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Implications of Deep-Marine Miocene Deposits on the Evolution of the North Patagonian Andes

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ABSTRACT
The North Patagonian Andes (38°–43°30′S) present topographic and geologic characteristics distinct from those of the Central Andes to the north, including a lower altitude, a predominance of plutonic rocks, and the presence of a major N-S-trending, dextral strike-slip, intra-arc discontinuity known as the Liquiné-Olqui Fault Zone (LOFZ). The timing for the uplift of the North Patagonian Andes remains poorly constrained principally because high exhumation rates resulted in the erosion of most of its volcano-sedimentary cover during the late Cenozoic. The strongly deformed Cenozoic marine deposits ascribed to the Ayacara Formation, which crop out on the western flank of this range at ∼42°S, can provide for a better understanding of the recent geologic evolution of this range. To discern this formation’s age and sedimentary environment, we have integrated sedimentology, ichnology, foraminiferal paleontology, and geochronology. Our results indicate that this unit was deposited in a deep-marine environment during the Early-Middle Miocene. Correlation of the Ayacara Formation with coeval deep-marine deposits in the forearc to the west indicate that deposition was caused by a major regional event of tectonic subsidence probably related to subduction erosion. An Early-Middle Miocene age for the Ayacara Formation is a reliable maximum age for the deformation and uplift of the western flank of the North Patagonian Andes as well as for the commencement of transpressional deformation associated with the LOFZ.

Online enhancements: appendix, supplementary tables.

Introduction
The Andean Cordillera, the longest and highest mountain belt formed in a convergent setting, extends for ~4000 km along the Chilean margin. The origin of this range is related to the subduction of the oceanic Nazca (formerly Farallon) plate beneath South America, at least since the Jurassic (Mpodozis and Ramos 1989). Uyeda and Kanamori (1979) related the formation of the Andean range to the strong mechanical coupling between the upper and lower plates that characterizes the “Chilean subduction mode.” They ascribed such a strong coupling to a high convergence rate and a relatively young and buoyant oceanic plate. More than 30 yr of research, however, has shown that their model is too simplistic, and several issues remain highly debated. First, the uplift of the Andean Cordillera has been ascribed to various causes, such as (a) the convergence rate between the oceanic and continental plates (e.g., Pardo-Casas and Molnar 1987); (b) the absolute westward motion of South America (e.g., Silver et al. 1998); (c) the amount of sediment along the plate interface, which determines shear
stress in the subduction zone and is ultimately controlled by climate [Lamb and Davis 2003]; [d] the age of the subducting plate [Yañez and Cembrano 2004]; [e] a combination of the overriding velocity of the South American plate, crustal thickness, and friction coefficient in the subduction channel (Sobolev and Babeiko 2005); and [f] the subduction of horizontal slab segments (Martinod et al. 2010). Second, the long-lived Andean margin has not been in a permanent state of compression, as envisaged by Uyeda and Kanamori’s model. On the contrary, it appears that tectonic shortening and Andean uplift occurred principally during three pulses in the upper Cretaceous, Eocene, and Neogene, whereas other periods are characterized by moderate shortening or extension (Martinod et al. 2010 and references therein). Third, despite rather uniform convergence conditions, the Andean margin is characterized by significant along-strike segmentation [e.g., Mpodozis and Ramos 1989; Martinod et al. 2010; Orts et al. 2012]. Maximum mean elevations, width, and crustal shortening in the Central Andes are much larger than those in the southern and northern parts [e.g., Martinod et al. 2010 and references therein]. Segmentation is also evident in the margin’s geologic record, as tectonic phases that caused Andean growth did not occur uniformly along the whole mountain chain [e.g., Jordan et al. 2001; Martinod et al. 2010 and references therein]. Fourth, although the Andean Cordillera is considered to have formed by crustal shortening through the advance of predominantly east-verging fold and thrust belts [Farias et al. 2010 and references therein], some areas are characterized by important margin-parallel strike-slip systems whose origin is debated [e.g., Cembrano et al. 2002; Thomson 2002].

The North Patagonian Andes [38°–43°S] are a poorly studied segment of the Andean Cordillera that present important differences compared with the Central Andes to the north. South of ~38°S, the broad and high (>4000 m) Central Andes narrow considerably, and maximum elevation descends to ~2000 m (Hervé 1994). The North Patagonian Andes have a crustal thickness of only ~40 km [Lüth et al. 2003], in contrast to the 70 km for the Central Andes [Allmendinger et al. 1997]. They are mainly composed of plutonic rocks of the Patagonian Batholith [Adriasola et al. 2006], show little deformation (Hervé 1994; Orts et al. 2012), and have a different timing of the main contractional phases than the Central Andes [García Morabito et al. 2011 and references therein]. A distinctive tectonic feature of the North Patagonian Andes is a ~1000-km-long, N-S-trending, dextral strike-slip fault system referred to as the Liquine-Ofqui Fault Zone [LOFZ; Hervé 1994]. This fault system accommodates part of the margin-parallel component of oblique subduction. However, the long-term nature, timing, and causes of lateral motion along this important fault system remain poorly understood. Also distinctive of this segment of the Andean margin is the subduction of the Chile Rise spreading center that defines the Chile triple junction between the Nazca, South American, and Antarctic plates. Subduction of the Chile Rise beneath the southernmost tip of the South American plate began at ~14 Ma and migrated north to its present location at ~46°S [Adriasola et al. 2006]. Finally, an important singularity of the geologic evolution of the North Patagonian Andes is the occurrence of late Cenozoic marine deposits across the entire forearc and in the western and eastern flanks of this range, indicating that uplift of this part of the Andes occurred rather recently [Levi et al. 1966; Ramos 1982; Encinas et al. 2012b].

