

# Mio–Pliocene volcanoclastic deposits in the Famatina Ranges, southern Central Andes: A case of volcanic controls on sedimentation in broken foreland basins

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## Abstract

A well-constrained record of Miocene–Pliocene explosive volcanism is preserved within the broken foreland of Western Argentina along the Famatina Ranges. This paper focuses on the volcanoclastic record known as the Río Blanco member of the El Durazno Formation. Three facies can be recognized in the study area: (1) massive tuffs; (2) volcanoclastic conglomerates and (3) pumiceous sandstones. These facies are interpreted as primary pyroclastic flow deposits (ignimbrites) and reworked volcanogenic deposits within interacting volcanic–fluvial depositional systems. Alternation between ignimbrites and volcanogenic sandstones and conglomerates suggest a recurrent pattern of sedimentation related to recurrent volcanic activity. Considering the facies mosaic and relative thicknesses of facies, short periods of syn-eruption sedimentation (volcanoclastic deposits) seem to have been separated by longer inter-eruption periods, where normal stream-flow processes were dominant. The volcanoclastic component decreases up-section, suggesting a gradual reduction in volcanic activity. The mean sedimentation rate of the Río Blanco member is higher (0.44 mm/year) than those obtained for the underlying and overlying units. This increase cannot be fully explained by foreland basement deformation and tectonic loading. Hence, we propose subsidence associated with volcanic activity as the causal mechanism. Volcanism would have triggered additional accommodation space through coeval pyroclastic deposition, modification of the stream equilibrium profile, flexural loading of volcanoes, and thermal processes. These mechanisms may have favored the preservation of volcanoclastic beds in the high-gradient foreland system of Famatina during the Mio–Pliocene. Thus, the Río Blanco member records the response of fluvial systems to large, volcanism-induced sediment loads.

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## 1. Introduction

The influence of volcanism on nonmarine sedimentation has been addressed in several studies (e.g., Kuenzi

et al., 1979; Walton, 1986; Smith, 1987a, 1991; Waresback and Turbeville, 1990; Nakayama and Yoshikawa, 1997; Smith et al., 2002; among others). However, little work (e.g., Smith, 1988; O'Halloran and Gaul, 1997) has dealt with volcanism-induced sedimentation within fold belt and “Laramide-type” broken foreland settings (cf. Jordan and Allmendinger, 1986; Jordan, 1995).

Explosive volcanic eruptions involve rapid delivery of large volumes of pyroclastic material and the instan-

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taneous modification of topography and drainage patterns (Smith, 1991). Hence, the sedimentary processes operating within an active volcanic terrain are different compared with those of non-volcanic settings and cannot be adequately explained by the existing or “background” alluvial facies models (Smith, 1987a). Whereas in non-volcanic fluvial systems aggradation and basin fill are related to basin subsidence and/or high sediment influx from adjacent uplifts (Miall, 1981; Devlin et al., 1993; Sobel et al., 2003), in volcanoclastic environments most sediment aggradation occurs during short syn-eruption periods separated by longer inter-eruption periods of incision when streams adjust to reestablish their former graded profile (Smith, 1987a, 1991). These fluctuations occur at a much higher frequency than tectonically or climatically induced modifications (Smith, 1991). Syn-eruption deposits have a high preservation potential, because they are relatively thick, laterally extensive sheets with relatively low-erodability due to their inherent primary welding. Conversely, preservation of inter-eruption alluvial successions is independent of volcanism and requires base-level adjustments within the basin (Smith, 1991).

Late Miocene–Early Pliocene explosive volcanic activity is recorded within a thick Neogene synorogenic succession (>3700 m) in the intermontane region of Famatina, southern Central Andes, western Argentina. This volcanism, more than 200 km east of the Cordillera Principal (Fig. 1), was related to flat-slab subduction of the Nazca plate inferred by the eastward broadening of arc volcanism (Kay et al., 1988; Kay and Mpodozis, 2002; Dávila et al., 2004).

This paper examines the response of an alluvial system to a large volcanism-induced sediment load recorded within the broken foreland and compares the volcanoclastic interval with those units devoid of volcanoclastic influence. We propose that the recorded episodes of explosive volcanism greatly increased instantaneous sediment accumulation rates, contributing to the preservation of the unusual stratigraphic thickness independent of tectonic and climatic factors.

## 2. Geological setting

The Famatina Ranges in western Argentina (Fig. 1) constitute the westernmost part of the Sierras Pampeanas

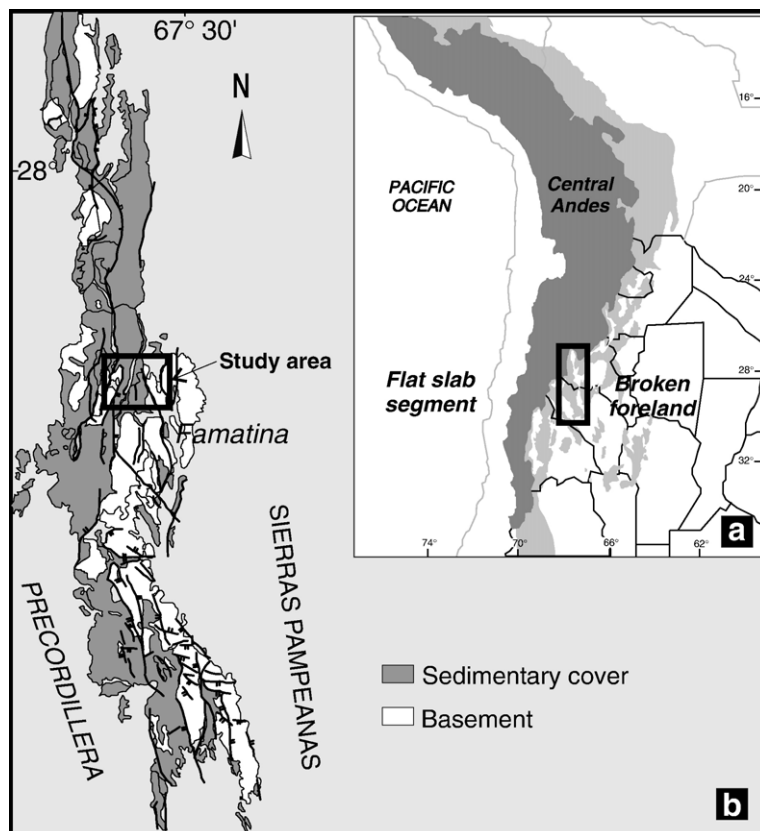


Fig. 1. (a) Location of the Famatina belt within the southern Central Andes. (b) Geological map of Famatina, depicting the study area.

nas broken foreland (sensu Jordan, 1995) between ~27° and 31° S latitude. This ~N–S mountain range was recently interpreted as a transition zone between the conventional Cordilleran thin-skinned thrust belts and the thick-skinned Andean broken foreland (Dávila, 2003), analogous to the structural overlap belt described by Kulik and Schmidt (1988) in the west of North America. Despite its cratonward position, Famatina is the highest region within the broken foreland, with peaks over 6000 masl.

