Intermontane Late Paleogene–Neogene Basins of the Andes of Ecuador and Peru: Sedimentologic and Tectonic Characteristics

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Abstract

An important characteristic of Neogene basin evolution in the Andean Cordillera was the formation of intermontane basins. These basins were initiated in the late Oligocene with reactivation of Andean tectonism and were abandoned in the latest Miocene (about 7 Ma). Their sedimentary fill and structures record the Neogene tectonic history. The sedimentary fill of these basins comprises two megasequences. The first consists of fluvial and overlying lacustrine deposits attributed to basin opening. The second is composed essentially of proximal fluvial sedimentary units and reflects the closure of the basins in the latest Miocene.

Structural analysis of the Neogene basins shows that their evolution was controlled by the regional tectonic stress. Synsedimentary folding and fracturing show that the direction of stress experienced a clockwise rotation in the Neogene, thus explaining variations in the behavior of the faults bordering the basins as well as the different stages of their evolution.

Resumen

Una característica importante de la evolución de las cuencas neógenas de la cordillera de los Andes ha sido la formación de las cuencas intramontañosas, cuya creación se inició en el Oligoceno superior con la reactivación de la tectónica andina, y cuyo relleno finalizó en el Mioceno terminal (aproximadamente 7 Ma). Su relleno sedimentario y sus estructuras registran la historia tectónica del Neógeno. El relleno sedimentario de tales cuencas comprende dos megasecuencias. La primera consiste en depósitos fluviales y lacustres sobreyacentes que corresponden a la apertura de las cuencas. La segunda se compone básicamente de sedimentos fluviales proximales y refleja el cierre de las cuencas en el Mioceno terminal.

Según los resultados del análisis estructural de la cuencas neógenas, su evolución ha sido controlada por el esfuerzo tectónico regional. Plegamientos y fracturas sinsedimentarios muestran que la dirección del esfuerzo ha experimentado una rotación horaria durante el Neógeno, explicando así las variaciones en el comportamiento de las fallas del borde de las cuencas y en las diferentes etapas de su evolución.

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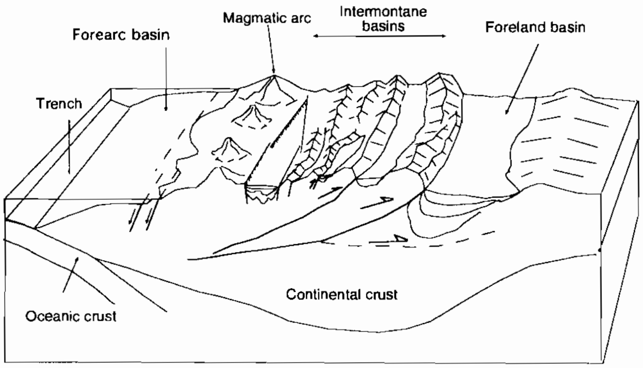


Figure 1—Location of basins in an Andean-type orogenic belt.

INTRODUCTION

The phenomena that cause folding and uplift of orogenic belts are also responsible for the genesis of foreland basins and intermontane basins (Figure 1). The sedimentary fill of these basins and their bounding structures are controlled by the tectonic forces that build the orogenic belt. Sedimentologic, stratigraphic, and structural analysis of the synorogenic basins allows a reconstruction of their development and their regional relationships. This type of study is essential to understanding the geodynamics of mountain belts.

We are particularly interested in the Andean intermontane basins of Ecuador and Peru because they are contemporaneous with Neogene tectonism and because they record structural reactivation. These processes have significance for the basins described by Mégard et al. (1984), Bonnot et al. (1988), Noblet et al. (1988), Bellier et al. (1989), Marocco et al. (1990, 1993), and Baudino et al. (1991). There are three types of intermontane basins: those linked to strike-slip faulting, those controlled by reverse faults, and those related to normal faults. Basins such as the Andean intermontane basins commonly develop through each of these types. The term *intermontane* is occasionally misused to define some Andean basins. For example, the Moquegua basin of the southern Peruvian coast has been defined as intermontane (Marocco et al., 1982; Marocco, 1984), although it is actually a Neogene continental forearc basin. Some small basins in the Eastern Cordillera of Perú, which are interpreted as Neogene intermontane basins because of their setting, are actually the early stages of the sub-Andean foreland basin. Examples include the Bagua basin of northeastern Perú and the Andamarca basin of central Perú. Mathalone and Montoya (1995) illustrate some examples.

Neogene intermontane basins are located in the inter-Andean region. This is a region of variable width separating the Western and Eastern Cordillera of Ecuador and Perú. It is known as the Inter-Andean Valley in Ecuador, as the Highlands (or High Plateau) in central Perú, and as the Altiplano in southern Perú and Bolivia (Figure 2).

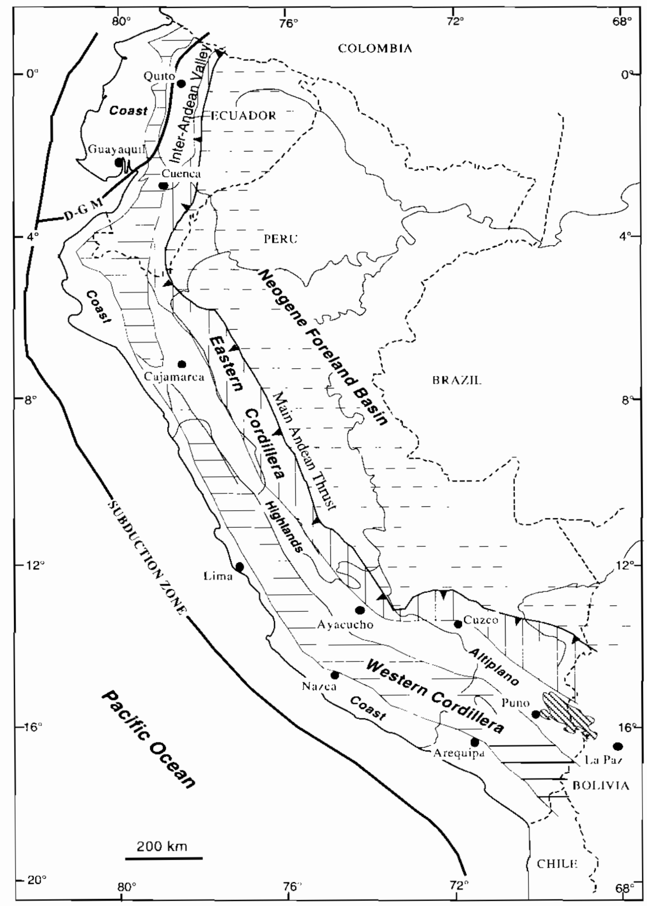


Figure 2—Morphostructural map of Ecuador and Peru. D-G M, Dolores-Guayaquil megashear, the tectonic border between the oceanic accreted coastal terranes (on the west) and the South American continental lithosphere.

DATING THE BASIN FILL

The chronostratigraphy of the succession filling the intermontane Neogene basins is not well known. Sporadic fossil occurrences and radiometric ages date some specific strata. However, radiometric ages of the volcanic rocks underlying and overlying the sedimentary fill confirm a Neogene age.

Age of the Volcanic Rocks

In Ecuador and Peru, the formation of the Neogene basins began in about the late Oligocene when arc magmatism recorded the effects of Andean tectonic reactivation. Most recent work on Andean deformation, chronology, and related sedimentation (Noblet et al., 1988; Sébrier et al., 1988; Lavenu and Noblet, 1989; Baby et al., 1990; Lavenu et al., 1990; Marocco et al., 1990; Marocco, 1991; Sempere, 1991; Sempere et al., 1991) agrees that, in the region presently occupied by Neogene intermontane basins, Andean tectonism was reactivated during the late Oligocene. This is the Aymara phase of Sébrier et al. (1988) and was apparently the start of a long tectonic phase (see Sempere, 1991).

In Ecuador, the Saraguro volcanics immediately preceding the formation of intermontane basins (Baldock, 1982) have an age bracketed between 35.3 and 26.8 Ma (Lavenu et al., 1992). The oldest dated rocks within the sedimentary succession yield an age of 22 Ma (Lavenu et al., 1992); they form the lower part of the Biblián Formation in the Cuenca basin (Noblet et al., 1988). The presence of volcanic units, dated at 19.5–14.2 Ma by Kennerley (1980), is evidence that volcanic activity was persistent during basin subsidence (Marocco et al., 1993).

In Peru, the youngest volcanic series underlying the intermontane basins belongs to the Calipuy Formation in the north (Cobbing et al., 1981), to the Huanta, Castrovirreyna, and Ayacucho formations in central Peru (Mégard, 1978), and to the Tacaza Formation in the south (Newell, 1949). The Calipuy Formation has been dated at 54–14 Ma (Cobbing et al., 1981). In central Peru, Noble (1973), Noble et al. (1974), and Dalmayrac et al. (1980) showed that the volcanic rocks enclosing the intermontane basins had an age between 41 and 6 Ma. Finally, the age of the Tacaza Formation of southern Peru ranges from 27.2 to 8.9 Ma (Sébrier et al., 1988). Therefore, in Peru as well as in Ecuador, the enclosing volcanic suites were emplaced before and during the evolution of the basins.

The sedimentary fills of the basins are generally overlain by volcanic materials. This recent volcanic cover, known as the Barroso Group in southern Peru (Mendivil, 1965), has an age of 7.2–0.17 Ma (Sébrier et al., 1988). Where present in Ecuador (e.g., Cuenca basin) (Noblet et al., 1988), the volcanic cover has a younger age, from 3.59 Ma to present (Barberi et al., 1988; Lavenu et al., 1992).

Age of Basin Sedimentary Rocks

Available dates for the sedimentary fill of the intermontane basins are both paleontologic and radiometric because significant volcanic activity occurred in the vicinity of the basins while lacustrine and fluvial sediments were being deposited. This is reflected in the basins by the occurrence of lava or pyroclastic flows and by the deposition of river-transported volcaniclastic materials.

Lacustrine facies present in almost all the basins have yielded most of the reported animal and plant fossils, which indicate an early–late Miocene age. In the northern Peruvian basins of Namora, San Marcos, and Cajabamba, Bellier et al. (1989) have described diatomite associations of early–middle Miocene age in the Cajabamba Formation and of late Miocene age in the Namora Formation. In the latter beds, a volcanic tuff dated radiometrically at 7.2 ± 0.6 Ma confirms the Tortonian age of these fossil associations.