Geologic studies of the North Patagonian Andes have been limited by a paucity of Mesozoic-Cenozoic strata. High exhumation rates during the late Cenozoic resulted in extensive erosion of the volcano-sedimentary cover [Adriasola et al. 2006], and what remains is mostly obscured by thick vegetation. As a consequence, there is little stratigraphic control to constrain the timing and onset of deformation. However, study of Cenozoic marine strata exposed on the western flank of this range at ~42°S (figs. 1, 2) can enable us to better understand the recent geologic evolution of the North Patagonian Andes. These marine strata, referred to as the Ayacara Formation [Levi et al. 1966], are strongly deformed and constitute a rhythmic succession of sandstone and siltstone. The age of this unit has been ascribed to the Eocene-Miocene [Levi et al. 1966; Rojas et al. 1994]. Late Eocene–Late Oligocene [Bourdillon, in SERNAGEOMIN 1995], and Miocene [Martínez-Pardo 1965]. Another unresolved issue is its possible correlation with the Neogene marine strata of the Chilean forearc, which intrigued the first geologists who studied this area [e.g., Katz 1965]. Hervé [1994], Rojas et al. [1994], and Hervé et al. [1995] speculated that the Ayacara Formation and correlative units to the south formed in short-lived pull-apart basins related to the dextral motion of the nearby LOFZ. As noted by Cembrano et al. (2002), this interpretation is inconclusive, as there is little structural and chronological evidence to support it.

In this study, we investigate the formation and subsequent destruction of a forearc basin at a particular segment of an ocean-continent convergent
Figure 1. Geologic map of the study area. Insets show the location of the studied sections detailed in figure 2. (After Ramos 1982; SEGEMAR 1995; Giacosa et al. 2005; Duhart 2008.) LOFZ = Liquiñe-Ofqui Fault Zone.
margin that has been influenced by strike-slip tectonics and the subduction of an active mid-oceanic ridge. For that purpose, we focus on the marine deposits of the Ayacara Formation that crop out on the western flank of the North Patagonian Andes. We integrate sedimentology, ichnology, foraminiferal micropaleontology, and laser ablation multicollector inductively coupled plasma mass spectrometry (LA-MC-ICPMS) U-Pb geochronology on detrital zircons to discern the age and sedimentary environment of this unit. Our findings have important implications for the geologic evolution of this segment of the Andean Cordillera, as they (1) indicate the onset of a major event of subsidence that affected the entire forearc and at least the western flank of the North Patagonian Andes and (2) provide a reliable maximum age for the uplift and deformation of the western flank of the range. On the basis of our findings and previous work in this region (fig. 1) are, from west to east, the Coastal Cordillera, the Intermediate Depression, and the Main Andean Cordillera. The Coastal Cordillera is a subdued mountain range located in the western part of Chiloé Island (Duhart and Adriasola 2008). The Intermediate Depression (also referred to as the Longitudinal Depression or the Central Valley) is a low-lying area between the Coastal Cordillera and the Main Andean Cordillera that includes eastern Chiloé Island and the adjacent floor of the Golfo de Ancud (Duhart and Adriasola 2008). The Main Andean Cordillera extends from the mainland coast to eastern Argentina and includes the highest peaks and active volcanoes (Duhart and Adriasola 2008).

The Coastal Cordillera (fig. 1) has a Paleozoic-Triassic metamorphic basement interpreted as a paleoaccretionary wedge (Duhart and Adriasola 2008). Locally, Eocene dacitic dikes and sills, as well as granodioritic stocks, intrude the basement (Duhart and Adriasola 2008). Coal-bearing rocks of the Caleta Chonos Formation (Eocene?-Oligocene?) occur in a small area of northwestern Chiloé (Antinao et al. 2000). Late Oligocene–Early Miocene volcanic and subvolcanic rocks of the Complejo Volcánica de Ancud and equivalent units are mostly exposed in northwestern Chiloé (Antinao et al. 2000). Younger Neogene marine strata of the Lacui Formation and equivalent units overlie the metamorphic basement or the volcanic rocks of the Com-