The relatively simple stratigraphy of the Famatina region comprises >10-km thick, Cambrian to Cenozoic rocks (Astini, 1998; Fig. 2). Poorly exposed Cambrian metasediments intruded by Ordovician granites compose

the basement locally. These rocks are largely overlain by Ordovician volcanic, volcanoclastic and marine sedimentary rocks. An extensive Late Paleozoic–Triassic non-marine succession laps onto the different Early Paleozoic rocks. A thick section of ~4000 m Andean synorogenic strata, represented by the Early Miocene Del Crestón Formation, the Middle Miocene–Pliocene Angulos Group, and the Pleistocene Santa Florentina Formation complete the stratigraphic record in the region. We herein focus on the uppermost interval of the Angulos Group, known as El Durazno Formation (Fig. 3).

The Sierras Pampeanas is the easternmost region within the Andean belt and has been acknowledged as one of the most outstanding examples of continental

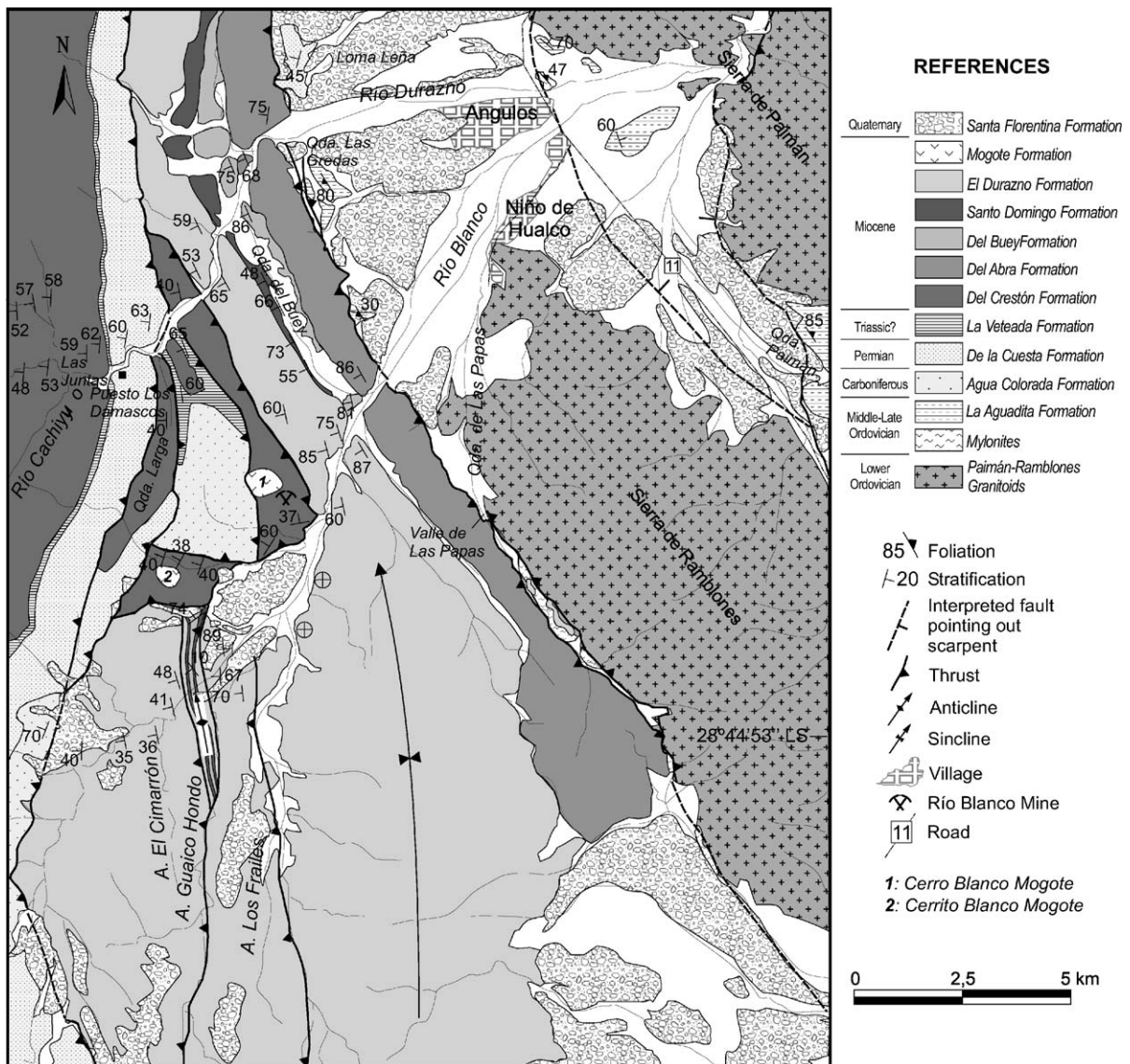


Fig. 2. Geological map of the study area in the Rio Blanco region.