Lacustrine fauna and flora have been studied in more detail in Ecuador. In the Cuenca basin, Marshall and Bowles (1932) and Liddle and Palmer (1941) describe ostracods (*Cyprideis* aff. *Howei*), crabs (*Necronectes proavitus*), gastropods (*Limnopomus* [*ampullarius*] cf.

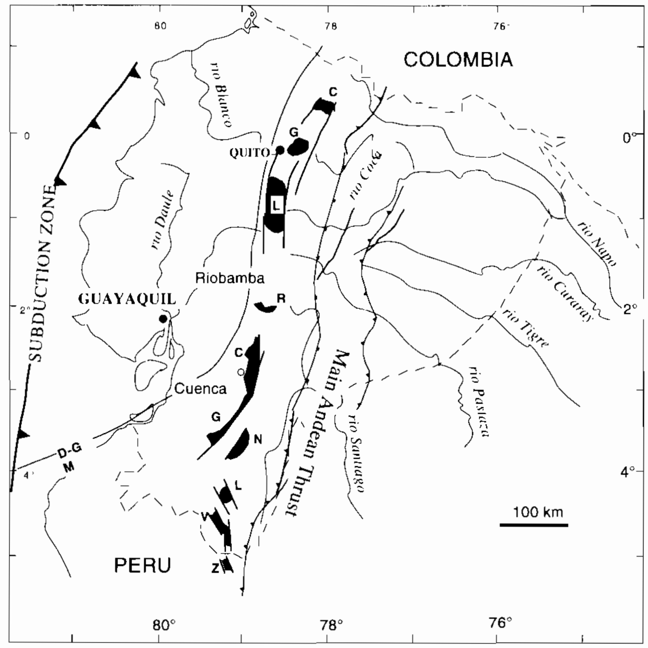


Figure 3—Location of the Ecuadoran Neogene intermontane basins. D-G M., Dolores-Guayaquil megashear; Ch, Chota basin; G, Guayabamba basin; L, Latacunga; R, Riobamba basin; C, Cuenca basin; G, Girón–Santa Isabel basin; N, Nabón basin; L, Loja basin; V, Vilcabamba basin; Z, Zumba basin.

manco, *Poteria* [*pseudoaperastoma*] *bibliana*), and bivalves (*Ecuadorea bibliana*) in the lacustrine intervals of the Loyola Formation (Noblet et al., 1988). These indicate a broadly Miocene age (Bristow and Guevara, 1974). Notoungulae (Toxodontidae) have been discovered in the overlying Mangán Formation (Repetto, 1977). The ostracod *Cyprideis stephensoni* and plant remains occur in the coal at the top of the lacustrine intervals in the Loja and Vilcabamba-Malacatos basins (Kennerley and Almeida, 1975; Marocco et al., 1993). These plants, studied by Berry (1929), include *Camphoromea speciosa*, *Cassia longiflora*, *Heronymia lehmanni*, and *Tapirina lanceolata* and belong to the late Miocene.

Radiometric dates complement the stratigraphic information. In some cases, they support the paleontologic ages, such as in the Cuenca basin (Figure 3), where volcanic intercalations show ages of 22–5.2 Ma (early–late Miocene) (Barberi et al., 1988; Noblet et al., 1988; Lavenu et al., 1992), and in the Namora basin of Peru (Figure 4). Elsewhere radiochronology is the only method for dating the sedimentary column. This is the case for the Peruvian Rumichaca basin (see Figure 8), where the base of the succession has been dated at 22 Ma (McKee and Noble, 1982; Mégard et al., 1983) and the Ayacucho basin (see Figure 7), where volcanic intercalations in the volcaniclastic succession yield ages of 18.3–6 Ma (Mégard et al., 1984). In southern Ecuador, the fluviolacustrine sedimentary rocks of the Nabón basin (Figure 3) have been dated at 22–7.9 Ma (Winkler et al., 1993).

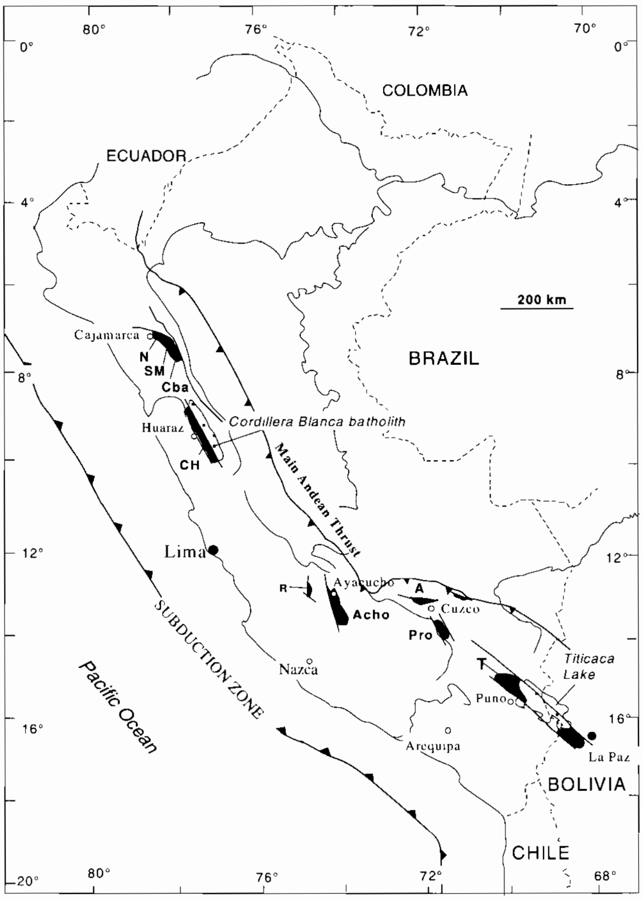


Figure 4—Location of the Peruvian Neogene intermontane basins. N, Namora basin; SM San Marcos basin; Cba, Cajabamba basin; CH, Callejón de Huaylas basin; R, Rumichaca basin; Acho, Ayacucho basin; A, Anta basin; Cc, Ccatca basin; Pro, Paruro basin; T, Tinajani basin.

LITHOSTRATIGRAPHY

The basins are filled with fluvial and lacustrine continental sedimentary rocks and significant amounts of intercalated volcanic material. Because the sedimentary environments of the Ecuadorian intermontane basins have been studied in most detail, we have taken the following examples from the works of Noblet et al. (1988), Noblet and Marocco (1989), Barragán (1992), and Marocco et al. (1993).

Depositional Environments

Only the fluvial and lacustrine sedimentary environments are represented in these intermontane basins (Figures 5, 6, 7, 8).

Fluvial Sedimentary Rocks

The fluvial sedimentary units are typically characterized by fining-upward sequences 0.5–7 m thick (Figure 9A). The relatively small area of Andean intermontane basins (rarely exceeding 1000 km^2) and the proximity of provenance is reflected in proximal fluvial systems with sequences of the type shown in Figure 9A. The sequence begins with an erosive base and is overlain by a lower conglomeratic or coarse pebbly sandstone bed with common large-scale trough cross stratification and pebble imbrication, corresponding to channel fill. Above these are finer grained conglomerates of longitudinal bar origin (see Collinson, 1986), with or without coarse horizontal lamination. The overlying sandstones have horizontal lamination (Figure 9A) or ripple structures attributed to low water stage. The sequence is commonly capped with several meters of claystone or siltstone in which small floodplain channels may be incised. These sedimentary rocks can be divided into larger decameter- or hectometer-scale sequences. The latter are either fining-upward, as in the lower parts of the sedimentary fill of the Cuenca (Noblet et al., 1988) and Vilcabamba basin (Marocco et al., 1993), or coarsening-upward, as in the middle part of the B megasequence of the same basins (Figure 5). The abundance of debris flows intercalated in the fluvial sequences confirms the proximal character of the sedimentation.

Another type of fluvial environment is characterized by conglomeratic coarsening-upward sequences 10–100 m thick (Figure 9B). These sedimentary units are interpreted as alluvial fans that form the upper part of the fill in each of the basins studied (Figures 5, 6, 7, 8). These alluvial fans reflect rapid filling of the basins controlled by tectonic processes that uplifted the basin flanks. The crests of the alluvial fan sequences sometimes contain boulders several meters in diameter, such as in the upper sequences of the B megasequence in the Vilcabamba basin (Figure 5) (Marocco et al., 1993). The conglomerate beds are variable. They may be structureless or locally channelized or may consist locally of sheetflood deposits, sieve deposits, or debris flows. These facies types suggest a semi-arid climate during alluvial fan sedimentation (see Collinson, 1986). The clays that commonly form the lower parts of the alluvial fan sequences may contain red oxidation intervals and calcareous crusts, such as in the Vilcabamba basin (Marocco et al., 1993). These argillaceous lower parts indicate lower energy and reduced coarse sedimentation (such as during tectonic quiescence or long droughts) that favor soil formation.

Lacustrine Sedimentary Rocks

Lacustrine sediment units are important in the lower part of the sedimentary fill of each basin. Two main types of facies are represented: a sedimentary facies related to quiescent periods and a catastrophic sedimentary facies contemporaneous with intense tectonic activity. The granulometry of the lacustrine fill may thus reflect either tectonism or climatic changes. We ascribe the change from fine-grained to coarse-grained sedimentation to tectonic processes. Although there may be a climatic overprint, there are no data available on Neogene climates in the Andes.