Figure 2. Detailed geologic maps showing sections from figure 3 (arrows) as well as the strike and dip of beds and inverse fault at Fiordo Pichicolo depicted in figures A3 and A4, available in the online edition or from the Journal of Geology office. LOFZ = Liquiñe-Ofqui Fault Zone.
The age of the Lacui Formation and correlative units is controversial. Molluscan biostratigraphy and Sr isotopes indicate a latest Oligocene–late Early Miocene age (Finger et al. 2007; Nielsen and Glodny 2009). Samples from different sections yielded three different assemblages of planktic foraminifera, indicative of the Burdigalian-Serravallian (Antinao et al. 2000), the Tortonian (Finger et al. 2007), and the Zanclean (Finger et al. 2007). According to these studies, this unit ranges from Early Miocene to Early Pliocene, but the younger markers reported in Finger et al. (2007) were misidentified, and most of the assemblages are probably Early Miocene. Benthic foraminifera indicate lower-bathyal depositional depths for the Lacui Formation (Finger et al. 2007). Shallow-marine Pliocene strata of the overlying Caleta Godoy Formation occur in some parts of northern Chiloé Island (Antinao et al. 2000). Pleistocene deposits, mostly glacial, and Holocene continental and marine sedimentary beds constitute the younger successions in the area.

The geology of the Longitudinal Depression (fig. 1 in this area is poorly known because most of this physiographic feature is submerged. Subsurface data from a 4010-m well drilled by ENAP (the Chilean National Oil Company) at Puerto Montt (Katz 1965; Elgueta et al. 2000) transected a succession comprising [1] a lower, 1480-m-thick interval of interbedded tuff and volcaniclastic breccia; [2] 310 m of Miocene marine rocks that probably correlate with the Lacui Formation, [3] 950 m of Pliocene continental? or marine? sedimentary rocks; and [4] 1300 m of Pleistocene continental and marine deposits mostly of glacial origin. Miocene marine strata and Pleistocene marine and continental deposits also occur in some areas along the coast of eastern Chiloé and on the nearby islands (Quiroz et al. 2004; Duhart and Adriasola 2008).

The oldest rocks of the Main Andean Cordillera (fig. 1) are Middle-Late Paleozoic metamorphics in the western part (Duhart 2008). Mesozoic-Cenozoic plutonics of the North Patagonian Batholith (Pankhurst et al. 1992) are the most extensive rocks in the Main Andean Cordillera. Jurassic, Cretaceous, Paleogene, and Neogene volcanic, volcaniclastic, and sedimentary rocks occur on the eastern and western flanks (Giacosa et al. 2005; Duhart 2008). In Chile, Miocene marine sedimentary rocks of the Ayacara and La Cascada Formations (Levi et al. 1966; Thiele et al. 1978; Encinas et al. 2012a) occur along the western and eastern parts, respectively. Miocene marine deposits of the Rio Foyel Formation occur in westernmost Argentina (Bertels 1980; Ramos 1982; Encinas et al. 2012a). Pleistocene and Holocene volcanic, volcaniclastic, and glacial deposits constitute the youngest successions in the North Patagonian Andes.

The LOFZ extends for ~1000 km along the North Patagonian Andes (figs. 1, 2) and decouples the south-central Chilean forearc, which behaves as a northward-gliding sliver, according to Forsythe and Nelson (1985). The long-term nature and timing of lateral motion along this important fault system, however, remain poorly understood (Thomson 2002). Structural analyses, Ar-Ar geochronology, and fission-track thermochronology along the Northern Patagonian Andes (41°–46°S) provide solid evidence for right-lateral transpression along the LOFZ during the Late Miocene to Pliocene (e.g., Cembrano et al. 2000, 2002; Thomson 2002; Adriasola et al. 2006). However, Cembrano et al. (2000) document a 1-km-wide mylonitic zone that shows microstructural evidence of sinistral reverse-displacement and is crosscut by a 100-Ma ±2 Ma undeformed porphyritic dike at Liqueñé (39°S). The LOFZ appears to be linked to a zone of magmatic weakening of the crust, as suggested by alignment of its strike with Pliocene-Recent volcanic centers (Hervé 1994; Glodny et al. 2008).

**Previous Studies of the Ayacara Formation**

The Ayacara Formation (Levi et al. 1966; figs. 1, 2) is a succession of turbidites composed of sandstone, siltstone, conglomerates, and minor tufts between the Estuario Reloncavi and Chaitén, with the most complete sections at Caleta Ayacara and Isla El Manzano (figs. 1, 2). Its scarce fossil fauna includes solitary corals (Solano 1978), echinoderms (Alarco´n Manzano 2000) document a 1-km-wide mylonitic zone that shows microstructural evidence of sinistral reverse-displacement and is crosscut by a 100-Ma ±2 Ma undeformed porphyritic dike at Liqueñé (39°S). The LOFZ appears to be linked to a zone of magmatic weakening of the crust, as suggested by alignment of its strike with Pliocene-Recent volcanic centers (Hervé 1994; Glodny et al. 2008).