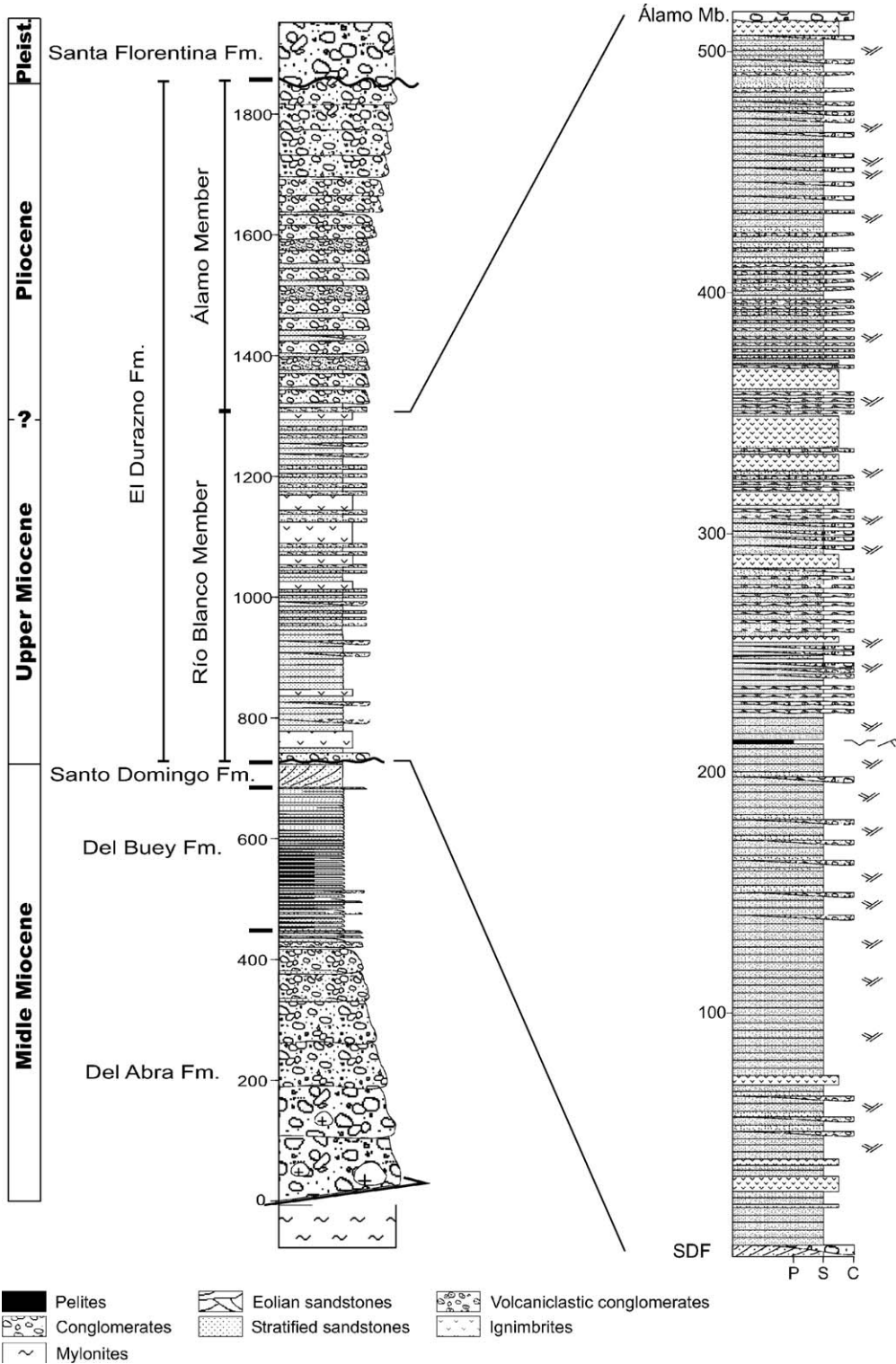


Fig. 3. Simplified stratigraphic column of the Angulos Group in the Rio Blanco section (~1970 m thick, see location in Fig. 2). The column to the right represents a more detail scale of the volcaniclastic Río Blanco Member stratigraphy.

lithosphere affected by flat subduction (Barazangi and Isacks, 1976; Cahill and Isacks, 1992; Gutscher, 2002). This tectonic regime is interpreted as the major control for the broken foreland development in western and central Argentina (Jordan and Allmendinger, 1986; Ramos et al., 2002). The modern flat slab is evidenced by the absence of Quaternary volcanism and recurrent seismic activity (Regnier et al., 1992). However, the progressive shallowing of the slab can be tracked from the Early Miocene by eastward shifting and broadening of volcanism since ~18 Ma (Kay et al., 1988, 1991; Kay and Mpodozis, 2002; Bissig et al., 2002; Dávila et al., 2004). The volcanic history roughly matches with the timing of the foreland basement thrusting, which also began in the Early Miocene (Dávila and Astini, 2003a; Dávila et al., 2004). Therefore, the modern Sierras Pampeanas landscape can be considered as a complex result of Neogene to Quaternary basement thrusting (Pilger, 1981; Jordan et al., 1983; Jordan and Allmendinger, 1986; Strecker et al., 1989; Costa and Vita Finzi, 1996; Reynolds et al., 2000; Ramos et al., 2002; Dávila et al., 2004) with coeval volcanism tracking the slab flattening, in close relationship to the southward migration of the Juan Fernandez ridge (Gutscher et al., 2000; Yañez et al., 2001; Kay and Mpodozis, 2002).

In Famatina, the El Durazno Formation records the Late Miocene–Pliocene explosive volcanism, approximately contemporaneous with those described in the easternmost Sierras Pampeanas (Pocho and El Morro volcanic fields in the Sierras de Córdoba and San Luis respectively, >200 km east; Ramos et al., 1991; Kay and Gordillo, 1994), and in the Main Cordillera (Bissig et al., 2002). These are the last manifestations of Andean volcanism in this segment of the southern Central Andes, recording the maximum broadening of the magmatic activity (Kay and Mpodozis, 2002) and the beginning of the flat slab regime according to James and Sacks (1999).

### 3. The El Durazno Formation

The El Durazno Formation (Turner, 1962; amended by Dávila, 2005) constitutes a non-marine, volcanoclastic–clastic assemblage deposited between the Late Miocene and Early Pliocene (Tabbutt, 1990). This unit represents the thickest (>1100 m) succession within the Angulos Group (Fig. 3). Although some volcanoclastic features have already been documented (Turner, 1971; De Alba, 1979; Tabbutt et al., 1989; Tabbutt, 1990; Dávila, 2003, 2005), their description as well as the origin and significance of the succession remain unclear. The El Durazno Formation laps onto a re-

gional unconformity developed above the aeolian deposits of the Santo Domingo Formation (Dávila and Astini, 2003b; Dávila, 2005). In the easternmost part of the region it locally covers non-conformably the basement granites (Dávila and Astini, 2003c). The unit is unconformably covered by the coarse-conglomerates of the Santa Florentina Formation (Fig. 3). This angular discordance can locally reach ~60°. The El Durazno Formation can be subdivided into two members (Fig. 3): (1) the Río Blanco and (2) the Álamo Members (Dávila, 2005), mainly differentiated by their distinctive colors, which in turn reflect the amount of volcanoclastic content.

The Río Blanco Member (535 m in the study area) consists of an alternating succession of off-white to pink sandstones and conglomerates with various interbedded massive tuffs each up to 10 m thick. Within the unit, two volcanoclastic-rich intervals are recognized, defining two major eruptive cycles (cf. Fisher and Schminke, 1984). According to the geochronological data, these volcanic events occurred between  $5.2 \pm 0.85$  (Dávila, 2003; Dávila and Astini, in revision) and  $4.0 \pm 0.8$  Ma (Tabbutt, 1990; Tabbutt et al., 1989). Towards the top, the El Durazno Formation shows a subtle coarsening, dominated by clast-supported coarse conglomerates, locally interlayered with sandy beds that represent the bulk of the overlying Álamo Member (633 m). In contrast with the underlying Río Blanco member, this upper member lacks significant volcanoclastic horizons.

The lithologies together with stratigraphic stacking patterns suggest that the El Durazno Formation was deposited in a broken foreland basin (cf. Jordan, 1995), within a volcanic–alluvial interactive system (Martina et al., 2003; Dávila, 2003, 2005), recording pulses with recurrent explosive volcanic activity, followed by periods devoid of volcanism and leading to accumulation of non-volcanoclastic alluvial successions.

### 4. Primary and reworked pyroclastic units

Three volcanoclastic facies characterize the Río Blanco Member in central Famatina: (1) massive tuffs, (2) volcanoclastic conglomerates and (3) pumiceous sandstones; all of them associated with non-volcanoclastic beds.

