and mudstone require a proximal clastic source area and the combination of horizontal laminations and minor current ripples implies generally slow velocities and a unidirectional current (Harms et al., 1975). We interpret the interbedded siltstone, mudstone and gypsum as being deposited in a playa mudflat (e.g. Flint, 1985; Hartley et al., 1992). The presence of fine-grained, laminated beds and gypsum is indicative of an arid, or at least episodically dry environment (e.g. Hardie et al., 1978).

The transition from siltstone and claystone dominated to gypsum dominated suggests an overall decrease in clastic detritus. The massive (>5 m with no discernible laminations) gypsum (Em) beds indicate a restricted lagoonal environment, whereas the interbedded laminated gypsum (El), limestones (Lm), algal mats (Fig. 8d), and nodular gypsum (En), record an open, supratidal environment (Butler et al., 1982). The variability in lithologies is a product of fluctuating water chemistry, biological activity and clastic sedimentary input.
The limestones in the BB and MW areas are dominantly fossiliferous grainstones (Lf) composed of a single type of bivalve fossil, indicating a high-productivity, low-diversity ecosystem, similar to those seen in other parts of the Neuquén Basin and described in detail by Lazo & Damborenea (2011). Intraclasts (Li) likely require that the limestone episodically experienced strong wave and/or storm disturbance (James & Choquette, 1984). Peloidal grainstones (Lp) — comprised of faecal pellets generated by mud ingesting organisms and micritized by micro-organisms — are indicative of relatively low energy, protected lagoonal settings (Reid et al. 1992).

During the deposition of the Huitrín Formation, the northern Neuquén Basin was a restricted marine environment where the rate of evaporation outpaced recharge allowing for the precipitation of gypsum. The massive gypsum (Em) facies indicates a semi-continuous connection to normal salinity sea water, which is required to maintain constant gypsum deposition without other common evaporite minerals (e.g. halite, dolomite and carbonate). This was combined with a lack of clastic sedimentation in the basin and slow, steady subsidence (Handford, 1991; Warren, 1991).

Diamante Formation

The contact between the Huitrín and Diamante formations is a paraconformity in the Malargüe area. The contact is generally covered, but where it is exposed (BB area) it is a sharp contact between either tabular limestone or thinly laminated carbonate rich siltstone and medium-grained ripple cross-laminated (Sr) and trough cross-stratified (St) sandstone.

The Diamante Formation is informally divided into three units based on dominant grain size. The lower, middle and upper members are composed of generally fine-, coarse- and fine-grained sandstone, respectively, which
are interpreted to have been deposited under different flow conditions in varying environments, described below (Fig. 7b,c).

**Lower Diamante Formation**

*Description*: The lower Diamante Formation is variable between sections, but in general is fine-grained (dominantly siltstone to medium sand size) and contains minor trough cross-stratification (St), horizontally laminated (Sh), ripple cross-laminated (Sr) and massive (Sm and Fsm) sandstone. Only the basal 80 m of the Diamante Formation is preserved in the LA section (Fig. 7a), with the majority of the section having been removed by erosion or covered by Miocene volcanic rocks (Fig. 6). The lithology is dominated by red, trough cross-stratified (St) and horizontally laminated (Sh), medium- to thick-bedded, fine- to medium-grained sandstone. The Diamante Formation in the LA area coarsens upwards over the first 10 m to thickly bedded lenses of minor gravel-pebble conglomerate (Gt), which fines upwards to trough cross-stratified (St), horizontally laminated (Sh) sandstone with ripple cross-laminated (Sr) tops (Fig. 7a–c). Beds are lenticular on the outcrop scale generally ranging from 2 to 5 m wide and 1 to 3-m thick.

The lower Diamante Formation is ca. 100-m thick in the BB area and ca. 150 m in the MW area (Fig. 7b,c). Facies observed in the BB area are similar to those described in the LA area, although the lenticular nature of the beds is less pronounced at the outcrop scale, with single packages up to 20 m wide. The base of beds tend to be erosive with medium-grained trough cross-stratified (St) and ripple cross-laminated (Sr) sandstone with red siltstone and mudstone rip-up clasts that fine upwards over 4-5 m to massive, fine-grained sandstone to siltstone (Sm and Fsm). Palaeocurrent measurements taken from the limbs of troughs indicate west-directed palaeoflow in both the LA and the BB areas (Fig. 7a,b).
In the MW area, the base of the Diamante Formation has a thin (ca. 50 m), laterally discontinuous (ca. 100 m wide) interval of medium red to light green, massive fine-grained sandstone and siltstone (Fsm) with abundant carbonate nodules. The fine interval coarsens upwards into sandy, lenticular facies similar to that observed in the BB area with medium-grained trough cross-stratified (St) and ripple cross-laminated (Sr) sandstone with red siltstone and mudstone rip-up clasts that fine upwards over 3–5 m to massive and fine-grained to siltstone (Fsm).