Levi et al. (1966) assigned an Eocene-Miocene age to the Ayacara Formation on the basis of Martínez-Pardo’s study of foraminifers and silicoflagellates. Martínez-Pardo (1965) ascribed a Middle Miocene age to a foraminifer sample from Isla El Manzano. This age was supported by Fuenzalida (1979), who identified the planktic foraminifer *Orbulina universa* in the Ayacara Formation at Peninsula Huequi. Bourdillon (1994) criticized these studies because the foraminifera were identified only from petrographic thin sections, a methodology of limited accuracy. He later extracted foraminifers from several Ayacara Formation samples that were indicative of a Middle
Figure 3. Representative sections from Isla El Manzano (IMAN), Caleta Ayacara (CAYA), and Isla Ica (IICA; see fig. 3 for location). LM = lower member, UM = upper member. Diagonal lines in the scales represent long covered intervals. Grain size: (a) clay, (b) silt, (c) very fine sand, (d) fine sand, (e) medium sand, (f) coarse sand, (g) very coarse sand, (h) granule, (i) pebble, (j) cobble, (k) boulder. LT = lithology: (1) clast-supported conglomerate, (2) matrix-supported conglomerate, (3) breccia, (4) sandstone, (5) siltstone, (6) tuff, (7) andesitic dike, (8) covered interval. SS = sedimentary structures: (9) trough cross-bedding, (10) planar cross-bedding, (11) hummocky cross stratification, (12)
Eocene–Late Oligocene age [in SERNAGEOMIN 1995]. Radiometric $^{40}\text{Ar}/^{39}\text{Ar}$ dating of plagioclase from an Ayacara Formation tuff near the top of the Caleta Ayacara succession, however, yielded an age of 16.5 ± 0.5 Ma (F. Munizaga, written comm. in Rojas et al. 1994).

**Sampling Sites**

Because of the thick vegetation cover in the study area, Ayacara Formation exposures are limited to coastal cliffs and a few roadcuts and river valleys. The formation underlies Pleistocene deposits (Levi et al. 1966), but its basal contact has not been observed. We performed detailed sedimentologic and ichnologic analyses and measured stratigraphic thicknesses at Isla El Manzano, Caleta Ayacara, and Isla Ica (figs. 2, 3), where the unit is thickest and least weathered. In addition, we collected rock samples for foraminiferal analysis and U-Pb zircon geochronology.

This unit was divided into two members by Levi et al. [1966]. The lower member, 180 m thick, crops out only at the Caleta Ayacara section, whereas the upper member, up to 570 m thick, occurs in all of the studied localities (fig. 3). The transition between the two members at Caleta Ayacara is covered.

We also measured bedding attitudes in different outcrops (fig. 2). The average strike and dip of the beds are 335°/75° SW in the lower member and 325°/75° SW in the upper member at Caleta Ayacara, 335°/85° SW at Isla Ica (fig. A1, available in the online edition or from the Journal of Geology office), and 35°/40°–70° NW at Isla El Manzano. Strata of the coastal outcrops of Fiordo Pichicolo average 360°/66° W but become vertical to the east (fig. A2, available in the online edition or from the Journal of Geology office). In a newly opened roadcut that runs parallel to this fjord, we measured a 270°/75° W reverse fault but did not observe slickenslides. The fault has associated tight folds and a 30-cm-thick tectonic breccia (figs. A3, A4, available in the online edition or from the Journal of Geology office).

Ayacara Formation contacts with other geologic units are obscured by dense vegetation. However, the intense deformation of this unit in some localities, as well as the occurrence of exhumed Miocene granites in this region, suggest that some of these contacts are inverse faults.

**Sedimentary Facies of the Lower Member**

**Matrix-Supported Conglomerate (Cgl1).** This facies consists of sharp- to erosively based massive, matrix-supported conglomerate (fig. 4A) that forms meters-thick successions. It is poorly sorted and contains subangular to subrounded clasts composed primarily of volcanic rocks [andesite with aphanitic and trachytic textures, tuff, and minor dacite] and secondarily of plutonic rocks [diorite and tonalite] and sandstone. Clasts show random orientations and range in size from millimeters to several decimeters. This facies locally grades upward into the clast-supported conglomerate of facies Cgl2.

This facies is interpreted as debris-flow deposits on the basis of their poor sorting, randomly oriented floating clasts, and abundance of matrix (e.g., Nemec and Steel 1984; Hwang and Chough 1990). **Clast-Supported Conglomerate (Cgl2).** This facies consists of slightly erosive to, more rarely, sharp-based clast-supported conglomerate that generally has a lot of sandy matrix. Beds range in thickness from decimeters to, more commonly, meters. Conglomerate is generally well to moderately sorted. Clasts range in size from a few millimeters to several centimeters, albeit outsized boulders up to 80 cm occur locally. Clasts are subangular to rounded and compositionally similar to those of facies Cgl1. Some beds show abundant rip-up sandstone clasts up to 40 cm across. Clasts are generally arranged parallel to bedding, and some are imbricated. Beds are generally massive or show crude stratification. However, some strata show normal or inverse grading and, more rarely, planar cross-bedding (fig. 4B). Conglomerates in this facies sometimes grade into sandstones of facies Ls.