#### 4.1. Massive tuffs

##### 4.1.1. Description

Consist of medium to fine-grained off-white, slightly welded, poorly sorted tuffs, arranged into tabular

bodies of 1 to 10 m thick and tens of kilometers in lateral extent. These facies commonly show positive relief in the field (Fig. 4a) and contain sparse, pumiceous and lithic fragments of lapilli and block size. Contacts are sharp, with non-erosional, irregular to wavy bases and strongly irregular tops. Erosional features (e.g., cut and fill structures) frequently truncate the top of the massive tuffs with incisions of up to 1 m deep, vertical walls (Fig. 4b). Along strike these paleochannels show a sub-parallel alignment oriented toward the east. At an outcrop scale the tuffaceous bodies are homogenous and largely structureless. However, closer textural analysis reveals the existence of normal grading of lithic fragments and reverse grading of pumice blocks. Locally, in the upper part of some individual tuffs pseudocolumnar structures (0.04 m width and 0.2 m length) crosscutting internal features and perpendicular to the upper surface of the beds are present (Fig. 4c). These have a distinct brecciated clast-supported fabric.

The coarser particles within this facies are represented by lithic and pumice fragments in approximately equal proportion. Lithics are subangular to angular, reach 0.04 m maximum grain size, and are primarily composed of porphyritic dacites (~80%), granites and metapelites. The pumiceous fragments, in contrast, are rounded to subrounded and show clast sizes between 0.01 and 0.15 m. Internally, the pumice clasts present rounded to ellipsoidal vesicles perpendicular to the flow foliation, and more elongate tube-like vesicles parallel to the flow direction. Calcite and dolomite partially fill the vesicles. The pumiceous fragments are rich in crystals (30–35%).

The matrix of the pumiceous tuffs is composed of crystals, crystal fragments and cusped and platy glass shards, locally recrystallised (Fig. 4d). The crystals consist of euhedral to subhedral, striped-twinning plagioclases (25–30% and 0.0625 mm to 2 mm size); straight extinction quartz (10–15%); large acicular crystals of green hornblende (3 mm maximum size); tabular, greenish-brownish biotites and subordinate opaque

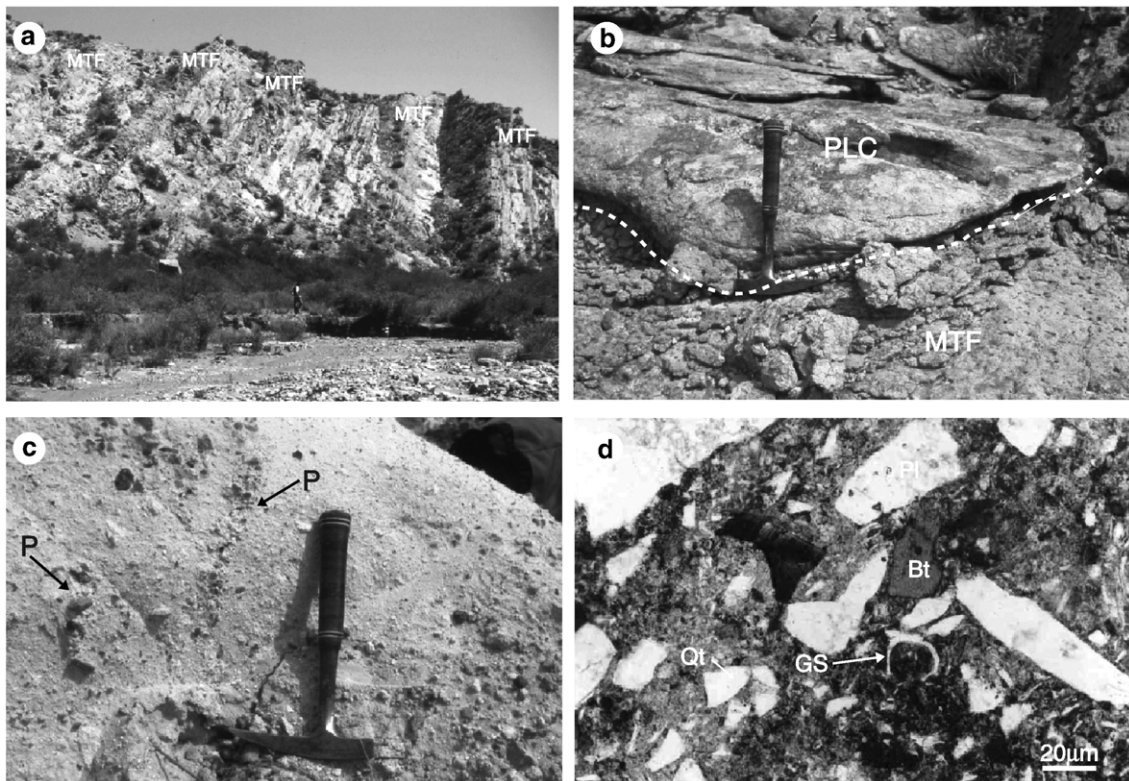


Fig. 4. Massive tuff facies (MTF) of the Rio Blanco Member (lower Durazno Formation). (a) Northward view of the Rio Blanco volcanic Member. Note the recurrent development of MTF interbedding high-gradient fluvial packages, and locally interfluvial layers, supporting an environment where interacted volcanic and fluvial processes. (b) Polymictic conglomerate incisive channels (PLC) dissecting MTF during the inter-eruptive stages, suggesting early welding of the pyroclastic beds. (c) Gas segregation pipe structures (P) vertically cutting volcaniclastic layers. Notice the contrasting fabric between the columnar pipes filled with coarser and more clast-supported volcanic fragments, and the finer cineritic matrix of the volcaniclastic flow. (d) Photomicrograph showing the matrix composition of the ignimbrites in thin section (plane polarized light). Note the outstanding development of cusped glass shards (GS) and angular phenocrystals of Plagioclase (Pl), Biotites (Bt) and Quarts (Qt).

minerals. The plagioclase composition varies from An<sub>47–63</sub>. K-feldspar, epidote, muscovite, zircon, apatite and titanite constitute minor components. All crystals are very angular and lack alteration. Microfolded biotites are occasionally observed.

#### 4.1.2. Interpretation

The physical features together with composition allow interpretation of this facies as primary pyroclastic flow beds. The massive and homogenous nature of these deposits and non-erosional bases suggest emplacement from laminar flows (cf. Sparks, 1976; Fisher and Schminke, 1984). The poor sorting can be attributed to a high particle concentration in the flow (Sparks, 1976), suggesting inertial processes. The accumulation of lithic fragments at the base of beds and of pumice towards the top indicates buoyancy during transport by pyroclastic flow (Smith, 1960; Sparks, 1976; Wilson, 1980). The degree of lithification of these beds could be explained by high-temperature emplacement. This is consistent with the ductile deformation

shown locally by the microfolded biotites. The columnar structures that perpendicularly cut the layers are interpreted as gas segregation pipes. The incisions at the top of individual massive beds indicate early welding. The subparallel pattern in this drainage network, in turn, indicates relatively high surface gradients  $\geq 2^\circ$  (cf., Phillips and Schumm, 1987). Hence, it is suggested that the above-mentioned features are consistent with slightly welded ignimbrites, deposited from dense pyroclastic flows (low gas/solid ratio). The size and degree of preservation of the pumice clasts suggest distances for the pyroclastic accumulation of the order of tens of kilometers from the eruptive volcanic sources.