The quiescent sedimentation facies are of two types that reflect their proximity to the shore zone. Clastic sedimentation occurred near river mouths where small

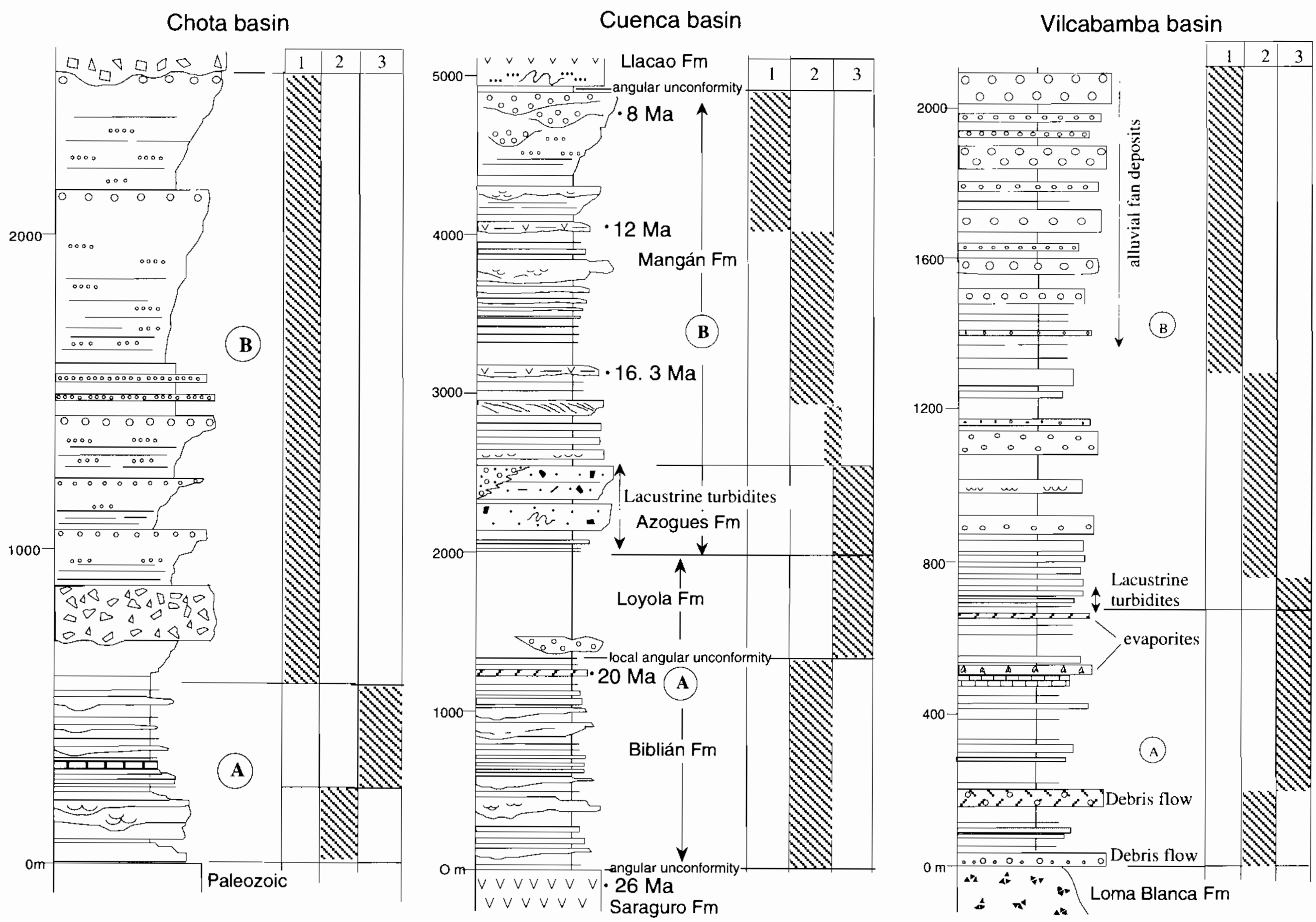


Figure 5—Stratigraphic sections of the Chota (Barragán, 1992), Cuenca (Noblet et al., 1988), and Vilcabamba (Marocco et al., 1993) basins, Ecuador. Circled letters refer to megasequences A and B. Columns: (1) alluvial fan, (2) fluvial, and (3) lacustrine.

Gilbert-type deltas were formed (Gilbert, 1885) which distributed sediments into the lake. The river floodplain sediments are mainly argillaceous and suggest swamps where bioturbation and rooting occurred. Biochemical sedimentation may have also occurred in these shallow zones, resulting in limestone with algae lamination and a fetid smell due to hydrocarbons, such as in the Vilcabamba (Figure 5) (Marocco et al., 1993) and Rumichaca basins (Figure 8) (Mégard et al., 1983). Evaporites are attributed to lake margin precipitation, such as in the Vilcabamba (Figure 5) and San Marcos basins (Figure 6) (Bellier et al., 1989). In deeper offshore zones, sedimentation was mainly due to suspension processes. This resulted in shales and very fine grained deposits that are white or pale yellow, well-stratified, and laterally persistent. Slump blocks and slump scars locally indicate the paleoslope of the lake and suggest some tectonic instability (e.g., Vilcabamba basin) (Marocco et al., 1993). Thin, low-density Bouma-type (1962) turbidite beds of centimeter to decimeter scale are commonly intercalated within the fine-grained sedimentary beds. Finally, in all of the basins, the abundance of interbedded volcanic and volcaniclastic strata in the lacustrine fill suggests contemporaneous volcanic activity.

The catastrophic sedimentary facies are represented by high-density megaturbidites that locally form thick successions in the Cuenca (Noblet et al., 1988; Noblet and Marocco, 1989) and Girón-Santa Isabel basins (Mediavilla, 1991). Figure 10 shows an idealized lacustrine megaturbidite sequence containing all the elements observed in various outcrops. The lower part of the sequence consists of debris flows (see Middleton and Hampton, 1976; Lowe, 1982) with erosional basal discontinuities that may be channeled but are generally flat with flute, prod, or groove casts. Clasts in a typical debris flow are centimeter to decimeter scale, whereas intraclasts of lower lacustrine strata can reach a diameter of several meters. Large clasts are commonly concentrated at the top of the bed, indicating that they were transported at the top of the turbidity flow before it solidified (Middleton and Hampton, 1976; Lowe, 1982).

Sandstones overlying the debris flows are characteristic of the high-density turbidites described by Lowe (1982). The S1 unit (or facies) of Lowe is the lowest. It is composed of coarse sandstones that are locally microconglomeratic and preserve some traction structures as well as poorly developed flat or oblique laminations. The second unit, S2, is composed of thin, centimeter-scale

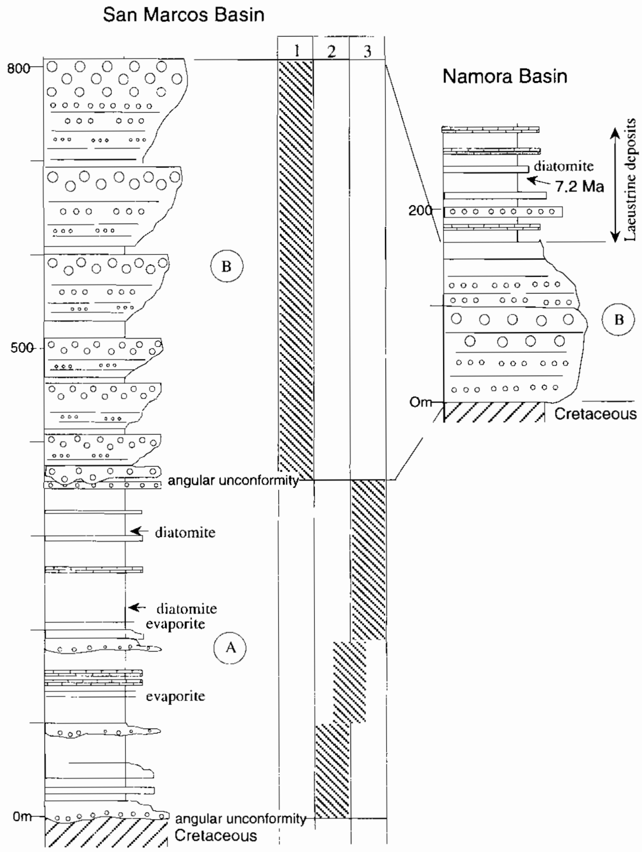


Figure 6—Stratigraphic sections of the San Marcos and Namora basins, northern Peru. Circled letters refer to megasequences A and B. Columns: (1) alluvial fan, (2) fluvial, and (3) lacustrine. (After Bellier et al., 1989.)

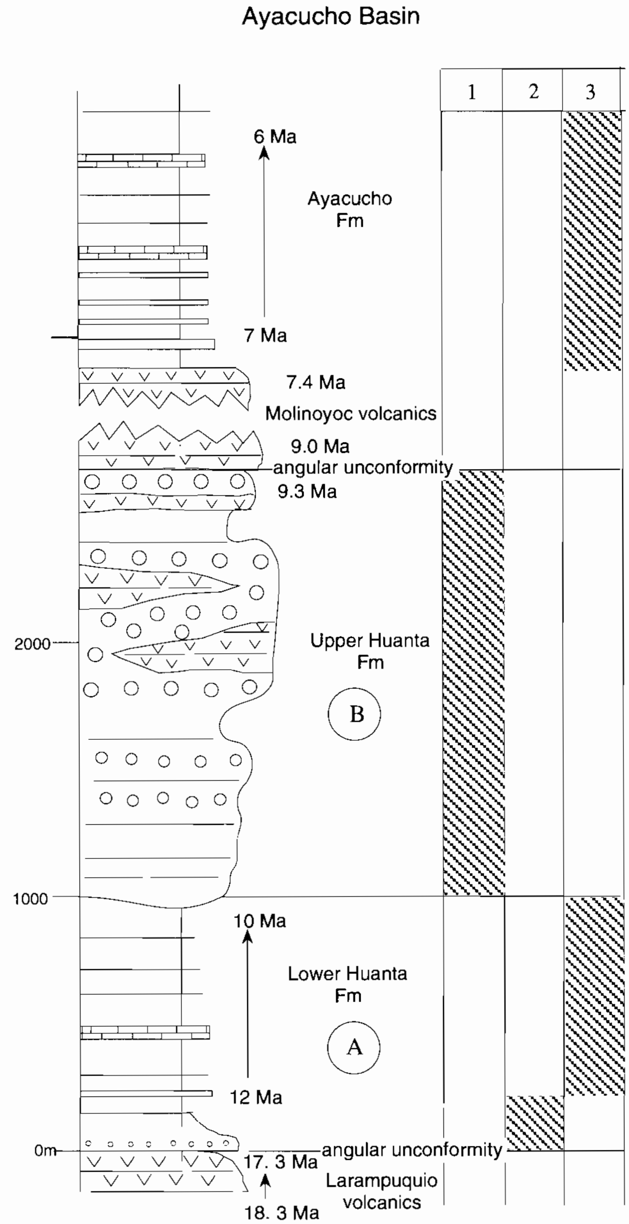


Figure 7—Stratigraphic section of the Ayacucho basin, south-central Peru. Circled letters refer to megasequences A and B. Columns: (1) alluvial fan, (2) fluvial, and (3) lacustrine. (After Mégard et al., 1984.)

coarsening-upward sandstone beds that formed by "traction carpet" processes. Their genesis, due to traction and suspension mechanisms, is linked to the increasing instability of the turbiditic flow. The S2 facies are rare in the intermontane basins studied, especially the Cuenca basin where lacustrine megaturbidites were recognized. Noblet and Marocco (1989) attribute this to the low proportion of spherical grains in the sediment, which is composed mainly of idiomorphic volcanic minerals.