Figure 4. Characteristic facies and trace fossils of the lower member of the Ayacara Formation at Caleta Ayacara section. A, Matrix-supported, poorly sorted conglomerate (facies Cgl1). B, Clast-supported conglomerate (facies Cgl2). The bed indicated by the hammer shows planar cross-bedding. The upper bed is sandstone. C, Coarsening-upward parasequence formed by a basal sandy-siltstone interval that grades into sandstone to the top [facies Ls]. The base (flooding surface) is indicated by the arrow. D, Hummocky cross stratified (HCS) sandstone (facies Ls). Note the low-angle curved intersections of stratification characteristic of HCS, indicated by the pen. E, Sandstone with symmetrical wave ripples (facies Ls). F, Undetermined radially branching burrows in sandy siltstones from the basal interval of the parasequence shown in C.
grading illustrates that clasts within the flow were free to move relative to each other, suggesting little or no cohesion (Van der Straaten 1990). Planar cross-bedding represents 2-D dune migration. These sedimentary features indicate transport by fluidal flows. The common abundance of sand matrix and outsized clasts suggests that transport by hyperconcentrated flows was the dominant process (Iverson 1997).

**Laminated Sandstone (Ls).** This facies consists principally of laminated sandstone interbedded with siltstone or conglomerate forming two different associations. Facies association Ls1 consists of meters-thick, coarsening-upward successions characterized by a basal sandy-siltstone interval that grades upward into sandstone (fig. 4C). Siltstone is dark gray and in sharp contact with the underlying rock. Sandstone is fine grained to, more commonly, medium to coarse grained and usually shows hummocky cross stratification (fig. 4D) and scarce intercalations of conglomerate a few centimeters thick. Facies association Ls2 comprises a medium- to very coarse-grained sandstone interbedded with conglomerate of facies Cgl2. Some beds have an erosional basal contact overlain by a conglomerate that grades into sandstone. Strata are typically tabular and decimeters to a few meters thick. Some sandstone beds show symmetrical ripples (fig. 4E), while others display trough or planar cross-bedding. Sandstone is composed primarily of quartz, andesite lithics, and plagioclase and secondarily of clinopiroxene, although some beds consist of shards, pumice, and quartz.

Sandstone in this facies shows sedimentary structures that indicate wave reworking. Facies association Ls1 is interpreted as shallow-upward parasequences, where basal siltstones represent shelf deposition and hummocky cross stratified sandstone indicates reworking by storm waves in the lower shoreface (Clifton 2006). Thin intercalations of conglomerates are interpreted as storm deposits. Laminated sandstone in facies association Ls2 reflects wave influence. Sandstones showing symmetrical ripples are typical of the lower shoreface, whereas those with cross-bedding indicate deposition in the upper shoreface (Clifton 2006).

The Ayacara Formation’s lower member presents an overall fining- and thinning-upward succession, where facies Cgl1 (matrix-supported conglomerates) is present only in the lower part. Facies Cgl2 (clast-supported conglomerates) occurs along the whole interval but is more abundant and thicker in the lower part, and facies Ls (laminated sandstone and minor siltstone) is present only in the upper half of the succession.

**Sedimentary Facies of the Upper Member**

**Rhythmically Interbedded Sandstone and Siltstone (T).** This facies consists of fine- to very coarse-grained sandstone rhythmically interbedded with siltstone (fig. 5A). Bed thickness is variable, but sandstone beds are typically decimeters to meters thick, whereas siltstone intervals are typically centimeters to decimeters thick. They form partial Bouma cycles where sandstone predominates and is locally amalgamated. They typically show Tbc divisions [parallel-convolute or, more rarely, ripple cross-lamination, fig. 5B], but Tab [structureless-parallel-laminated], Ta, and Tb are also common. Sandstone has sharp and, more rarely, erosive basal contacts and flute casts (fig. 5C) and occasionally contains ripped-up mudstone clasts up to some decimeters in size. Sandstone beds commonly present convolute lamination, locally overturned, in intervals up to 1 m thick. Fine-grained (Tde) intervals consist of siltstone and, more rarely, tuff. They are generally azoic, but some of the tuffaceous siltstone contains abundant foraminifers. Flames, load casts, and, more rarely, ball and pillow structures locally occur at the contact between sandstone and siltstone beds. Sandstone is primarily composed of plagioclase, quartz, and volcanic lithics and secondarily of shards, pumice, and igneous lithics. Interbedded sandstone and siltstone (T) and conglomerate (Cgu) are the most common facies in the upper member of the Ayacara Formation.

The rhythmically interbedded sandstone- and siltstone-forming Bouma cycles indicate that this facies represents classic turbidites (Bouma 1962; Posamentier and Walker 2006). The predominance of basal Bouma divisions (Ta, Tb), as well as the coarse grain size and thick bedding, suggests proximal turbidites. The common presence of convolute lamination, as well as the occurrence of flames, load casts, and ball and pillow structures, indicates rapid sediment deposition that causes trapping of pore fluid and deformation of primary structures (Posamentier and Walker 2006). The presence of rip-up clasts and flute casts in some beds are evidence of turbulent erosion of the substrate immediately prior to deposition.