#### 4.2. Volcaniclastic conglomerates

##### 4.2.1. Description

Matrix to clast-supported, largely disorganized pumice conglomerates arranged in tabular and more rare lenticular bodies (Fig. 5a). Beds range between 0.05

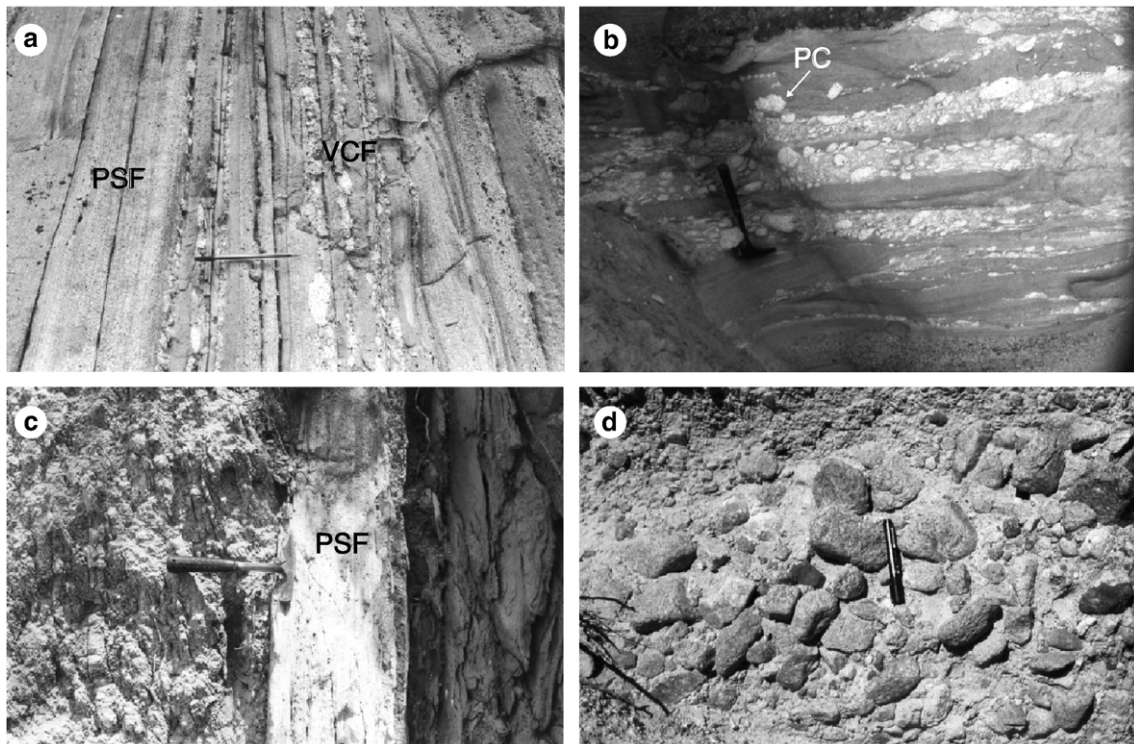


Fig. 5. Facies of the Durazno Formation. (a) Pumiceous sandstones (PSF) interlayering volcaniclastic conglomerates (VCF) in the Río Blanco Member. Note the marked tabular geometry of the beds. Base to the right and pen for scale. (b) Detail of the VCF. Note the size and incipient imbrication of the pumice clasts (P). The pick points to the base. (c) Thin bedding in the PSF. Note the plane and low angle cross stratification. The pick points to the base. (d) Coarse alluvial conglomerates at the upper Álamo Member in the upper most section of the Durazno Formation. Note the clast size over 20 cm (and locally, out of the picture, ~1 m), and polymictic composition, with predominance of granite fragments, compositionally similar to those exposed in the Sierra de Famatina. Marker is ~10 cm long.

and 0.5 m thick and commonly have sharp erosional contacts. Whereas bases are planar to wavy, tops are seldom markedly irregular due to subsequent erosion. Amalgamation of separate beds was recognized based on subtle fabric and grain size variations. Crude stratification and poor sorting is most common, although locally beds exhibit remarkable normal grading and within the upper part of the beds, low-angle cross- and parallel stratification. Within the sandy caps, a few beds show wavy lamination where boundaries with subsequent facies are most erosional. Conglomerates are predominantly composed of subangular to subrounded pumice blocks (~90%), suspended in a fine- to coarse-grained, cineritic–epiclastic, sandy matrix. Individual fragments have a maximum clast size of ~0.2 m (Fig. 5b). High concentration of pumiceous blocks toward the top of the beds allows formation of confined, clast-supported fabrics. Very few imbricate clasts were observed. However, one persistent characteristic is the abundant flattening of pumiceous pebbles and cobbles with *a*-axis parallel to the layer boundary. The lithic fragments (<10%) are remarkably smaller than the pumiceous (0.025 m maximum size), and are composed mainly of porphyritic dacites and by minor amounts of granite clasts and other volcanic and sedimentary rocks present in the local underlying stratigraphy.

#### 4.2.2. Interpretation

Sharp erosional bases and mechanical structures suggest highly turbulent flows and waning tractional deposits (Fisher and Schminke, 1994). The sandy, matrix-supported fabrics are interpreted as arising from high-particle concentration during deposition, which would have favored the trapping of large amounts of fines (Cole and Scarpato, 1993). Upward fining together with the sedimentary structures recording lower-stage flow regime toward the upper part of the beds suggest gradual waning conditions, likely associated with sheet-like flash floods. The concentration of large pumiceous clasts on top of beds is attributed to flotation due to differential buoyancy developed within hyperconcentrated, denser pulses (Postma et al., 1988). The elongate shape of pumiceous clasts in turn is attributed to compaction (early diagenesis). Similar facies deposited from hyperconcentrated flood-flow facies (cf., Smith, 1986) are an important component of volcanoclastic sequences (Smith, 1987b). The clast and matrix composition suggest that these flows scoured and reworked a substrate of pyroclastic nature, similar to that described for facies 1, but locally involved other sources farther west as shown from paleocurrent data and granite boulders presently exposed in the higher ranges

within the Famatina thrust belt (Dávila and Astini, in revision).