Interpretation: Trough cross-bedded sandstones within lenticular sand bodies are interpreted to be...
deposited by the migration of dune bed forms within fluvial channels (Miall, 1996). Massive fine-grained sandstone and siltstone lithofacies (Sm and Fsm) commonly contain evidence for weakly to moderately developed soils including root casts, mottling, absence of stratification, and carbonate nodules, which are interpreted as floodplain deposits (Retallack, 1988). The red colour and abundant carbonate nodules indicate an oxidized subsurface and an arid to semi-arid environment (Hardie et al., 1978), whereas green units likely formed under reduced conditions (Bowen & Kraus, 1981). The close vertical association between channel and floodplain deposits suggests deposition onto a fluvial plain (Miall, 1996), which could be either representative of meandering or braided river systems (Miall, 1978; Smith, 1987). The prominence of the palaeosol facies in combination with minor trough cross-bedded sandstones, and climbing ripples suggests a meandering fluvial environment (Miall, 1978).

Table 2. Lithology codes

<table>
<thead>
<tr>
<th>Facies code</th>
<th>Lithofacies</th>
<th>Sedimentary structures</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lf</td>
<td>Limestone, fossiliferous</td>
<td>Generally massive, sometimes graded &gt;2 m fossils</td>
<td>Shallow marine to shelf deposits</td>
</tr>
<tr>
<td>Li</td>
<td>Limestone, intraclastic</td>
<td>Generally massive, sometimes graded</td>
<td>High energy, shallow marine</td>
</tr>
<tr>
<td>Lp</td>
<td>Limestone, peloidal</td>
<td>Thinly bedded</td>
<td>Low energy, protected lagoonal setting</td>
</tr>
<tr>
<td>Lo</td>
<td>Limestone, oolitic</td>
<td>Generally massive, sometimes graded</td>
<td>High energy, waver reworked, restricted, shallow marine or lacustrine</td>
</tr>
<tr>
<td>Em</td>
<td>Evaporite, massive</td>
<td>Massive</td>
<td>Continually recharged lagoon</td>
</tr>
<tr>
<td>En</td>
<td>Evaporite, nodular</td>
<td>Nodular</td>
<td>Supratidal banded layers</td>
</tr>
<tr>
<td>El</td>
<td>Evaporite, laminated</td>
<td>Thinly laminated to thinly bedded</td>
<td>Supratidal deposition, with alternating layers of evaporite beds, limestones, algal mats</td>
</tr>
<tr>
<td>Fcl</td>
<td>Laminated black or brown claystone</td>
<td>Thinly laminated to thinly bedded</td>
<td>Suspension settling in ponds and lakes</td>
</tr>
<tr>
<td>Fsl</td>
<td>Fine-grained sandstone, siltstone, mudstone</td>
<td>Thinly laminated to thinly bedded</td>
<td>Palaeosols, usually calcic</td>
</tr>
<tr>
<td>Fsm</td>
<td>Fine-grained sandstone, siltstone, mudstone</td>
<td>Massive, bioturbated, with common carbonate nodules</td>
<td>Palaeosols, usually calcic</td>
</tr>
<tr>
<td>Sm</td>
<td>Sandstone, fine- to coarse-grained</td>
<td>Massive, bioturbated, with common carbonate nodules</td>
<td>Palaeosols, usually calcic</td>
</tr>
<tr>
<td>Sr</td>
<td>Fine- to medium-grained sandstone</td>
<td>With small, asymmetric, 2D and 3D current ripples</td>
<td>Migration of small ripples under weak (ca. 20–40 cm s(^{-1})), unidirectional flows in shallow channels</td>
</tr>
<tr>
<td>Sh</td>
<td>Fine- to medium-grained sandstone</td>
<td>Plane-parallel lamination</td>
<td>Upper plane bed conditions under unidirectional flows, either strong (&gt;100 cm s(^{-1})) or very shallow</td>
</tr>
<tr>
<td>Sp</td>
<td>Medium- to very coarse-grained sandstone</td>
<td>Planar cross-stratified</td>
<td>Migration of large 2D ripples under moderately powerful (ca. 40–60 cm s(^{-1})), unidirectional channelized flows; migration of sandy transverse bars</td>
</tr>
<tr>
<td>St</td>
<td>Medium- to very coarse-grained sandstone</td>
<td>Trough cross-stratified</td>
<td>Migration of large 3D ripples (dunes) under moderately powerful (40–100 cm s(^{-1})), unidirectional flows in large channels</td>
</tr>
<tr>
<td>Gt</td>
<td>Pebble conglomerate, well sorted, M, bioturbated, matrix-supported, poorly sorted, unstratified</td>
<td>Clast-supported, trough cross-stratified</td>
<td>Deposition by large gravelly ripples under traction flows in relatively deep, stable fluvial channels</td>
</tr>
<tr>
<td>Gm</td>
<td>Pebble conglomerate, matrix supported</td>
<td>M, bioturbated, matrix-supported, poorly sorted, unstratified</td>
<td>Bioturbated gravel-fine pebble conglomerate</td>
</tr>
</tbody>
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**Middle Diamante Formation**