**Conglomerate (Cgu).** Individual conglomerate beds are typically decimeters to meters thick and are usually interbedded with turbidites, sandstones, or siltstones. Basal contacts are sharp or slightly erosional and sometimes show load and flame structures (fig. 5D), pseudonodules, descendant clastic dikes, or disruption of underlying siltstone beds. Conglomerate is clast supported, is moderately well sorted, and contains subangular...
Figure 5. Characteristic facies and trace fossils of the upper member of the Ayacara Formation. A, Rhythmically interbedded sandstone and siltstone (facies T) at Isla Ica. A human for scale is encircled. B, Sandstone showing partial Tbc (parallel-convolute lamination) Bouma cycle (facies T) at Caleta Ayacara. C, Base of turbidite with flute casts (facies T) at Isla Ica. Paleocurrent direction (157°) is indicated by the arrow. D, Flame formed by tuffaceous siltstone injecting into granule conglomerate (facies Cgu) at Isla El Manzano section. E, Slumped tuff beds (facies Slp) at Caleta Ayacara section. F, Chondrites isp. in very fine-grained sandstone at Isla Ica section. Note the typical branching of this trace fossil.
to rounded clasts, typically a few centimeters in size. Clasts are mainly aphanitic and trachytic andesite and minor sedimentary and metamorphic lithics. They present a sandy matrix composed of angular granules of plagioclase and quartz, some of them showing embayments. Conglomerates occasionally contain rip-up clasts up to a few decimeters in size that consist of fragments of stratified siltstones, tuffs, or interbedded siltstone and sandstone, some of which display folding. Some beds, however, are composed entirely of angular rip-up clasts. On the basis of Walker’s [1975] classification, we distinguished three different types of conglomerates: the Cgu1 type is normally graded with subtle stratification, generally grading upward into pebbly sandstone and rarely showing planar cross-bedding; the Cgu2 type is normally graded and not stratified; and the Cgu3 type is inversely to normally graded. Conglomerate Cgu2 is predominant, Cgu1 is less common, and Cgu3 is rare. Normal grading in Cgu1 and Cgu2 is generally very subtle and may be absent.

The presence of grading and, more rarely, cross-bedding in this facies suggest transport by turbulent flows in which grains move freely relative to one another. Conglomerates in deep-marine successions that show similar facies to those described above are interpreted as having been deposited by turbidity currents [Posamentier and Walker 2006]. The presence of rip-up clasts is evidence of turbulent erosion of the substrate immediately prior to deposition, whereas the occurrence of disruptive sedimentary structures, such as flames and load casts, indicates rapid deposition.

**Structureless Sandstone (Ss).** Structureless sandstones are medium- to very coarse-grained, usually massive sandstones, although they may locally display zones of indistinct parallel lamination, and some beds show dish structures. Structureless sandstone successions are decimeters to meters thick and locally form amalgamated stacks that have erosional contacts or rip-up clasts occasionally folded up to 1 m in size between individual beds. Mineralogical composition of this facies is similar to that of sandstones in facies T.

Structureless sandstones are commonly interpreted as deposited by turbidity currents on the basis of the common association of this facies with classic turbidites [Posamentier and Walker 2006]. The presence of dish structures that indicate initial deposition of a fluid-rich sediment-water mixture also suggests this mode of deposition.

**Siltstone and Very Fine-Grained Sandstone (Sif).** This facies consists of dark gray to brownish siltstone and very fine-grained sandstone forming intervals up to several meters thick. Locally it shows thin intercalations of coarse-grained sandstone and granule conglomerate. Bedding thickness ranges from centimeters to decimeters. Beds are massive or show poorly defined lamination. Fossils of bivalves, gastropods, scaphopods, echinoderms, bryozoans, and solitary corals are, with few exceptions, present only in this facies of the Ayacara Formation. Foraminifers, mostly planktic, are abundant, and pyrite is common.

This facies is interpreted as having been deposited by distal turbidity currents and hemipelagic fallout. Interbedded sandstone and conglomerate are coarser-grained turbidites. Thick, fine-grained intervals within turbiditic successions are usually interpreted as condensed sections deposited at extremely low sedimentation rates during maximum regional transgression [Weimer and Slatt 2006]. The abundant and diverse biota, the high ratio of planktic to benthic foraminifers, the high content of organic matter suggested by the dark color of this facies, and the presence of pyrite are also characteristic features of condensed sections [Weimer and Slatt 2006].

**Tuff (Tf).** This facies consists principally of ash and secondarily of lapilli tuff and is commonly interbedded with sandstone and conglomerate. Beds typically show a green color and are commonly decimeters thick, but at the Caleta Ayacara section there are intervals up to 18 m thick. Tuff consists mostly of shards and subordinate pumice, quartz, and plagioclase [fig. A5A, A5B, available in the online edition or from the *Journal of Geology* office]. Shards are generally transformed to zeolites. This facies usually shows a massive aspect, but parallel lamination, convolute bedding, and, more rarely, slumps are present in some beds. Siltstone rip-up clasts up to 80 cm across, some of which are deformed, are present locally.