### 4.3. Pumiceous sandstones

#### 4.3.1. Description

Thin-medium bedded, pumice-rich, medium to coarse grained, poorly sorted, sandstones (Fig. 5b). Individual beds between 0.05 and 0.3 m thick are laterally persistent and typically have sharp boundaries with slightly erosional basal surfaces and wavy tops. Stratification is typically imparted by the alternation of coarse and fine-grained layers. Internally, beds show normal grading. Planar and low-angle (<15°) cross-bedding are the most recurrent traction structures. Ripple cross-lamination at the top of beds is also common. Texturally the pumiceous sandstones are moderately immature (grains range from angular to subrounded) and poorly sorted. Thin sections also show euhedral to subhedral broken plagioclase (An<sub>43–52</sub>) and quartz crystals, volcanic lithics (dacites and ignimbrite fragments) and relict pumice clasts. Minor amounts of devitrified cusped shards and microfolded biotites are also present. The fabric is dominantly grain-supported. Composition of sandstones, determined by standard point-counting technique, is 53% plagioclase, 20% quartz, 18% lithic fragments and 5% opaque minerals; hornblende, biotite and potassium feldspar together represent ~4%. According to their composition they plot as lithic–feldespathic sandstones (cf. Pettijohn et al., 1987). Pumice fragments are in general oversized, reaching granule to pebble size. Heavy mineral concentrations are locally important and remark the primary lamination.

#### 4.3.2. Interpretation

The presence of well-developed mechanical sedimentary structures, thin stratification and relatively well-rounded grains suggest deposition from dilute turbulent flows, where the transport was essentially by traction (Lowe, 1982). However, the internal sequence of structures and grading suggest a flashy discharge regime with decelerating flows, compatible with those described for event beds by Olsen (1989). The tabular geometry of the sandy beds suggests extended and unconfined (sheet-like) flows like those recorded in modern and ancient ephemeral streams, terminal fans or poorly evolved distal braided fluvial systems (e.g., Kessler, 1971; Tunbridge, 1981; Parkash et al., 1983; Stear, 1983; Horton and DeCelles, 2001). Sheet floods are short-lived, relatively shallow and often supercritical flows (Rahn, 1967) adequately

explaining the sheet-like bed geometry and the upper regime planar and standing wave laminations described within the beds. These flows would have remobilized a pyroclastic substrate as suggested by the pumiceous-rich composition of the sandstones, euhedral phenocrysts and relict glass shards. Sub-rounded quartz grains and volcanic and pyroclastic rock fragments indicate limited reworking. These sandstones can be classified as volcanogenic deposits (cf. Tucker, 1991).

## 5. Alluvial facies association

This association is volumetrically the most important within the upper Álamo Member, but also punctuates the volcanoclastic association that characterizes the Río Blanco Member. The alluvial facies association comprises an assemblage of epiclastic coarse conglomerates, sandstones and siltstones without significant volcanic components. A detailed sedimentological and stratigraphic analysis of these alluvial deposits has been carried out (Dávila, 2005). Thus, we herein only summarize the essential features. Conglomerates are polymictic, poorly organized, coarse to medium grained, clast-supported, and subrounded to subangular. They are arranged in lenticular to semi-continuous bodies,  $\leq 3.5$  m thick (Fig. 6), showing locally amalgamation inferred from textural inversions and contrasting fabrics. Scour and fill structures are common and erosionally truncate the other facies. Deep incisions with overhanging walls particularly affect the tops of mas-

sive tuffs and associated volcanoclastic deposits. Conglomerates are mostly interbedded with poorly sorted, graded and well laminated arkosic sandstones. The sandstones are tabular and laterally persistent (hundreds of meters). Very fine-grained sandstones–siltstones and reddish mudstones are also present.

Lenticular conglomerates are interpreted as stream-flow deposits related to braided–fluvial systems (e.g., Miall, 1981; Bridge, 2003). Poor internal organization is interpreted as derived from hyperconcentrated flows, consistent with abundant channels and large maximum clast sizes, which in turn indicate relatively high gradient. Sheet-like sandstones and siltstones suggest over-bank deposits, whereas remaining fine-grained rocks may represent deposition from waning sheet floods in more distal alluvial or ponded settings. At a larger scale, the coarsening and thickening-upward arrangement of the El Durazno Formation indicates alluvial fan progradation with more proximal environments up-section (Dávila, 2003, 2005).

## 6. Depositional setting of the Río Blanco Member

The Río Blanco Member was deposited coevally with intense volcanism as constrained by the widespread ignimbrites extending more than 50 km along the eastern flank of the Sierra de Famatina. These massive tuffs were generated from explosive eruptions. Volcaniclastic sandstones and conglomerates grade vertically from one to another suggesting subsequent fluvial reworking during inter-eruption periods. The

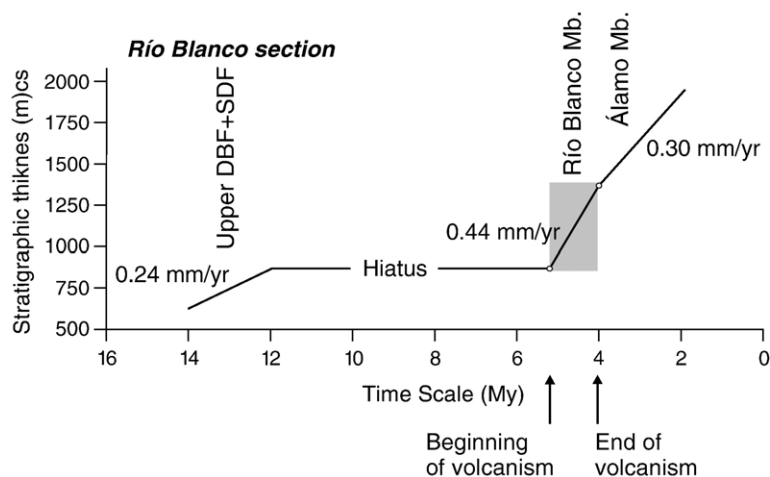


Fig. 6. Sedimentation rates from the mid (Del Buey and Santo Domingo Formations) to upper most (Durazno Formation) sections of the Angulos Group. The stratigraphic thickness-vs.-age chart depicts clearly a marked change in the accumulation rate in the volcanoclastic member (Río Blanco), if compared with the rest of the non-volcanoclastic units. This rate change would have occurred under similar climate conditions. The major Mio–Pliocene basement uplifting event, in contrast to expect from this calculus, would have occurred during deposition of the Álamo Member (Dávila, 2005; Dávila and Astini, in revision).

volcaniclastic association constitutes extensive sheets separated by several meters of non-volcanic alluvial deposits that erosionally truncate the top of volcaniclastic units. The regular alternation between pyroclastic facies and volcanogenic sandstones and conglomerates suggests episodic sedimentation punctuated by strong volcanic influence.