*Description:* The contact between the lower and middle Diamante Formation is marked by the first thick (>2 m) conglomerate bed of the formation and exhibits similar facies in both the BB and MW areas (not present in LA area), which have a total thickness of 240 and 440 m respectively (Fig. 7b,c). The middle Diamante Forma-
tion is thicker bedded, coarser grained (dominantly sand to pebble grain size), and contains more prominent sedimentary structures (St, Sr, Sh, Sp, Gt) than the lower and upper Diamante Formation (Fig. 7c). The middle Diamante Formation comprises an overall fining upward sequence that is characterized by 2–3-m-thick packages of both clast- (Gt; Fig. 7b,c) and matrix-supported (Gm) conglomerate, interbedded trough cross-stratified gravelly (St and Gt) sandstone (Fig. 8e) along with minor, medium-grained, massive sandstone (Sm) with abundant bioturbation and carbonate nodules, which increases in abundance up section (Fig. 7b,c). The matrix-supported conglomerate (Gm) beds tend to be massive (2–3-m thick) with nonerosive bases, whereas gravel- to pebble-rich trough cross-stratified (Gt and St) sandstone beds exhibit erosive bases with abundant mud rip-up clasts and fine upwards to medium- to fine-grained sandstone with ripple cross-lamination (Sr) and planar laminated sandstone (Sp). Trough cross-stratified beds are lenticular in nature, over a horizontal scale of 5–10 m (Fig. 8e), and commonly cut one another stacking both vertically and diagonally. Beds fine upwards ranging from pebble and gravel at the base to fine sand to siltstone size and are 2–3-m thick. Medium- to thick-bedded massive sandstones, with minor parallel laminated sandstone (Sp) at the base of the beds, are also present. Palaeocurrent measurements indicate east-directed flow in both the BB and MW areas throughout the middle Diamante Formation (Fig. 7b,c).

**Interpretation:** The interbedded massive and planar laminated sandstones (Sm and Sp) are interpreted as being deposited by high-density flows during waning flow conditions (Rasmussen, 2000). Matrix-supported conglomerate (Gm) beds are interpreted as high viscosity debris flows due to their nonerosive bases and massive bedform (Buck, 1983; Schultz, 1984). We interpret the lenticular conglomerates (Gt) as resulting from deposition in shallow, gravely, bed load channels, with (e.g. Nemec & Steel, 1984). Trough cross-stratified sandstones and pebble conglomerates (St and Gt) are

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**Fig. 8.** Photo panel of important facies and sedimentary structures in the Agrio (a, b), Huitrín (c, d), Diamante (e, f) formations and Malargüe Group (g, h). (a) Is thinly laminated organic-rich shale with nodular thickly bedded carbonate beds. (b) Bivalve packstone at the base of the Malargüe West section. (c) Massive gypsum from the Malargüe West section. (d) Laminated gypsum. (e) Lateral accretion sets from the Malargüe West section. (f) Example of bioturbation common in the Malargüe West section. (g) Contact between the red beds of the upper Diamante Formation and the yellow Malargüe Group. (b) Cross-bedded, glauconitic sandstones from the Malargüe Group.

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interpreted as being derived from the migration of dune bed forms within sand–gravel bed channels in a fluvial depositional environment (Miall, 1996). Scour-based, trough cross-stratified sandstone (St) grading into climbing ripples (Sr) represents a lower flow regime within a fluvial environment (e.g. Allen, 1963). The minor mudstones are likely to be floodplain deposits. Due to the combination of debris flow deposits, vertically stacked channel deposits, climbing ripples, and an overall lack of fine-grained material and soil formation, we interpret this facies association to indicate a mixed braided fluvial and distal alluvial fan environment (Miall, 1978).

Upper Diamante Formation

Description: The total thickness of the upper Diamante Formation, where it is completely preserved in the MW area, is 480 m, whereas in the BB area, where the top of the upper Diamante Formation is not present, the total measured thickness is 412 m. The upper Diamante Formation is dominantly massively bedded (Sm and Fsm) due to a large amount of thick, bioturbated, heavily weathered (covered in the sedimentary logs) intervals of sandstone and siltstone (Fig. 7b, c and f); although minor cross-stratification and lenticular- to wedge-shaped bedding are still sporadically preserved, especially in the uppermost deposits in the BB area (Fig. 7b). Abundant in situ and reworked carbonate nodules, at the base of conglomerate beds, are found in both the MW and BB areas (Fig. 7b,c). Rip-up clasts are also common at the base of individual beds. Amalgamated beds range from 1 to 15-m thick, and are arranged into thinning- and fining-upward patterns (Fig. 7b,c). Lenticular sand bodies are infrequent, but where present they are 1–3-m thick, up to 20 m wide, and tend to have erosive bases comprising St with very minor Gt, which then transition into Sr and Sm/Fsm. Overall, the upper Diamante Formation in the MW and BB area fines upwards with an increase in the thickness of siltstone and mudstone intervals and a decrease in sandstone with little to no conglomerate in the middle ca. 300 m and a slight increase again in the uppermost 100 m. All paleocurrent measurements taken on the limbs of trough cross-stratified sandstones in the upper Diamante Formation indicate east-directed flow in both the BB and MW areas (Fig. 7b,c).