This facies was initially formed by explosive volcanism (ashfalls and pyroclastic flows), as indicated by the abundance of shards and pumice clasts. Fine-grained volcanic material entered water and was deposited by settling. Coarser-grained pyroclastic flows were transformed into water-supported turbidity flows that displaced them farther and deeper into the marine basin, as suggested by the presence of parallel and convolute laminations [see Cas and Wright 1991].

**Slumped Sandstone and Siltstone (Slp).** This facies only rarely occurs and consists of folded interbedded sandstone and siltstone forming intervals decimeters to a few meters thick. These beds are underlain and overlain by flat-bedded successions [fig. 5E].
This facies was formed by sliding and folding of unconsolidated turbidites. They were previously deposited in the basin and subsequently displaced downslope.

**Trace Fossils**

Trace fossils in the lower member of the Ayacara Formation are indeterminate radially branching burrows (fig. 4F) present only in siltstone and very fine-grained sandstone. Bioturbation in the upper member is scarce and present only in discrete intervals (<5% of beds are bioturbated). Two distinct trace-fossil assemblages can be differentiated. Trace-fossil assemblage 1 occurs in siltstone and, more rarely, very fine-grained sandstone of facies T and Sil. It is dominated by *Chondrites* isp. (fig. 5E), with *Phycosiphon incertum* and *Planolites montanus* accessory elements and *Zoophycos* isp., *Schaubcylindrichnus freyi*, and *Taenidium* isp. rare components. The bioturbation index (BI scale by Taylor and Goldring 1993) is typically low (1–2), although it reaches 5 in some intervals, which are also characterized by an increase in ichnodiversity. This assemblage is ascribed to the *Zoophycos* ichnofacies, characterized by low oxygen levels associated with high organic detritus in quiet-water settings, and is typical of outer shelf to slope settings (Frey and Pemberton 1984). In the case of the Ayacara assemblage, it essentially characterizes interturbidite intervals of colonization.

Trace-fossil assemblage 2 occurs in medium- to coarse-grained sandstone of facies T and Ss and, more rarely, in tuff (facies Tf). It is dominated by *Thalassinoides suevicus* and *Ophiomorpha* isp., each of which commonly forms a monospecific suite characterized by extensive horizontal networks. *Palaeophycus tubularis* is locally present. Although typically low, the bioturbation index is somewhat variable, reaching BI 4 at the network levels. This trace-fossil assemblage characterizes the *Skolithos* ichnofacies, typical of high-energy shallow-marine environments but also widely reported in deepwater settings, and it reflects local environmental conditions, such as high energy, a sandy substrate, high levels of oxygenation, and an abundance of suspended organic particles (Buatois and Lópeza-Angriman 1992). The low bioturbation index of the deposits, the low trace-fossil diversity of both associations, and the predominance of morphologically simple structures probably reflect short colonization windows in a stressful environment characterized by rapid rates of volcaniclastic sedimentation (see also Mángano and Buatois 1997).

**Foraminiferal Biostratigraphy and Paleodepths**

To determine the age and paleobathymetry of the Ayacara Formation, we analyzed nine samples from the Caleta Ayacara and Isla El Manzano sections. Details about the extraction method are provided in the appendix, available in the online edition or from the *Journal of Geology* office. All of the isolated specimens show some degree of diagenetic alteration that masks diagnostic features. Thus, species identifications (table S1, available in the online edition or from the *Journal of Geology* office) were based mostly on experience and intuition stemming from familiarity with the Miocene fauna of south-central Chile.

Planktic foraminifera are generally more abundant than benthic specimens in the analyzed samples, reaching P : B ratios of 90–100 in some beds of the Caleta Ayacara section, where planktics are so profuse that they constitute as much as 75% of the rock volume. Only five samples, all from the Caleta Ayacara section, yielded planktic specimens that could be identified at the species level (tables S1, S2, available in the online edition or from the *Journal of Geology* office). Samples CAYA 17, 18, 19, 20, and 22 yielded *Paragloborotalia mayeri*, a Late Oligocene–Middle Miocene species with a zonal range of P22–N14. CAYA 22 also yielded *Globigerinoides quadrilobatus*, and the two species have a concurrent range of N6–N14 (Burdigalian–Serravallian). Finally, the occurrence of *Globigerinoides sicamis* in CAYA 17 restricts the sample to zones N8–N9 (Langhian, Middle Miocene). This determination narrows the range for four stratigraphically higher samples to N8–N14 (Langhian–Serravallian, Middle Miocene).