The S3 facies is the uppermost unit of the high-density turbidites and consists of medium- to fine-grained massive sandstones that show dish structures. The S3 unit was deposited by rapid settling of the remaining sediments still in suspension. Megaturbidite sequences vary from 5 to 20 m in thickness, and their areal extent reaches 250 km2. The S3 facies is commonly capped by a low density Bouma-type turbidite, either complete (Ta–Te divisions) or consisting of only the most distal components (Tc–Te). In the Cuenca basin, Noblet et al. (1988) have shown that megaturbidites were related to a lacustrine delta, which formed typical coarsening-upward mouth bar sequences and received an abundant supply of volcaniclastic material (3000-m level, Cuenca basin section, Figure 5).

Sequence Stratigraphy

The same pattern of stratigraphic evolution characterized all of the basins. Initial fluvial depositional systems were followed abruptly by lacustrine deposition, and fluvial environments were reestablished. The upper fluvial deposition was progressively more proximal in character and terminated in alluvial fan deposition. The sedimentary units exhibit a sequential organization that allows correlation, either among different sections of a basin if the sequence-generating processes were local (such as hydrodynamic changes or local morphologic modifications) or among basins if the processes were

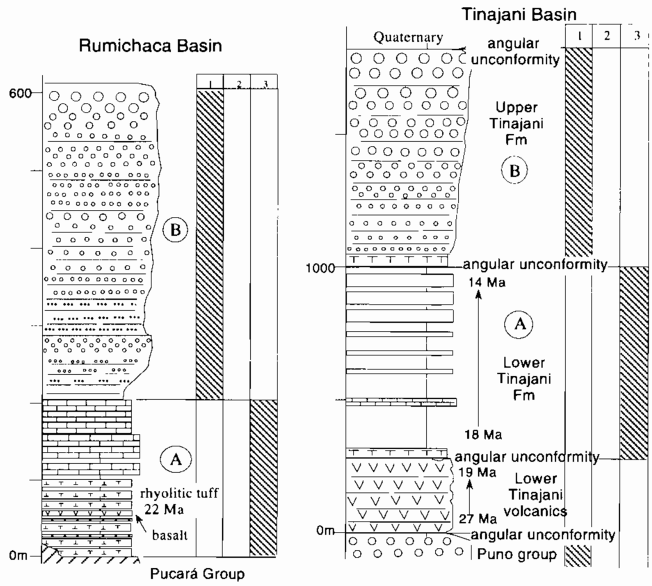


Figure 8—Stratigraphic sections of the Rumichaca (Mégard et al., 1983) and Tinajani (Sébrier et al., 1988) basins, south-central and southern Peru, respectively. Circled letters refer to megasequences A and B. Columns: (1) alluvial fan, (2) fluvial, and (3) lacustrine.

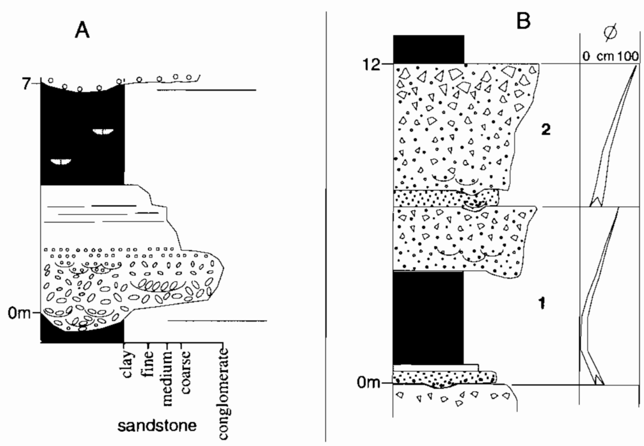


Figure 9—Two types of fluvial sequences observed in the sedimentary fill of the Río Chota Neogene intermontane basins, northern Ecuador. (A) Proximal fluvial sequence, lower part of the Río Chota basin sedimentary fill. (B) Two sequences (1 and 2) of alluvial fan deposits, upper part of the Río Chota basin sedimentary fill. (After Barragán, 1992.)

regional in nature (climatic changes, tectonic modifications, or eustatic variations in marine basins). This section addresses the large-scale organization of the deposits or megasequences in the basins.

The intermontane basins shown in Figures 5, 6, 7, and 8 preserve two types of stratigraphic columns: upward-fining set A and upward-coarsening set B. Thicknesses vary from hundreds of meters to kilometers in the Cuenca and Vilcabamba (Figure 5), Ayacucho (Figure 7), and Tinajani (Figure 8) basins. This two-set organization is also observed in other basins, including the Loja

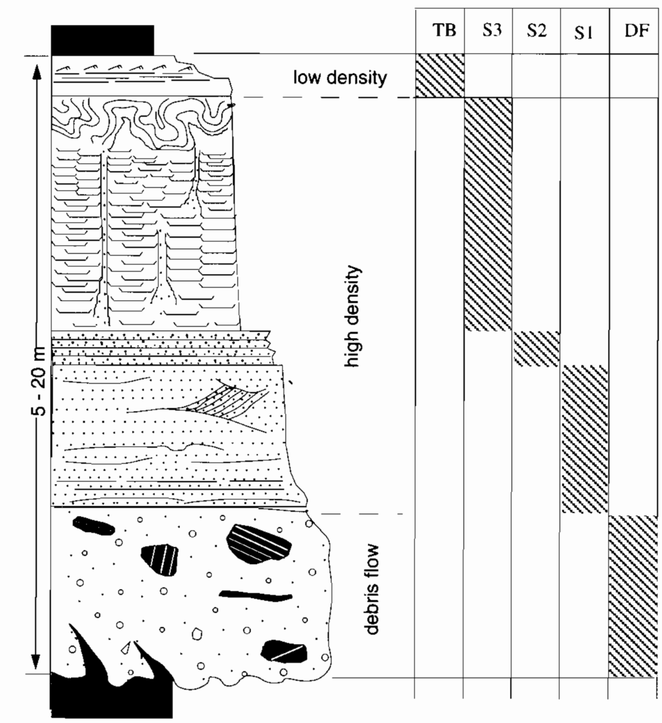


Figure 10—Example of a megaturbiditic sequence in the Azogues Formation, Cuenca basin, Ecuador. TB, Bouma turbidite; DF, debris flow; S1, S2, S3, Lowe's (1982) divisions of a megaturbidite. (After Noblet and Marocco, 1989.)

(Izquierdo, 1991), Girón-Santa Isabel (Mediavilla, 1991), and Nabón (Winkler et al., 1993) basins of Ecuador (Figure 3) and the Cajabamba (Bellier et al., 1989), Callejón de Huaylas (Bonnot et al., 1988), and Paruro (Mendivil, 1979) basins of Peru (Figure 4). The Pliocene–Pleistocene Anta and Ccatca basins of the Cuzco region in southern Peru (Figure 4) (Cabrera, 1988) show a fining-upward sedimentary organization, but they are younger than the basins of this study and are still evolving.

The available chronologic data show that sets A and B are not the same age in all basins. In the Cuenca basin (Noblet et al., 1988) (Figure 5), the age of the discontinuity between sets A and B is bracketed between 20 and 16.3 Ma. In the Ayacucho basin (Mégard et al., 1984) (Figure 7), set A starts at about 15 Ma and finishes at about 10 Ma. Finally, in the Tinajani basin, set A was deposited between 18 and 14 Ma. The term *megasequence* used here for convenience to name the sedimentary sets A and B is valid only at the scale of an individual basin. This may reflect the lack of detailed biostratigraphy. The fossil flora in the Ecuadorian basins was studied more than 50 years ago. A reexamination of these flora may indicate that the megasequences are contemporaneous even between basins.

The discontinuity between megasequences A and B is always well defined. It generally consists of an angular unconformity, as in the Cuenca (Figure 5), San Marcos (Figure 7), and Tinajani basins (Figure 8). However, even without discordances, the discontinuity is expressed by a drastic change in the sedimentary rocks, from distal beds at the top of megasequence A (low-energy lacustrine) to proximal deposits with a notable increase in grain size at the base of megasequence B (lacustrine megaturbidites, proximal fluvial).

Megasequence A

The lower megasequence (A) begins with proximal fluvial deposits that grade abruptly into deep lacustrine strata. In the upper part, sequences 20–100 m thick of shoaling-upward lacustrine deposits are generally present, showing deep facies in their lower parts and progressively shallower facies in their upper parts. In the southern Ecuador Vilcabamba basin (Figure 5) (Fierro, 1991; Marocco et al., 1993), the top of the lacustrine shoaling-upward sequences preserve swamp facies: limestones with algal laminations and a fetid smell and coal and evaporite beds that are exploited manually. The thickness of the lower fluvial part of the series varies among basins. It exceeds 1000 m in the Cuenca basin, where it constitutes the Biblián Formation (Figure 5) (Noblet et al., 1988), whereas it is generally thinner than 100 m in the other basins. The Cuenca basin, however, has subsided more than the other basins.

In summary, during deposition of megasequence A, a progressively more distal location of the source areas is reflected in the general fining-upward trend. This pattern is interpreted as a sedimentary response to basin subsidence. The initial depression was controlled by fault systems, which trended parallel to the latest Oligocene orogenic fabric (roughly north-south in Ecuador and NNW-SSE in Peru) and which concentrated drainage. The earliest deposits of the basin were fluvial. Subsequently, marked subsidence trapped the river systems structurally, resulting in widespread lacustrine deposition of the lower megasequence in the Neogene Andean basins. However, the early Miocene was also characterized by a eustatic sea level 100 m higher than the present (Haq et al., 1987). If the rate of uplift in the central Andes during the Neogene (0.15 mm/year) estimated by Sébrier et al. (1979, 1988) is valid, the lakes must have been at a relatively low altitude, a few tens or hundreds of meters above sea level 20 Ma ago. In this setting, the high sea level may have maintained a high base level, favoring preservation of a lacustrine regime, which was, nevertheless, controlled by subsidence.