Interpretation: The massively bedded sandstone and siltstone lithofacies (Sm and Fsm) commonly contain evidence of weakly to strongly developed palaeooslots (Retallack, 1988) including: root casts, mottling and carbonate nodules. The combination of closely stacked channel and floodplain deposits combined with the dominance and strongly developed nature of palaeosol facies, minor trough cross-bedded sandstones and climbing ripples indicates a meandering or anastomosing fluvial environment for the upper Diamante Formation (Miall, 1978; Smith, 1987).

Malargüe Group

Lower Malargüe Group

Description: A distinct shift in facies marks the contact between the Diamante Formation and the Malargüe Group with a change from the massive red beds (Fsm) of the Diamante Formation to the laminated green siltstone and shale (Fcl) of the Malargüe Group (Fig. 9g). Although most of the lower Malargüe Group (Loncoche, Jagüel and Roca Formations; Fig. 5) is recessive and exposures are limited, there are 5–10 laterally continuous, thin-bedded limestone beds (Fcl) within the green and yellow, recessive siltstone and shale (Fcl and Fsl) sequence. Contacts and facies variation are difficult to follow along-strike, but generally the section seems to be arranged in coarsening upwards packages that range from 5 to 10-m thick and define an overall fining upwards succession with a general increase in shale and decrease in sandstone (Fig. 7c). Two resistant medium-to thick-bedded marker intervals, composed of trough cross-stratified, glauconitic sandstone can be followed for hundreds of m along-strike (Fig. 9h). After a covered interval of ca. 200 m, there is a resistant ridge that contains thin-bedded oolitic grainstone (Lo) and medium-bedded fossiliferous (bivalves and gastropods) packstone (Lf).

Interpretation: The combination of laminated mudstone, siltstone and limestone at the base of the Malargüe Group in the MW are interpreted to have been deposited mostly by suspension settling in either a marginal marine or lacustrine setting. Oolitic grainstone (Lo) requires warm surface temperatures, constant wave action, and water saturated to supersaturated in calcium carbonate (Tucker & Wright, 1990) in shallow marine or lacustrine settings. Within the ridge-forming limestone beds, marine fossils have been documented by Aguirre-Urreta et al. (2011) and Parras & Griffin (2013), which along with distinctive glauconitic sandstones in the lower Malargüe Group indicate a marine environment in the latest Cretaceous–early Cenozoic.

Upper Malargüe Group

Description: In the Malargüe area, the upper Malargüe Group comprises the continental Pircala Formation that sits above a thick (ca. 300 m) covered interval, likely comprised of the recessive Roca Formation (Fig. 7c; fig. 5, Barrio, 1990a). The upper Malargüe Group is poorly exposed but a generalized section was measured in the MW area (Fig. 7c). There are discontinuous exposures of red, interbedded siltstone (Fsr) and fine medium-grained, trough cross-stratified (St), thin medium-bedded, bioturbated sandstone (Sm and Fsm) with minor carbonate nodules. Beds are lenticular over a horizontal distance of 10–20 m. The uppermost exposures are massive silt and mudstone (Fsm and Sm)
and form red and white amorphous mounds in the centre of the syncline in the MW area and are up to 10-m thick.

**Interpretation:** The thickness, mottling, root casts, absence of stratification, bleached horizons and carbonate nodules require an oxidized subsurface and an, at least episodically, arid environment (Hardie *et al.*, 1978). The minor trough cross-bedded (St) lenticular sandstones are interpreted as being formed by migrating dunes within a fluvial channel, which are at times capped by ripple cross-laminated (Sr) and planar laminated sandstone (Sp) caused by differing flow conditions/water depths on the upper parts of the migrating dune-forms (Miall, 1996). Due to the dominance and strongly developed nature of palaeosol facies, minor trough cross-bedded sandstones and climbing ripples, we interpret this association as typical of a meandering or anastomosing fluvial environment (Miall, 1978), similar to the fluvial setting interpreted for the upper Diamante Formation.

**Sandstone Petrology and Provenance**

**Description:** Sandstones from the Diamante Formation are dominantly feldspathic and lithofeldspathic with minor amounts of lithic arenite and contain abundant monocrystalline quartz (Qm) grains, with lesser amounts of polycrystalline quartz (Qp) and quartzose sedimentary lithic fragments (Qs). Plagioclase is much more abundant than K-feldspar. K-feldspar types include orthoclase (dominant), microcline and perthite. Sedimentary lithic fragments comprise dominantly chert, quartzose sandstone (Qss) and siltstone with minor limestone and shale. Volcanic grains include lathwork (Lvl), microlitic (Lvm), vitric (Lvv), felsic (Lvf) and rare mafic (Lvma) varieties. Minor amounts of micas, zircon, magnetite and Fe-oxide altered fragments are present. Modal sandstone compositions are regionally consistent, with monomineralic fractions dominated by either plagioclase or quartz, and lithic compositions dominated by volcanic grains (Fig. 9; Table 2).