Most sedimentation seems to have occurred during episodes of intense volcanism when large volumes of pyroclastic material were introduced into the alluvial system, triggering changes in drainage patterns and facies. When volcanism ceased or diminished, erosion and reworking processes became dominant (e.g., interstratal erosional surfaces, diastems). This alternation is consistent with the volcaniclastic facies model proposed by Smith (1991), who suggested that between major explosive volcanic events, rivers tend to recover their graded profile, incising channels through the existing substrate of pyroclastic composition. The following syn-eruptive period of sedimentation would again disrupt equilibrium and contribute to rapid aggradation. This alternation of aggradation and degradation periods is common in alluvial–volcanic settings (Smith, 1987b). The development of channel facies suggests more stable fluvial conditions probably related to a diminished volcanic activity (O'Halloran and Gaul, 1997). The recurrence of these lenticular conglomerates associated with high mobility braided stream systems (cf. Bridge, 1993), with wide marginal zones affected by episodic flash floods, can be compared to mixed fluvial–volcanic interaction systems (e.g., Smith, 1987b, 1988; Haughton, 1993). This model of periodic aggradation–degradation controlled by rapid volcanism-induced sedimentation is similar to those described in the literature (e.g., Smith, 1987b, 1988; Waresback and Turbeville, 1990; Nakayama and Yoshikawa, 1997) and analogous to different modern examples in the Central Andes of Argentina and Chile (e.g., Clavero et al., 2004).

Although air fall tuffs have been mentioned in nearby regions (e.g., Tabbutt, 1990), only pyroclastic flow deposits (ignimbrites) are documented in the central part of Famatina. These flows would likely have been generated from gravitational collapse of Plinian eruptive columns (cf. Sparks and Wilson, 1976) associated with stratovolcanoes. Volcaniclastic deposits become less abundant upward in the sequence reflecting diminishing volcanic activity.

Ignimbrites tend to fill valleys and depressions, leveling landscape and hence generating lenticular profiles. However, in the study region the Río Blanco ignimbrites show tabular geometry and nearly flat

bases suggesting deposition on a substrate without intrabasin highs. Besides, this landscape is consistent with the widespread sheet floods recorded in Famatina. Considering the deep incisions into the ignimbrites we suggest that the paleoslope was important, but  $<5.5^\circ$  as shown by experimental work on ignimbrites (Wooler et al., 2004). In this context, braided systems within the alluvial setting have potential to recurrently rework and remobilize previously emplaced primary pyroclastic deposits.

The volcaniclastic successions recorded at central Famatina would have been deposited in a proximal–medial setting according to the criteria developed by Smith (1988, pp. 960–968). This observation is supported by the following features: a) the absence of extremely proximal facies (e.g., block and ash flow deposits); b) the scarcity of debris-flow deposits, and c) the maximum grain size of pumiceous fragments in the ignimbrites of  $\sim 0.15$  m. According to geochronological data ( $\sim 5.0$  Ma; Losada-Calderón et al., 1994), these pyroclastic rocks appear to be linked with the numerous dacitic domes (e.g., Tabbutt, 1990; Losada-Calderón et al., 1994; Ramos, 1999), responsible for the gold mineralization in the central region of Famatina (Kay and Mpodozis, 2001). Such dacitic domes are presently located within the study area and are considered the roots of the Mio–Pliocene volcanism. Strong compositional similarities and the fact that shortening during the Late Miocene–Present was not significant in Famatina (Dávila, 2003), suggests that the volcanic centers were very close to the depocenter and functioned as sources for the ignimbrites.

## 7. Depositional rates, basin subsidence and preservation of volcaniclastic succession in the foreland setting

The highly tuffaceous composition of the Río Blanco Member and the presence of at least 11 major ignimbrites demonstrate coeval development of the alluvial systems with explosive volcanic events. This well-preserved interval of intense volcanism (of 535m) within the thick ( $>3100$  m) non-volcanic llastic synorogenic succession of Famatina allows reconstruction of the history of an important volcanic episode within the broken foreland, related to the flat subduction in the south Central Andes (see Kay and Mpodozis, 2002).

Compiling the existing information of Famatina and nearby regions and the data herein presented, some important issues regarding sedimentation rates in volcanic–alluvial settings, the generation of accommodation space coevally with pyroclastic deposition, and the

preservation within high-gradient foreland systems can be made.

According to the available geochronological data the intense volcanic activity persisted for 1.2 Ma. Dates of the bottom ( $^{40}\text{Ar}/^{39}\text{Ar}$ ) and top (zircon fission-track) of the succession are, respectively,  $5.2 \pm 0.85$  (Dávila, 2003; Dávila and Astini, in revision) and  $4.0 \pm 0.8$  Ma (Tabbutt, 1990). The maximum exposed thickness of this member is  $\sim 535$  m along the Río Blanco section (see Fig. 3, column). Although the weighted mean ages are indistinguishable within error, these data allow calculate a mean accumulation rate of 0.44 mm/year (0.23–16.3 mm/year with analytical error using the propagation of error formula). It is important to remark that the obtained value constitutes a minimum sedimentary rate due to the abundant truncation features that include complete erosional removal of pyroclastic units as observed during fieldwork. This is a common process in volcanic environments that interact with fluvial systems, which systematically reduce the primary thickness of the volcanic successions. The calculated sedimentation rate also did not take into account compaction rates, which would increase the values (Damanti and Jordan, 1989). The high sedimentation rate for the Río Blanco Member contrasts with those inferred for the underlying and overlying Miocene units,  $\sim 0.25$  mm/year and  $\sim 0.30$  mm/year, respectively (Fig. 6). These rates were calculated using mammal ages, palynology and stratigraphic correlations (Lencinas, 1994; Dávila and Astini, 2003c; Dávila, 2005; Barreda et al., in press). The contrast in sedimentation rates can also be inferred from sedimentological features. The underlying Del Buey Formation preserves numerous paleosol horizons (Dávila, 2005) and lacustrine sections, which suggest lowered sedimentation rates for the considered interval. From the base of the El Durazno Formation upward, the sedimentation is represented by high-gradient alluvial environments. In this context, the volcanoclastic input and its preservation was exceptional and only recorded in the Río Blanco Member. These exceptional sources were responsible for increasing the local sedimentation rate.

Considering the complete stratigraphy of the El Durazno Formation, Tabbutt (1990) linked the sudden facies changes (increase in clast sizes, thicknesses, etc.) in the El Durazno Formation with rapid uplift and basement-involved thrusting in Famatina. According to Tabbutt's work, basement unroofing would have changed the equilibrium in the fluvial systems and triggered development of more proximal alluvial deposits. However, studies using apatite fission-track data (Jordan et al., 1989;

Coughlin et al., 1998; Dávila et al., 2005b) do not favor such interpretations. In Famatina, most data show Paleozoic to Mesozoic thermochronologic ages, like in other ranges of the Pampean region, indicating that the Andean region experienced slow exhumation rates. This might be associated with the uplift history. Hence, an increasing sedimentary supply cannot be fully explained by foreland basement deformation. Besides, if uplift of the Famatina Range had generated the increase in the sedimentation rate, we would expect that this occurred in the coarser and more granite-rich conglomerate of the Álamo Member (Dávila and Astini, in revision), where there is compositional evidence suggesting basement unroofing. Considering there is no independent evidence in support of a climate change within the El Durazno Formation, it is possible that volcanism-induced sedimentation produced the contrasting rates. Thus, volcanism would have been an important source, supplying an extra load to the basin. Other cases in west Argentina serve as examples of high sediment accumulation rates associated with volcanism. In the fold-thrust belt of the Precordillera region, to the west of Famatina, a thick volcanoclastic–alluvial sequence of over 500 m records sedimentation rates  $> 1$  mm/year (e.g., Johnson, 1985; Beer et al., 1987).