Megasequence B

Megasequence B is marked in its lower part by the abrupt change from a low-energy sedimentation regime to one of coarse sedimentation. Thus, in the Cuenca basin (Figure 5), quiescent lacustrine sedimentation of the Loyola Formation grades upward into turbiditic and megaturbiditic sedimentation of the Azogues Formation (Noblet et al., 1988). This is also characteristic of the Girón-Santa Isabel basin located just southwest of the Cuenca basin (Mediavilla, 1991) and which was probably connected to it. In the Vilcabamba basin (Figure 5), the base of megasequence B is also composed of conspicuous coarse turbidites (0.5-mm-diameter clasts); however, megaturbidites such as those in the Cuenca and Girón-Santa Isabel basins are absent. Neither did the Vilcabamba basin receive as much volcaniclastic material as the Cuenca basin. Turbidites filled the lake and were overlain by proximal prograding fluvial strata; these in turn were succeeded by typical coarsening-upward alluvial fan deposits such as those previously described (Figure 9B). Turbidites or megaturbidites have not been recognized in the other basins. Thus, megasequence B reflects a significant change in sedimentation from quiescent lacustrine to progressively coarser. This change is attributed to the increasing tectonism that affected basin and provenance areas, as evidenced by the progressive discordances observed in this megasequence in the Cuenca (Noblet et al., 1988), Vilcabamba (Fierro, 1991), and Chota basins (Barragán, 1992).

In the Namora (Figure 6) (Bellier et al., 1989) and Ayacucho basins (Figure 7) (Mégard et al., 1984), the coarse alluvial fan deposits forming the top of megasequence B are overlain by lacustrine deposits of latest Miocene age, radiometrically dated at 7–6 Ma in the Ayacucho basin (Ayacucho Formation) and at 7.2 Ma in the Namora basin. This lacustrine sedimentation, which reflects a conspicuous modification of the tectonic regime about 7 Ma, coincides with a discontinuity or interruption in the compressive regime that was building the central Andean orogenic belt. Such uppermost Miocene lacustrine strata have not been observed in the other basins, either because they have been eroded or because volcanic activity concealed their presence. In the Namora and Ayacucho basins, the local structural framework and tectonic regime is believed to have resulted in reactivation of subsidence, thus creating a lacustrine basin contemporaneous with lesser amounts of thrusting in neighboring areas, which explains the absence of coarse sedimentation.

TYPES OF BASIN EVOLUTION

The tectonic evolution of the Neogene intermontane basins of the central Andes can be grouped into two main types according to the orientation of their bounding structures in relation to the orientation of the Neogene stress fields. The first type is the strike-slip basin, of which we review the Cuenca basin of Ecuador as an end-member (Noblet et al., 1988). The second type of basin is linked to reverse faulting activity; we emphasize the Rumichaca basin of central Peru (Mégard et al., 1983). However, because of the variable nature of Neogene stress fields, each basin has evolved through several stages of the stress regime: extension, transpression, and compression. Each basin is characterized by a dominant stress stage, the one that lasted longest and controlled most of the sedimentation of that basin and deformation of its environs.

From a structural perspective, our knowledge varies greatly from basin to basin, with the Cuenca basin being the best known. This complicates comparison, especially where an author has emphasized only one aspect of basin evolution. For example, in the northern Peruvian basins of Namora, San Marcos, and Cajabamba, Bellier et

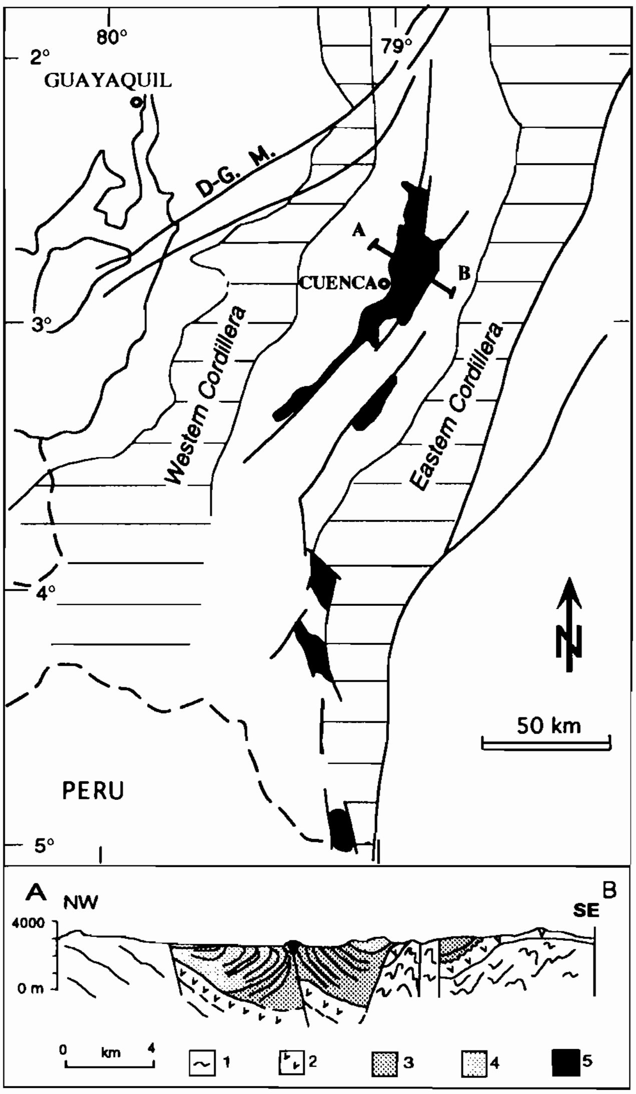


Figure 11—Tectonic setting of the Cuenca basin in Ecuador (after Noblet and Marocco, 1989). Key: 1, Mesozoic substratum; 2, Saraguro Formation; 3, first megasequence (A); 4, second megasequence (B); 5, Andesite of the Cojitambo dome (6.3 Ma; Barberi et al., 1988).

al. (1989) have stressed the extensional phase of basin evolution.

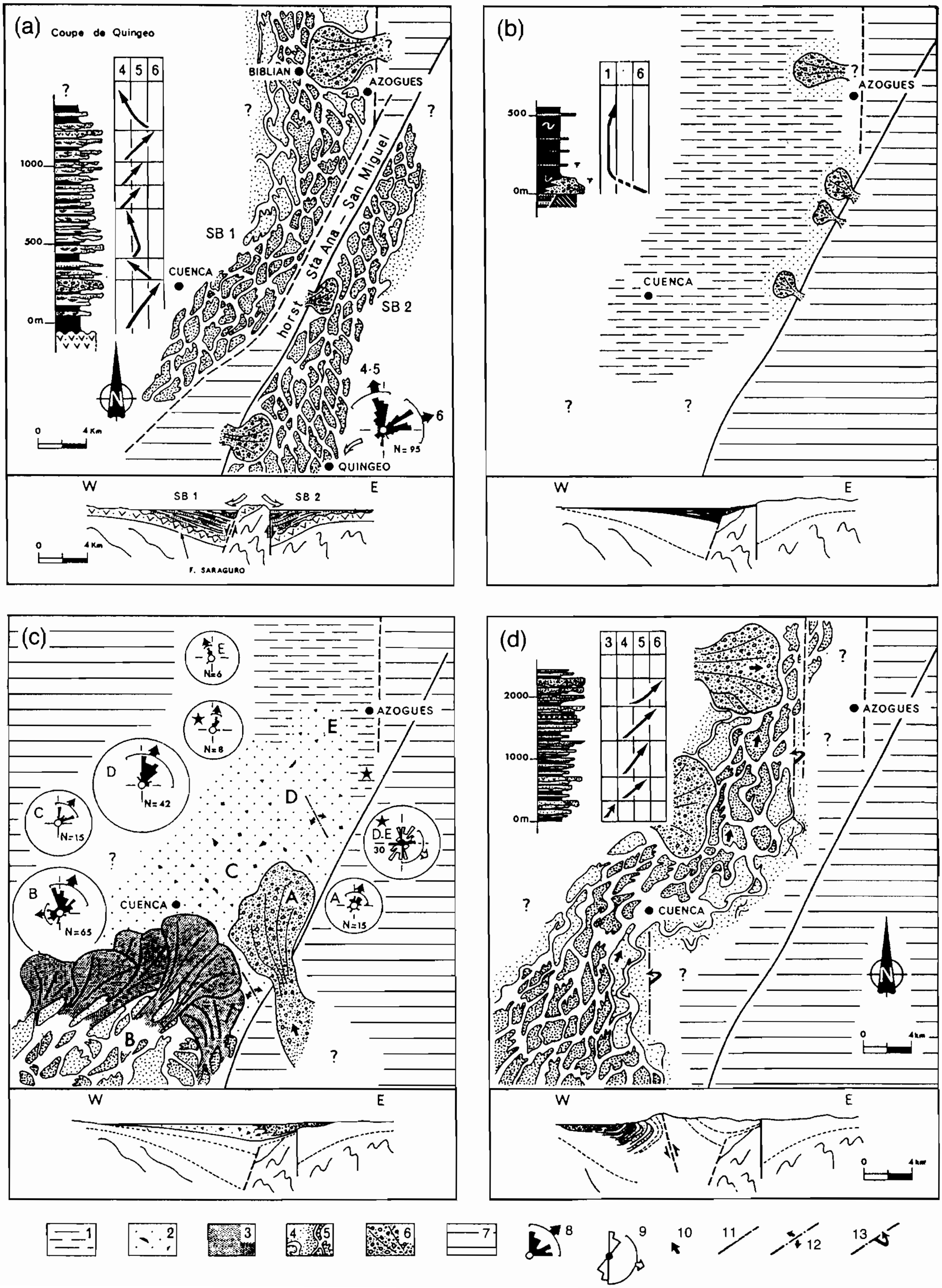
Cuenca Strike-Slip Basin

The Cuenca basin is located in southern Ecuador (Figures 3, 11) and has been studied by Noblet et al. (1988) and Noblet and Marocco (1989). It is the largest Neogene intermontane basin in this part of the Andes that we have studied (100 km × 30 km). It is controlled by faults trending N 170° E to north-south and NNE-SSW. Figure 5 summarizes the overall coarsening-upward succession that is more than 4000 m thick in this basin and consists of megasequences A and B. Its sedimentary evolution is summarized in Figure 12, where maps (a) and (b) represent megasequence A and maps (c) and (d) represent megasequence B.