Sandstones in the upper Diamante Formation are also generally feldspathic, but are more lithic-rich (especially in the MW area) and have a higher potassium feldspar to plagioclase ratio in the BB area than the lower and middle Diamante Formation (Fig. 9). The minor lithic component of the upper Diamante Formation is still dominated by volcanic lithics, but there is also a larger component (20% of total lithics) of sedimentary lithics, which include chert, siltstone, as well as a larger component of quartz in large sedimentary lithics as opposed to lower in the Diamante Formation where the majority of the quartz in lithics is plutonic.

**Interpretation:** In general, sandstones from the lower and middle Diamante Formation are plagioclase-rich, especially in the LA area (Fig. 9). The minor lithic component (<50%) is almost entirely volcanic, although there are some large (>60 μm) plutonic grains, and few to no metamorphic fragments. Quartz grains are dominantly monocrystalline with parallel extinction indicating dominantly igneous sources (Tortosa *et al.*, 1991). The presence of fine-grained plutonic fragments and plagioclase crystals that exhibit very little chemical decomposition require a local source and/or a semi-arid to arid environment. Samples from the LA area (only lower Diamante Formation) plot in the continental block uplift field of Dickinson *et al.* (1983), whereas the lower and middle Diamante Formation from the BB and MW areas plot into both the Continental Block and the Dissected Arc fields (Fig. 9). Sandstones from the upper Diamante Formation plot in the Continental Block, Dissected Arc and Recycled Orogen provenance fields (Fig. 9).
Detrital U–Pb geochronology and Provenance

Agrio Formation

Description: Detrital zircons from the Agrio Formation were collected from m-scale sand bodies interbedded with the organic-rich shale in the BB area (Figs 6 and 7b). Maximum depositional age, based on the mean age of the youngest population present within the sample is 134 ± 3.8 Ma \((n = 6)\). Peaks in the age distribution include: (i) a 150–200 Ma component, (ii) a dominant 250–300 Ma component and (iii) a minor Cambrian to Neoproterozoic component \((ca. 450–650)\). In addition, a few Mesoproterozoic \((ca. 1100 \text{ Ma})\) and Palaeoproterozoic grains \((ca. 1790–1900 \text{ Ma})\) are present (Fig. 10).

Interpretation: The majority of the grains from the Agrio Formation fall within the young end of the Gondwanan-Patagonian field (Fig. 10), which could be coming from the south or west (Franzese & Spalletti, 2001; Ramos, 2008; Naipauer et al., 2012). Jurassic- and Cretaceous-age detrital zircon populations require a western source, and could be from air-fall or minor east-directed tributaries coming off the volcanic arc. Pre-Jurassic detrital zircon grains could conceivably have originated from the north, but the abundance of Gondwanan-Patagonian and Andean-aged grains suggests that the Palaeozoic accretionary prism to the west is a likely source (AP, Fig. 3; Willner et al., 2008). Although it has been suggested that the Neuquén Basin experienced a post-rift thermal sag stage during the Late Jurassic–Early Cretaceous (Howell et al., 2005 and references therein), multiple authors (e.g. Vergani et al., 1995; Mosquera & Ramos, 2006; Garcia Morabito et al., 2011; Naipauer et al., 2012) have documented Upper Jurassic deformation on a structural high within the southern Neuquén Basin designated the Huincul Rise (Fig. 2). This arch may have continued to act as a structural high and as a sediment source to the northern Neuquén Basin throughout the Early Cretaceous (Fig. 11). A combination of thickening and coarsening of sandstone facies within the Agrio Formation to the south (Lazo et al., 2005; Archuby et al. 2011) imply a source of clastic material somewhere in the northern Pata-
gonian Massif. The southern signature (Gondwanan-Patagonian) was combined with a minor western source, which contained a combination of Jurassic–Cretaceous arc grains as well as a mix of older Palaeozoic and Proterozoic grains from the Palaeozoic accretionary prism (Willner et al., 2008).

**Huitrín Formation**

*Description:* The maximum depositional age of the Huitrín Formation, based on the youngest detrital zircon component, is $124 \pm 1.3$ Ma ($n = 2$, Fig. 10). The age distribution for the Huitrín Formation contains minor peaks at ca. 150 and 200 Ma. The majority of grain ages fall between 250 and 500 Ma with prominent components at 260, 315, 375, 420 and 480 Ma. Minor Proterozoic peaks (ca. 1000 Ma and a few older grains) are also present (Fig. 10).