Strata may be accumulated either by filling tectonically subsiding areas or by ponding and raising the base level in tectonically stable area between highlands (Jordan and Alonso, 1987). Within compression settings most basin subsidence is caused by the tectonic loading. Nevertheless, the volcanoclastic Río Blanco Member, as well as the rest Plio–Pleistocene units of Famatina, record thicknesses not accounted for by simple flexural analysis. In broken foreland regions, like Famatina and the Sierras Pampeanas of western and central Argentina; the relatively high elastic thickness, high-angle reverse faults bounding the basement blocks, and the small calculated shortening (between 25% and 2%; Jordan and Allmendinger, 1986; Costa et al., 1999, Dávila, 2003; Dávila and Astini, in revision) suggest that additional mechanisms are required to explain the accommodation space for deposition and increased preservation potential (see discussion in Dávila et al., 2005a, Dávila and Astini, in revision). We suggest that “volcanism-induced aggradation” (cf. Smith, 1987b, Smith et al., 2002) operated during deposition of the Mio–Pliocene Río Blanco Member, producing both a nearly instantaneous increase in sedimentation rate and larger preservation potential. Although we do not underrate the influences of tectonic and dynamic loading in the Mio–Pliocene basin developed in the Andean foreland region, it is necessary to

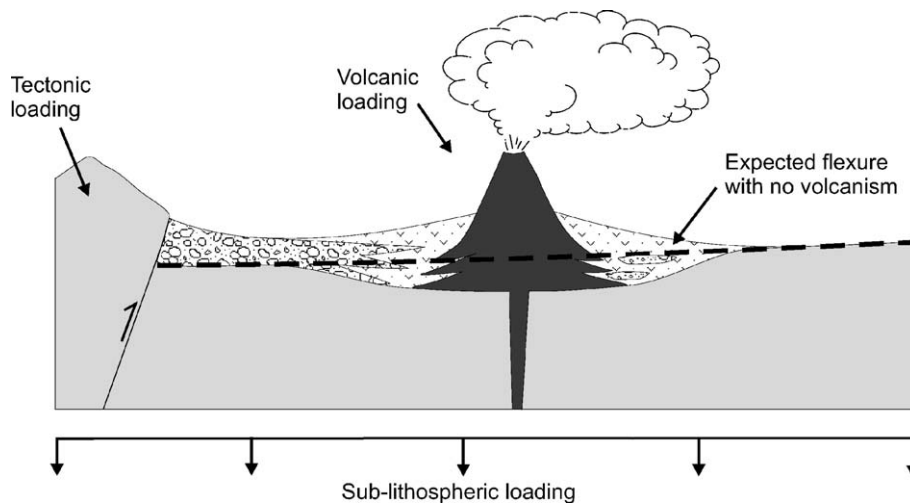


Fig. 7. Conceptual diagram showing subsidence mechanisms during deposition of Mio–Pliocene volcanoclastic succession of the Famatina belt. Note how volcanism adds extra loading to the tectonic subsidence regime, and also supplies additional material to the basin. The two effects can explain the recorded thickness, during a period with little shortening, and high sedimentation rates shown in Fig. 6.

include other factors (Fig. 7). Volcanic activity may have acted independently, overlapping the effects of basement thrust stacking, structural barrier controls (e.g., Jordan and Alonso, 1987; Sobel et al., 2003; Astini et al., 2004) and sublithosphere downwarping (e.g., Mitrovica et al., 1989; Liu and Nummedal, 2004), as recently suggested for the Pampean region by Dávila et al. (2005a) and Dávila and Astini (in revision).

During construction of the volcano (Fig. 7), the sediments were forced to aggrade in response to increase in the headwater elevation and sediment supply (modifying the stream longitudinal profile) (Smith et al., 2002). This process would also have favored the generation of extra accommodation space enhancing preservation and adding extra load to the basin. The volcanic load together with the volcanoclastic apron would have led to local flexural isostatic subsidence (Bahlburg and Furlong, 1996; Watts, 2001; Smith et al., 2002). By increasing heat flow, the volcanic activity may locally have reduced the effective elastic thickness of the crust, consequently allowing the addition of extra accommodation space. Largely contemporaneous lithosphere relaxation through thermal controls have been recently argued by Ramos et al. (2002) to explain the basement thrusting in western and central Argentina during the Late Miocene–Pliocene. A good example of subsidence surrounding volcanic edifices is the case of the Pocho volcanic field in the easternmost Sierras Pampeanas (Kay and Gordillo, 1994). This volcanic field, active between  $\sim 7.9$  and 4.5 Ma, developed on top of a near-flat, high-elevated hanging-wall base-

ment thrust sheet. The volcanoes lie near to the fault scarp and are surrounded by a striking volcanoclastic ring  $\sim 150$  m thick (Gordillo and Lencinas, 1979). Despite its supposedly low preservation potential atop an uplifted basement block, these deposits have been preserved, likely aided by former volcanism-induced subsidence.

As proposed by Smith (1987a, 1991) for the Cascade Range region in northwest North America, we consider that volcanism exerted a major control on the depositional systems and basin accommodation in the Andean foreland setting. This control was dominated by a combination of thermal processes (heating–cooling), flexural loading by the volcanic edifices and additional sedimentary loading, and by the local modification exerted on the stream equilibrium profile. Together these processes can produce an effect comparable to tectonic or climate forcing.

## 8. Conclusions

The Mio–Pliocene Río Blanco Member in Famatina, western Argentina, represents an exceptional example of where alluvial systems interacted with explosive volcanism within a broken foreland setting. The volcanic eruptions resulted in the rapid generation of a large volume of volcanoclastic sediments (aggradation stages) increasing the sedimentation rate. Pyroclastic flow deposits were intensively reworked by fluvial networks during inter-eruptive periods (degradation stages) when volcanism ceased. These alternating episodes of aggradation–degradation

exerted a primary control on both deposition and preservation. Thickness of successions, resultant geometry and good preservation potential of these volcanoclastic deposits would appear to indicate that explosive volcanic activity strongly influenced the basin dynamics, perhaps as much as tectonic controls. Although, we do not intend to disregard the tectonic and climatic influence on basin development, our objective is to emphasize the importance of volcanism as an independent control, overlapping and reinforcing other accommodation mechanisms.

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