Sedimentation began with establishment of fluvial systems on either side of the Santa Ana–San Miguel horst (Figure 12a); drainage of the Biblián Formation was toward the north and north-northeast. The Santa Ana–San Miguel horst was probably formed during the same tectonic events at the end of the Oligocene that were responsible for creation of the early Cuenca basin. As the Santa Ana–San Miguel horst disappeared in the earliest Miocene and depocenters were yoked together emphasizing the western depocenter, a lacustrine basin was established. The middle Miocene Loyola Formation was deposited in this lake (Figure 12b). Near the end of the middle Miocene, the influx of large volumes of detrital sediments into the basin (Figure 12c) marked the beginning of megasequence B. To the south of the basin, a fluvial system flowed into the lake, forming a delta that prograded toward the northeast. As a result of sediment stacking and tectonic instability, the lacustrine delta front is believed to have collapsed, resulting in the lacustrine turbidites and megaturbidites of the Azogues Formation. In the late Miocene, fluvial sediments of the Mangán Formation prograded into the lake and completely filled it.

The tectonic evolution of the Cuenca basin can be interpreted from analysis of the folding and faulting. Based on a study of the effects of different periods of synsedimentary folds, Noblet et al. (1988) established a basin model related to strike-slip movements of the bordering faults. Figures 13 and 14 show the synsedimentary folds that were formed during three main compressive periods:

1. The first period coincided with sedimentation of the Biblián Formation and resulted locally in conical folds with axes trending N 120° E (Figure 14A) and in sedimentary pinch-outs along faults trending N 20°–40° E. These structures are compatible with a dextral strike-slip tectonic regime along the approximately north-south trending faults with horizontal axes σ_1 and σ_3 approaching N 30° E and N 120° E, respectively.
2. The second period occurred at the beginning of turbiditic sedimentation of the Azogues Formation. It was a compressive event that also resulted in conical folds (Figure 14B, C) with axes trending close to N 150° E and showing a shortening direction of about N 60° E. These folds observed in the vicinity of the fault trending N 20°–40° E along the eastern border of the basin are compatible with dextral strike-slip movements along that trend.
3. The third prominent compressive period occurred throughout deposition of the Mangán Formation. Numerous hiatuses affected 2000 m of deposits (Figure 13). Analysis of these structures and of those induced by deformation of the underlying beds shows that the fold axes have an approximately north-south orientation (N 170° E in the northern part of the basin, N 20°–30° E in the south) and that shortening approaches an east-west direction. Diagrams D–H in Figure 14 illustrate this third compressional period.


(a) Coupe de Quingeo
4 5 6
?
1000
500
0m
BIBLIAN
AZOGUES
CUENCA
SB 1
SB 2
horst Sta Ana — San Miguel
4·5
6
N= 95
QUINGEO
0 4 Km
W
E
F. SARAGURO
(b)
1 6
500
0m
AZOGUES
CUENCA
(c)
AZOGUES
CUENCA
A
B
C
D
E
N=6
N=8
N= 42
N=15
N=65
D.E 30
N=15
(d)
3 4 5 6
2000
1000
0m
AZOGUES
CUENCA
0 4 km
1 2 3 4 5 6 7 8 9 10 11 12 13

Fracture analysis, particularly of the fault systems bounding the basin (Lavenu and Noblet, 1989), confirms the model of a basin related to strike-slip movements interpreted by Noblet et al. (1988) on the basis of fold deformation. This implies that between the latest Oligocene and late Miocene–Pliocene phase of basin subsidence compression directions experienced a clockwise rotation, from N 20° E to N 60° E (opening phase of the basin expressed in the Biblián, Loyola, and Azogues formations) and finally to N 100° E (closing phase of the basin reflected in the Mangán Formation). Figure 15 summarizes the geodynamic evolution of the Cuenca basin, as well as that of the south Ecuadorian Neogene basins of Nabón, Loja, Vilcabamba-Malacatos, and Zumba.

The nature of the structural evolution of the Cuenca basin, the thick sedimentary accumulation comparable to that of the Vienna pull-apart basin (Royden, 1985), and the thicker filling in the proximity of the most active faults all indicate that the Cuenca basin is linked to the array of strike-slip faults with characteristics similar to those described by Nilsen and McLaughlin (1985).

The Ecuadorian Vilcabamba (Fierro, 1991; Marocco et al., 1993) and Girón-Santa Isabel basins (Mediavilla, 1991) are also of pull-apart type, at least during part of their evolution (Figure 15). The Neogene basins of Peru are less well known; some of them may have had a geodynamic evolution comparable to that of the Cuenca basin.

Rumichaca Reverse Fault Linked Basin

The Rumichaca basin of central Peru (Figure 4) has been studied by Mégard et al. (1983). It is elongated in a north-south direction (Figure 16) and has modest dimensions (5 km × 1.5 km). It is limited to the west by a north-south trending reverse fault and by the Huapa anticline that deforms the Mesozoic succession.

The Rumichaca succession, attributed to the early–middle Miocene by Petersen et al. (1977), consists of approximately 600 m of continental, partially volcaniclastic sedimentary rocks (Figure 8). To the east, these strata rest with angular unconformity on paleorelief carved into the Liassic limestones of the Pucará Group; to the west, they are in reverse fault contact with Cretaceous sandstones and limestones (Figure 16A). The lower 100 m consists of volcanic tuffs that are variably altered and commonly resedimented, with thin decimeter-scale intercalations of lacustrine limestone beds. A tuff located in the upper part of this unit has been dated at 22 Ma (McKee and Noble, 1982). The tuffs are overlain by 100 m of lacustrine limestones with common algal laminations that pinch out toward the west and south. Along the western margin, near the basin-bounding fault, the limestones contain interbeds of conglomerates with clasts of Cretaceous sandstones and limestones, indicating fault activity during the lacustrine calcareous sedimentation. The upper 400 m consists of conglomerates of proximal fluvial or alluvial fan origin which reflect conspicuous activity of the western fault. Clast size of the conglomerates increases upward and from east to west.

The structure of the basin is dominated by the Rumichaca syncline (Figure 16). On the Lircay-Huachocolpa highway along the Huachocolpa River, the Rumichaca syncline is markedly asymmetric with a vertical western flank. Detailed mapping of the basin shows that it is bordered to the west by a reverse fault and that the sedimentary fill of the basin is punctuated by numerous synsedimentary unconformities near the

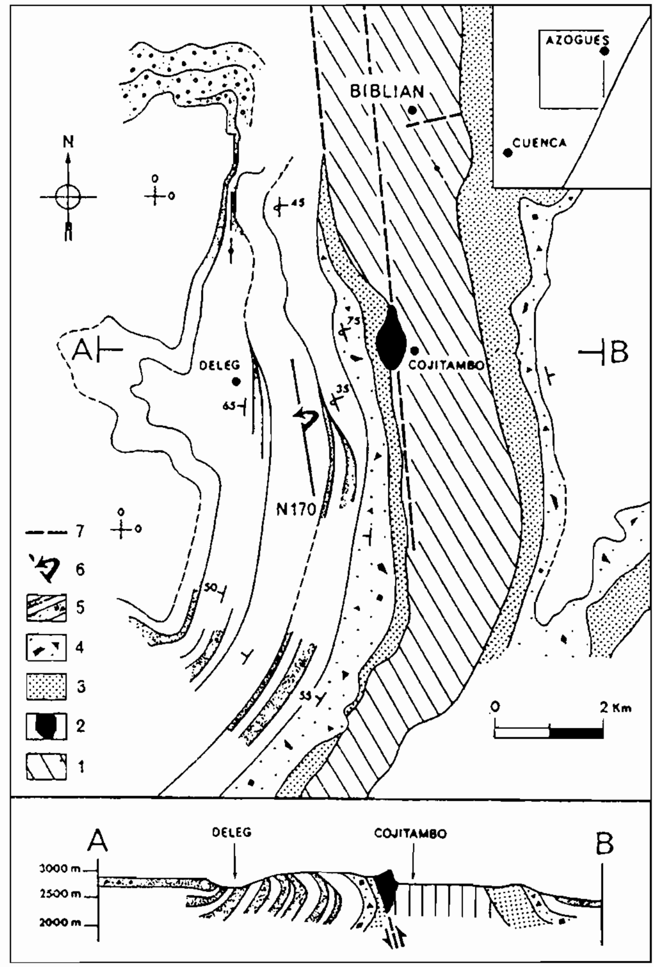


Figure 13—Outcrop map and synthetic cross section of the progressive discordance affecting the Mangán Formation (after Noblet et al., 1988). Legend: 1, Biblián Formation; 2, Cojitambo andesite; 3, Loyola Formation; 4, Azogues Formation; 5, Mangán Formation; 6, fold axes (third synsedimentary tectonic event); 7, probable fault.

(facing page)
Figure 12—Schematic paleogeographic map and synthetic cross section for each formation of the Cuenca basin (after Noblet et al., 1988). (a) Biblián Formation; (b) Loyola Formation; (c) Azogues Formation; (d) Mangán Formation. Legend: 1, lake; 2, lacustrine megaturbidites; 3, lacustrine delta; 4, flood plain; 5, braided river; 6, alluvial fan; 7, relief; 8, vector transport (N is number of measurements); 9, slump axes; 10, vector transport from some measurements; 11, fault; 12, conical fold axes of second synsedimentary tectonic event; 13, conical fold axes of third synsedimentary event. SB1 and SB2 are subbasins of the Biblián Formation.

reverse fault ("u" in Figure 16). In contrast, the Neogene strata of the eastern part of the basin are conformable, demonstrating that the eastern border was tectonically passive during Neogene sedimentation. From north to south in the basin, close to the contact with the western reverse fault, progressively lower levels of the stratigraphy crop out. The observed structural attitude of the strata is shown in Figure 16C. This structural–stratigraphic architecture resembles that along the active border of the Alto Cardener basin on the southern side of the Pyrenees (Riba, 1973, 1974).