*Interpretation:* The detrital zircon spectra of the Huitrín Formation differs from that of the Agrio Formation with a stronger Choiyoi signature (250–280 Ma), along with older (ca. 350–400 Ma) Gondwanan-Patagonian ages and fewer early Palaeozoic and Proterozoic grains (Fig. 10). The Huitrín Formation was likely receiving material dominantly from western sources, and noticeably less from southern sources. The change from a dominantly southwestern source in the Agrio Formation to western sources in the Huitrín Formation requires a palaeodrainage reorganization which may be related to uplift of the volcanic arc, aridification and restriction of the Neuquén Basin from open marine Pacific Ocean water in Aptian time (Figs 10 and 11).

**Lower Diamante Formation**

*Description:* The Diamante Formation shows variability in detrital zircon age distributions between different localities. The basal Diamante Formation in the LA area (Fig. 10) has a nearly identical age distribution to the underlying Huitrín Formation except for a significantly younger, $97 \pm 2$ Ma, maximum depositional age ($n = 5$, Fig. 10). The detrital zircon age spectra dominantly fall between 250 and 500 Ma, with a few Proterozoic grains. Samples from the lower Diamante Formation in the MW area contain a significant number of grains spanning ca. 230–300 Ma with almost no older grains and a very minor population of Mesozoic grains.

*Interpretation:* The detrital zircon spectra recorded from the lower Diamante Formation sample collected in the LA area require that the source material was either directly recycled from the Huitrín Formation, or that the Diamante Formation was receiving sediments from the same western source area. High plagioclase to total feld-
spar ratios (generally greater than 0.8; Fig. 9; Table 1) imply a volcanic source terrane (Dickinson, 1970), which is difficult to reconcile with an eastern source, unless Cho- iyoi igneous province rocks were exposed in the vicinity of the future San Rafael uplift, or if arid conditions allowed for the preservation and recycling of plagioclase between the Huitrín and Diamante formations. The near- depositional zircon ages from the lower Diamante Forma- tion indicate an active volcanic source. Palaeozoic–lower Triassic basement blocks uplifted immediately to the west of the MW section (Fig. 6) may have been potential sources for the pronounced Choiyoi and late Gondwanan- Patagonian signature; however, there is no evidence that they were exposed in the Late Cretaceous.

Upper Diamante Formation

Description: The detrital zircon signature of the upper- most Diamante Formation sample in the BB area is dis- tinct (Fig. 10). It does not have any grains near depositional age, or any grains younger than 200 Ma. There are major age components ranging from 200 to 280 Ma and another at ca. 400 Ma. There are also signifi- cant Proterozoic components that are not present in the two lower Diamante Formation samples.

Interpretation: The strong Choiyoi and Gondwanan- Patagonian signatures may be locally sourced immediately to the west of the BB area (Figs 4 and 6). The abundance of rhyolitic fragments in the conglomerates likely indicates a volcanic arc as a source. The re-emergence of Proterozoic grains either implies input from a northeastern source or recycling from older sedimentary units.

Lower Malargüe Group

Description: The detrital zircon signature of the lower, marine, Malargüe Group is a unimodal 69.1 ± 1.9 Ma peak; the signature also contains a few older grains, none of which form a distinct population.

Interpretation: This unimodal, ca. 70 Ma, population is syndepositional and derived from the volcanic arc. This sample does not provide any information about drainage patterns in the Maastrichtian, but it indicates that there was significant volcanic activity during Malargüe Group deposition that did not occur during deposition of the upper Diamante Formation as a result of the continued eastward migration of the volcanic arc or higher volcanic activity during that time (Fig. 11).

Upper Malargüe Group

Description: The detrital zircon signature of the upper Malargüe Group is dominated by two main peaks, one at ca. 300 Ma and another ca. 65 Ma (Fig. 10). Upon closer inspection, (Fig. 10) the younger ca. 65 Ma age com- ponent likely represents two subgroups of grains, one at 69.5 ± 2.3 Ma, similar to the syndepositional peak from the lower Malargüe Group, and a younger one, at 61.7 ± 2.1 Ma which is likely synchronous with deposition of the upper Malargüe Group.

Interpretation: The age range between 60 and 72 Ma implies continuous activity and input from the volcanic arc in the latest Cretaceous and early Cenozoic. The older, ca. 300 Ma component, is consistent with what is observed in the lower Diamante Formation from the MW area suggesting that a similar drainage pattern continued in the MW area during deposition of the Diamante For- mation and Malargüe Group.

DISCUSSION

Basin evolution in the Malargüe area

Our new sedimentological and provenance data show two major changes in depositional environments: (i) the marine-marginal marine evaporite transition between the Agrio and Huitrín formations; and (ii) the unconformity and change to fluvial-alluvial environment in the Diamante Formation. The unconformity between the Hu- itrín Formation and the Diamante Formation records a significant period (25 Myr) of erosion and/or nondeposi- tion in the Neuquén Basin. In a foreland basin system, the only depozone where uplift is expected is the forebulge (Decelles & Giles, 1996). Palaeocurrents measured in the Diamante Formation show a shift in palaeocurrent from east-directed in the lower Diamante Formation to west-directed in the middle and upper Diamante Formation. When considered in a foreland basin setting, the change in flow direction is likely a product of forebulge migration at ca. 90 Ma. During deposition of the lower Diamante Formation, the forebulge was a topographic high, supplying sediment to the distal foredeep. As the foreland basin system moved eastward, the majority of the sediment came from the fold-and-thrust belt to the west and was deposited in the proximal foredeep (Fig. 11). A switch from eastern to western sources is supported by a change in sandstone composition containing basement and volca- nic lithic detritus in the lower Diamante Formation to more sedimentary rock sources in the middle and upper Diamante Formation. Provenance data from this study do not show a major change in dominant sediment sources between the deposition of the Aptian Huitrín Formation and the Cenomanian Diamante Formation. The similarity in detrital zircon signatures can be explained by the uplift and erosion of Huitrín Formation stratigraphy which was then recycled into the lower Diamante Formation. A large shift in provenance is observed instead between the lower and upper Diamante Formations, which is here inter- preted as the product of a change from eastern to western sources. The combination of palaeocurrent, sedimentary petrography and detrital zircon analysis indicate an eastern and southern – dominantly volcanic and igneous base- ment – source for the lower Diamante Formation and a western – composed of a combination of volcanic, base- ment and sedimentary rocks – source for the middle and
upper Diamante Formation. Growth structures documented by Mescua et al. (2013) in Upper Cretaceous red beds in Chile imply movement on a thrust fault immediately to the west of the stratigraphic sections measured in the Malargüe area for this study (Fig. 2) and support our model of foreland basin initiation and migration.

According to Giambiagi et al. (2012), there has been a minimum of 20 km of shortening in the Malargüe area and the location of the thrust front migrated ca. 180 km since the Cretaceous when it was located in Chile (Fig. 2; Mescua et al., 2013) leading to an estimated 200 km of flexural wave migration of the foreland basin depozones since the Cretaceous. The geomorphic expression of the modern forebulge as documented by Niviere et al. (2013), has a length of 150 km and a height of 250 m and sits in the La Pampa High (Fig. 1; Chace et al., 2009). Based on the aforementioned shortening values and the modern location of the Andean forebulge at 35°S (Chase et al., 2009; Niviere et al., 2013), it is plausible that the Malargüe area was in the forebulge position during the Late Cretaceous, and as the flexural wave moved eastward the depositional system migrated into the foredeep and wedge top depozones. Eastward migration of the thrust front seems to have occurred mainly in the Miocene in association with potential flattening of the Nazca Plate, which led to the uplift of the San Rafael Block (Pascual et al., 2002), whereas the amount and magnitude of Cretaceous and Palaeogene shortening is less well documented and understood. Thus, if our foreland basin model is correct it raises issues associated with the amount and timing of shortening and thrust front migration, which will need to be considered in future studies.

**Along-strike variation in timing of foreland basin initiation**

Initial uplift of the South American Cordillera ranges from Late Cretaceous to Palaeogene along the length of the Andes (Fig. 1). Our study suggests that the unconformity between the Huitrín and Diamante formations marks the passage of the forebulge and that the Diamante Formation was deposited in the foredeep implying that shortening in the Malargüe area began by at least 97 ± 2 Ma (maximum age). This age is in relative agreement with the ca. 92 ± 1 Ma (Fildani et al., 2003; Romans et al., 2010) timing of foreland basin initiation in the Patagonian Andes to the south. In the central Neuquén Basin, timing of foreland basin initiation is either interpreted to coincide with the base of the Neuquén Group (100 ± 8; Tunik et al., 2010; Di Giulio et al., 2012) or with the unconformity between the Rayoso Formation and the Neuquén Group (Cobbold & Rossello, 2003), anywhere from 110 to 90 Ma. These ages also correlate with timing of foreland basin initiation in the Malargüe area.

Timing relationships north of the Malargüe area are less well known. The Neuquén Basin narrows significantly to the north, but similar stratigraphy (Huitrín and Diamante formations) is preserved in the Cordillera Principal all the way to the southern San Juan Province (31°S; Fig. 2). Although the stratigraphy is the same, it is unclear if the depositional system remained similar to the north. Growth structures identified in Cretaceous stratigraphy remain the only documented evidence of Cretaceous shortening in the Cordillera Principal at 33°S (Orts & Ramos, 2006). Mpodozis et al. (2005) and Arriagada et al. (2006) present evidence for mid to Late Cretaceous shortening in the Cordillera de Domeyko in the Atacama Basin (Fig. 1); although these units lack good age control, they potentially represent the northern equivalent of the Diamante Formation.

At 26°S, stratigraphy in the Eastern Cordillera and Salta region, ca. 100 km east of the Cordillera de Domeyko, record foreland basin initiation and migration in the Palaeocene (Fig. 1; Decelles et al., 2011), and active rifting was occurring in the back-arc region during the Cretaceous (Grier et al., 1991). The ca. 30 Myr age discrepancies between northern Argentina and central–southern Argentina could either be a result of (i) different boundary conditions along the Pacific margin during the Cretaceous, potentially due to multiple oceanic plates with variable convergence vectors; (ii) a lack of precise age control on Cretaceous terrestrial units along-strike; or (iii) expected changes in timing of foreland basin initiation as the thrust front migrates east.