Synsedimentary deformation of the Rumichaca basin began after 22 Ma and was related to formation of the Huapa anticline. Deformation probably ceased at about 10.5 Ma, the age of the Julcani rhyodacitic domes (McKee and Noble, 1982) that cross cut the structures affecting the Rumichaca basin fill.

The Río Chota basin in northernmost Ecuador (Figure 3) has an evolution comparable to that of the Rumichaca basin during megasequence B deposition. Marked synsedimentary tectonic activity, with an initial shortening direction of N 120° E that later changed to east-west, affected the conglomerates of the upper part of the succession. These are alluvial fan deposits that prograded eastward as a consequence of the activity of the reverse fault system that forms the western margin of the basin (Barragán, 1992). In contrast, deposition of the lower megasequence (A) is characterized by a N 130° E extensional regime. Evolution of the Neogene Nabón basin of southern Ecuador (Figure 3) is also linked to activity of a bordering reverse fault (Winkler et al., 1993).

Similar processes controlled the northern Peruvian Namora, San Marcos, and Cajabamba basins (Bellier et al., 1989), in which deposition of megasequence A (Figure 6) was coeval with an extensional regime (σ_3 changed from ENE-WSW to northeast-southeast). There, coarsening-upward megasequence B is coeval with a compressional regime in which the σ_1 directions changed from ENE-WSW to north-south, with the former normal faults that controlled deposition of megasequence A being reactivated as reverse faults.

CONCLUSIONS

Figure 17 summarizes the evolution of the Neogene intermontane basins of Ecuador and Peru. Their evolution begun about 28–26 Ma ago with renewed Andean tectonism after a long period of relative inactivity following the 42-Ma late Eocene compressive episode (Sébrier et al., 1988). The evolution of these basins ended at about 7 Ma. The Ayacucho and Tinajani basins in Peru do not follow the two nearly simultaneous

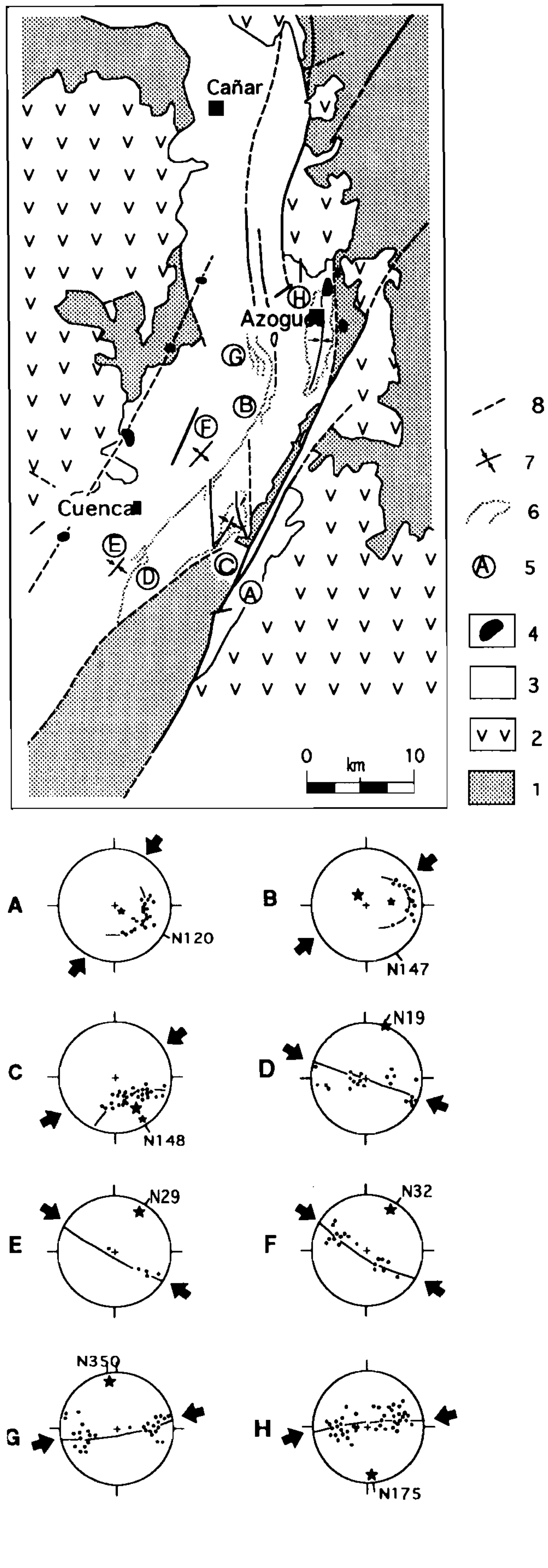


Figure 14—Schematic structural map of the Cuenca basin (after Noblet et al., 1988). Legend: 1, Mesozoic substratum; 2, Cenozoic volcanics; 3, Cenozoic sedimentary rocks; 4, travertines; 5, sites of tectonic measurements; 6, Azogues and Mangán formations; 7, fold axes; 8, fault. Diagrams AH are explained in text.

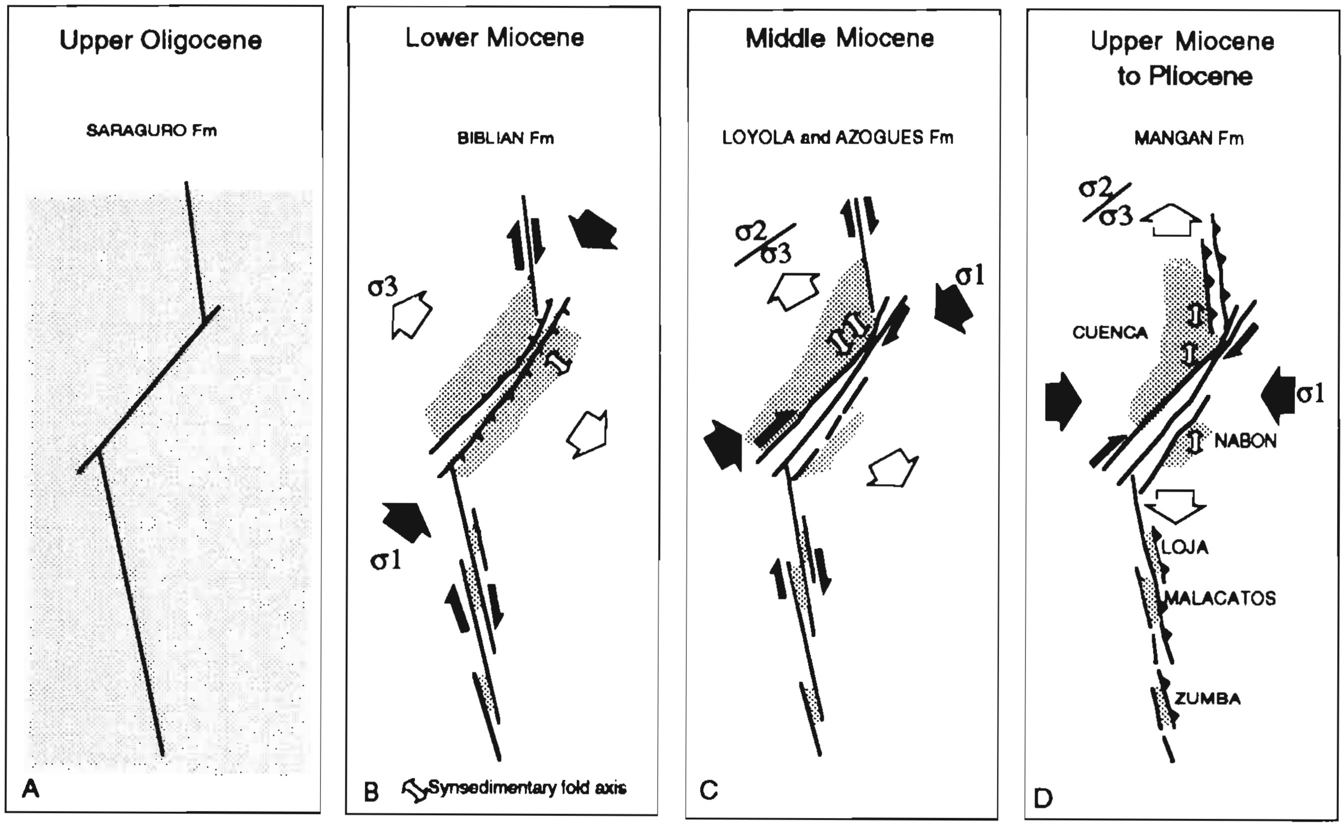


Figure 15—Evolution of the Neogene intermontane basins of southern Ecuador (after Noblet et al., 1988). (A) Location of probable ancient faults sealed by Saraguro Formation deposits. (B) First synsedimentary tectonic event, with compression oriented N 30° E. (C) Second synsedimentary tectonic event, with N 60° E compressional direction. (D) Third synsedimentary tectonic event, with an east-west compressional direction. Dotted pattern is sedimentary deposits synchronous with deformation.

megasequence models; the opening phase in the Ayacucho ended about 10 Ma (Figure 7) and in the Tinajani about 14 Ma (Figure 8). There are two possible explanations for this difference: either basin evolution in Peru south of Ayacucho is different from that of the northern regions or the chronostratigraphy (especially radiochronology) for all the Peruvian basins needs to be revised.

The initiation of basin formation and subsidence coincided with the last major reorganization of the Pacific oceanic plates at 27–25 Ma, which caused partition of the Farallon plate into the Cocos and Nazca plates (Handschumacher, 1976; Pilger, 1984). This new organization was associated with reorientation (N 80° E) and acceleration of the convergence rate between the Nazca and South American plates (Minster and Jordan, 1978; Pilger, 1983; Duncan and Hargraves, 1984).

Taking global geodynamic reorganization into account, Lavenu and Noblet (1989) proposed a model for the evolution of the Ecuadorian Neogene basins (Figure 14). The 50° obliquity between the convergence direction and the orientation of the Ecuadorian active margin caused the northward displacement of the coastal block, which accreted in the latest Cretaceous (Mégard et al., 1986). This displacement induced a dextral translation movement along the preexisting north-south faults, resulting in opening of the basins. Progressive blocking of the wrenching movement caused clockwise rotation of the stress, until it reached an east-west shortening direction (about 7 Ma), roughly parallel to the direction of convergence. This caused reverse faulting activity along the north-south faults, which provoked closure of the basins.