**Global context**

Recent plate reconstructions since 200 Ma by Seton et al. (2012) appear to be in agreement with Early Cretaceous shortening in the central and southern Argentinian Andes. Rifting in the South Atlantic progressed from south to north in the Late Cretaceous (Daly et al., 1989) and was associated with substantial intracontinental deformation within Africa and South America (Unternehr et al., 1988; Nürnberg & Müller, 1991; Eagles, 2007; Torsvik et al., 2009; Moulin et al., 2010). The South Atlantic Ocean was opening at the latitude of Malargüe around 130 Ma with absolute motion of South America towards the northwest (Fig. 4). Complex spreading patterns in the west Pacific Ocean led Seton et al. (2012) to add two plate segments along the western margin of South America (Chasca and Catequil plates; Fig. 4), breaking up the Farallon Plate into multiple segments in the Late Cretaceous. The Chasca Plate appears to subduct nearly orthogonally beneath the western margin of South America at ca. 120 Ma (Fig. 4; Seton et al., 2012; Maloney et al., 2013). Apparent motion between the Chasca and South American plates looks to be convergent between 120 and ca. 80 Ma, which would be consistent with shortening beginning in the Malargüe area in the Late Cretaceous along with a migration of depozones. By the Cretaceous–Palaeogene boundary there is no longer differential movement between the Chasca and Catequil plates, so the Farallon Plate has a single vector along the entire margin of South America (Fig. 4). The Farallon
Plate has a similar absolute motion to South America in the latest Cretaceous to early Palaeogene which should have led to either a neutral boundary or extension (Pardo-Casas & Molnar, 1987; Seton et al., 2012), and may explain why the Malargüe area stays in the foredeep position during the entire deposition of the Diamante Formation and Malargüe Group. However, the central Andean rotation pattern of Arriagada et al. (2008) still predicts dominantly orthogonal convergence in the Cenozoic, so more work will need to be done to fully constrain plate boundary condition during Andean mountain building. The next two major pulses of shortening coincide with high convergence velocities in the Eocene and Miocene (Pardo-Casas & Molnar, 1987) and likely caused major uplift and an eastward migration of the thrust front, which incorporated the Malargüe area into the thrust belt, uplifting and potentially eroding much of the foreland basin stratigraphy.

CONCLUSIONS

The Mesozoic to early Cenozoic history of the Malargüe area records the transition from thermally to flexurally controlled subsidence by at least 97 ± 2 Ma. This change coincides with a shift in facies from marginal marine to continental sedimentation, and an increase in clastic input into the northern Neuquén Basin. Palaeocurrents indicate a change from east- to west-derived sediment between the lower and middle Diamante Formation, and this is interpreted as a change in sediment source area from the forebulge to the fold-and-thrust belt, requiring forebulge migration between ca. 95 and 80 Ma (Fig. 11). Following the Late Cretaceous migration of the foreland basin system, it remains relatively stable during deposition of the Malargüe Group in the latest Cretaceous and early Cenozoic, however the volcanic arc appears to increase in activity between 70 and 60 Ma (Figs 10 and 11). Following the deposition of the Malargüe Group, continued eastward migration of the fold-and-thrust belt and volcanic arc in the Eocene and/or Miocene cause uplift and erosion of much of the Upper Cretaceous foreland basin stratigraphy. The timing of initiation and migration of the foreland basin system in the Malargüe area (97 ± 2 Ma) is relatively consistent to the south, both in the Central Neuquén Basin (100 ± 8 maximum depositional age; Tunik et al., 2010; Di Giulio et al., 2012) and in the Patagonian Andes (92 ± 1 Ma maximum depositional age; Fildani et al., 2003). North of the Neuquén Basin, in the central Andes, there is strong evidence for Palaeogene shortening (Horton & Decelles, 2001; Decelles & Horton, 2003; Carrapa et al., 2011, 2012; Decelles et al., 2011) and less well constrained evidence for shortening in the mid to Late Cretaceous in the Atacama basin to the west (Mpodozis et al., 2005; Arriagada et al., 2006). The initiation of shortening in the Malargüe area seems to coincide with shortening to the south, in the Patagonian Andes, and may be contemporaneous with shortening to the north requiring a plate-scale mechanism. If shortening to the north does not begin until the Palaeogene, the transition from extension to compression requires a regional-scale mechanism and a sharp boundary between older (ca. 95 Ma) ages in the south and younger (ca. 65 Ma) ages to the north.

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SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Appendix S1. Tables with point counting information. (A) Abbreviations used and equations for recalculated values shown in Table 1. (B) Raw point count data.

Appendix S2. Detrital zircon extended methods (A) and datatables (B).

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Cretaceous strata of the northern Neuquén Basin