The extension observed in the Río Chota basin during deposition of megasequence A is difficult to interpret. It is perpendicular to the compression direction that affected the more southward regions (such as the Cuenca and Vilcabamba basins) during the same period. Because of local characteristics, extensional structures in the Río Chota basin probably presented a better expression than compressive structures. The similarities in patterns of evolution of the Ecuadorian basins is also observed in the Girón–Santa Isabel (Mediavilla, 1991), Nabón (Winkler et al., 1993), and Loja (Izquierdo, 1991) basins, including the ages of the sequences and their bounding surfaces and their tectonic characteristics. Only the Zumba basin, straddling the Ecuador–Peru border, is largely unknown.

The Peruvian margin, at least south of the Huacabamba deflection (south of 5° S lat), does not comprise accreted terranes, which may explain the important differences in geodynamic evolution seen among the Neogene basins of Peru and Ecuador. Basin

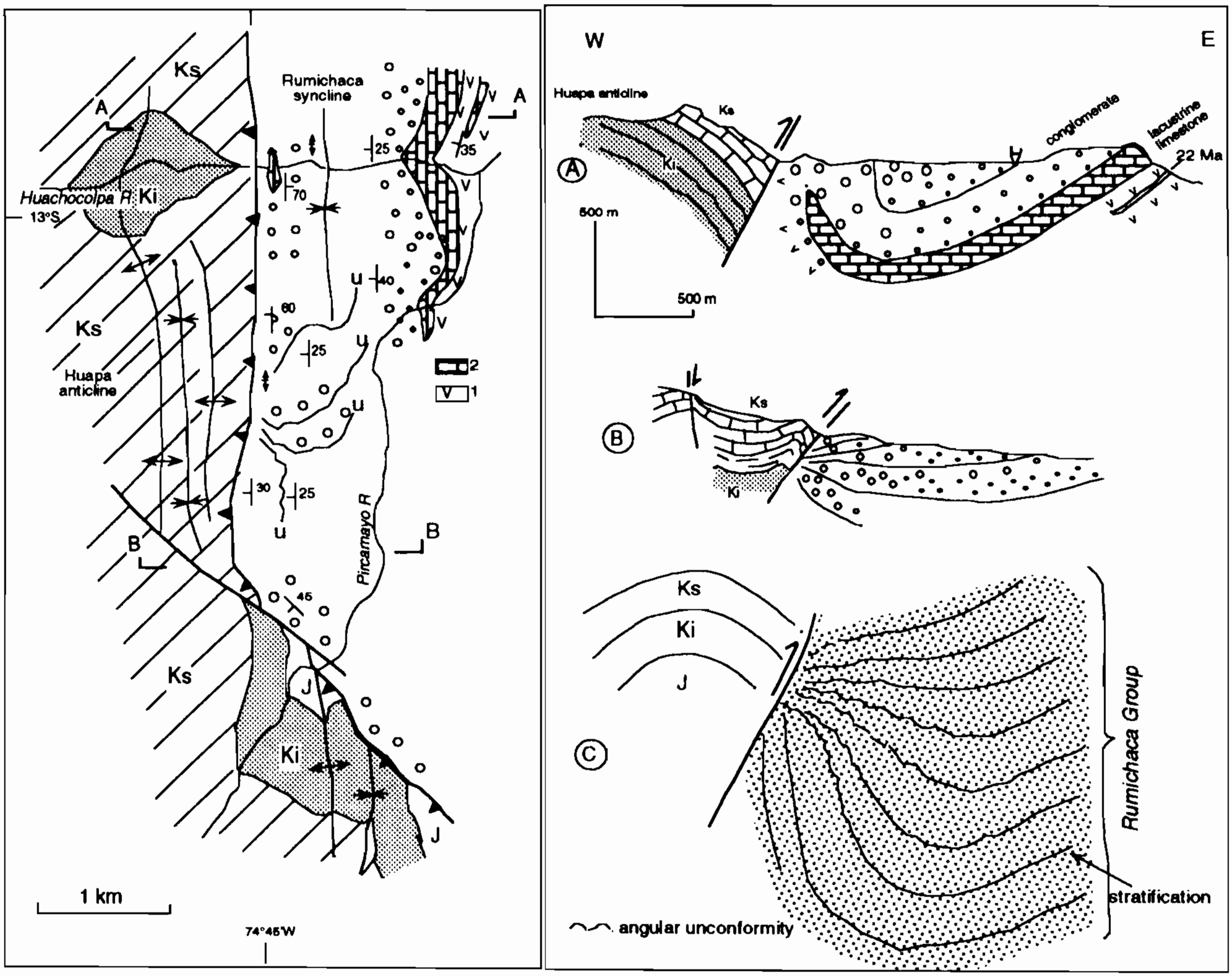


Figure 16—Geologic map and cross sections (AC) of the Rumichaca basin (after Mégard et al., 1983). J, Jurassic (Pucara Group); Ki, Lower Cretaceous (Goyllarisquizga Formation); Ks, Upper Cretaceous (Chulec Formation); 1, volcanics; 2, lacustrine limestones; u, intraformational angular unconformity.

evolution in Peru has been controlled by active margin processes and Andean fold and thrust belt deformation since the late Oligocene. Simultaneous with eastward encroachment of the thrust belt, the Andes were uplifted. The adaptation of preexisting structures to these two processes explains the origin of the Neogene intermontane basins. Bellier et al. (1989) interpreted the extension during the deposition of megasequence A in the Namora, Cajabamba, and San Marcos basins as a result of the flow of Andean material toward the trench due to gravity forces and weak coupling within the subduction zone. If this explanation is correct, then the presence in Ecuador of a coastal terrane between the trench and the rising Andes must have opposed this gravity flow.

Evolution of the Neogene intermontane basins of the Andes of Peru and Ecuador follows two stages:

1. The first stage represented by megasequence A spans the establishment of the initial basin in which progressively more distal sediments accumulated. This period was contemporaneous with extension or compression that was oblique to the preexisting faults controlling the basins. This opening phase of basin formation lasted until the end of the early Miocene, that is, when the main stress directions changed from NNE-SSW to northeast-southwest (Figure 15).
2. The second stage represented by megasequence B defines basin closure during the middle–late Miocene as compression progressively approached an east-west orientation. Movement along the bordering faults in a reverse sense caused uplift of the margins and influx of progressively coarser sediments that overfilled the basin.

There is insufficient stratigraphic information to correlate megasequences A and B of the Neogene intermontane basins with those of the coeval sub-Andean foreland basin or coastal basins with confidence. In the sub-Andean basin of northern Peru, Marocco (1993) has shown that sedimentation proceeded in three coarsening-upward megasequences (sequence N1, 28–10 Ma;

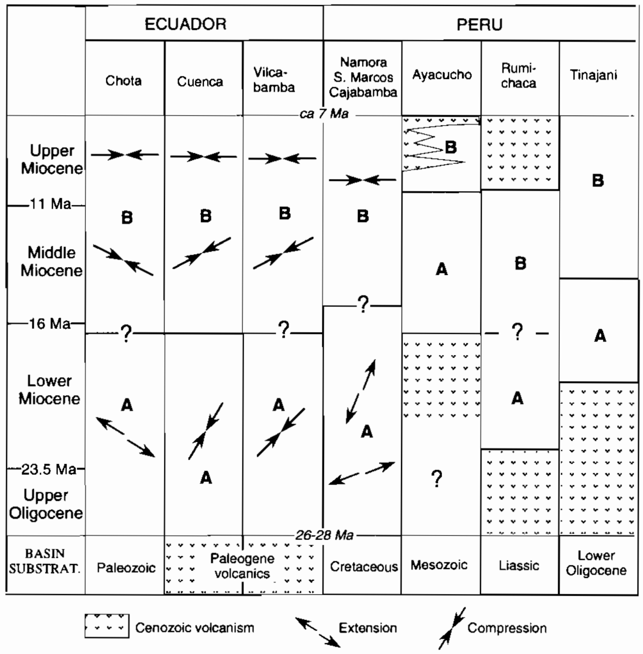


Figure 17—Summary of the evolution of the Andean Neogene intramontane basins.

sequence N2, 10–7 Ma; and sequence N3, 7–2.7 Ma), thus forming a coarsening-upward succession. Sequences N1, N2, and N3 represent three stages in the eastward propagation of the Main Andean thrust (Figure 2). Although tenuous, we correlate the intermontane megasequences A and B with the first two sequences (N1 and N2) of the sub-Andean foreland basin.

West of the continent, the few available syntheses of the Ecuadorian coastal forearc marine basins of Neogene age (Baldock, 1982; Egüez et al., 1991) show that their fills are organized into four fining-upward sequences: N1 (28–26 to 17–15 Ma), N2 (17–15 to 10 Ma), N3 (10–7 Ma), and N4 (7–2 Ma). The geodynamics of the coastal basins are controlled by prominent eustatic variations and indirectly by Andean tectonics, which in each pulse supplied coarser detrital deposits to the coastal basins, thus forming the base of the megasequences. It is impossible to establish intrabasin chronologic correlations of sequences in the three morphostructural regions (coast, Andean zone, and sub-Andean zone) because the processes controlling the evolution of the basins in each region were different.

The results of the study of the intermontane basins of Ecuador and Peru expand our understanding of Neogene tectonism in both countries (Dalmayrac et al., 1980; Sébrier et al., 1988). During the Neogene, the Andes experienced persistent compression, as demonstrated by the synsedimentary discordances and the chronology of brittle deformation observed in most of the basins. For the Neogene at least, the hypothesis of deformation by short tectonic phases separated by long periods of tectonic quiescence is incorrect. What, then, is the significance of the main tectonic "phases" marked by angular unconformities that occurred simultaneously in the central Andes at 28–26, 17–15, 10, 7, and 2.7 Ma (Sébrier et al., 1988)? The angular discordances possibly express short breaks or kinematic modifications in the tectonic continuum. Sempere (1991) has come to a similar conclusion in his study of the Cenozoic basins of Bolivia.

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