Benthic foraminifera were present in four samples from the Caleta Ayacara section and in three samples from the Isla El Manzano sections (table S1), but all of the benthic assemblages were weak (i.e., consisting of relatively few specimens). Depositional paleodepths are derived from the upper-depth limits (UDLs) of benthic foraminifera in the Chilean Miocene, which are based on the depth distributions of similar taxa currently living along the Pacific margin of central South America, as recorded by Bandy and Rodolfo (1964), Ingle et al. (1980), and Resig (1981). Most of the studied assemblages are mixed-depth associations of shelf and slope species, indicating downslope displacement (table S1). Five samples (CAYA 10, 18, 22; IMAN
16, 30) yielded foraminifera indicative of upper-middle bathyal (500–1500 m) depths—Cyclammina incisa (Stache), Gyroidinoides nipponica (Ishizaki), Hansenisca sp., and Neouvigerina hispida (Schwager)—some of which are included in the cosmopolitan deepwater fauna of Van Morkhoven et al. (1986). One sample (CAYA 10) also had Siphonodosaria sagrinensis [Bagg], which occurs no shallower than lower-middle bathyal (1500–2000 m) depths. In addition, a fragment possibly of that species was found in IMAN 30. Considering the poor preservation and weak benthic assemblages, there can be little doubt that these occasional specimens represent a much more diverse fauna that most likely was deposited below 1500 m.

**U-Pb Geochronology**

Five samples, each consisting of ~10 kg of medium- to coarse-grained sandstones, were selected for U-Pb zircon geochronology, which was conducted by LA-MC-ICPMS at the Arizona LaserChron Center following the techniques described by Gehrels et al. (2008) and detailed in the appendix. Two of the five samples were from the Isla El Manzano section, and three samples were from the Caleta Ayacara section [table S3; fig. 6]. Four samples correspond to the upper member of the Ayacara Formation (CAYA 14, CAYA 24, IMAN 14, and IMAN 20), and one sample (CAYA 4) belongs to the lower member of this unit. Crystals in these samples are predominantly prismatic and have a clear pinkish color except for a few reddish, subrounded grains in sample CAYA 4. The U/Th ratio is <7 in the grains from the Isla El Manzano and Caleta Ayacara stratigraphic sections; it indicates that most of the zircons are derived from igneous rocks [Rubatto 2002].

The majority of detrital zircons from the Ayacara Formation’s upper member yield ages near 16 and 23 Ma (fig. 6) and show mostly a unimodal distribution in their relative probability age spectra. The sample from the lower member similarly shows a young age peak, but with the addition of much older age peaks [Cretaceous ~121–131 Ma, Carboniferous ~303 Ma, and Proterozoic 1100–1200 Ma]. The maximum depositional ages for the full data set range from ~17.6 to 21.8 Ma [fig. 6].

**Discussion**

**Age of the Ayacara Formation and Paleogeographic Considerations.** Planktic foraminifera studied here indicate a Langhian (early Middle Miocene) age for at least part of the upper member of the Ayacara Formation [tables S1, S2]. Accordingly, U-Pb geochronology shows that the youngest detrital zircon populations from both the lower and the upper members indicate an Early Miocene maximum age [table S3; fig. 6]. Considering that Ayacara Formation strata contain abundant volcanic material, it is probable that sedimentation was contemporaneous with nearby volcanism, as previously suggested by Levi et al. (1966) and Rojas et al. (1994). This implies that the younger zircon ages are likely very close to the age of deposition. Thus, according to the aforementioned data, the age of the Ayacara Formation is Early-Middle Miocene. However, there is no complete section of this unit, so it could have a slightly wider age range. A minimum age for the formation can probably be interpolated from two radiometric ages (206Pb/238U in zircons) of 13.3 ± 0.2 Ma and 8.3 ± 0.4 Ma from dioritic porphyries and tonalite plutons at the Fiordo Pichico [Acosta and the Estuario Relloncaví, respectively (figs. 1, 2), which were correlated with dikes intruding the Ayacara Formation (Duhart 2008). These ages must be taken with some caution, as the analyzed samples were obtained from localities where crosscutting relationships between igneous and sedimentary rocks were not observed. Nevertheless, they are in agreement with fission-track data [Adriasola et al. 2006] and structural analysis (Cembrano et al. 2000), indicating that significant uplift and exhumation commenced in this area during the Late Miocene and likely put an end to deposition of the Ayacara Formation. Thus, a minimum age of late Middle-Late Miocene is inferred for the Ayacara Formation.

The Early-Middle Miocene age for the Ayacara Formation allows for some important paleogeographic inferences. According to these data, the Ayacara Formation correlates with at least part of the deep-marine Miocene strata of the Lacuí Formation exposed on Chiloé Island (see previous comment on the age of this unit; fig. 1) and equivalent units that crop out along the forearc of south-central Chile (see Encinas et al. 2008, 2012b and references therein), and probably the Traiguén Formation to the south [Hervé et al. 2001]. Subsurface data from the Puerto Montt 1 well in the Longitudinal Valley [Elgueta et al. 2000] indicate that below ~2000 m of Plio-Pleistocene deposits there is a Miocene marine succession that probably correlates with the Lacuí and Ayacara Formations. The Miocene section overlies 1450 m of volcaniclastic rocks that may also correlate with the Ayacara Formation [Katz 1965; Levi et al. 1966]. The likely source area, a contemporaneous volcanic arc, was
Figure 6. Relative probability age distribution plots of U-Pb ages (Ma) of detrital zircon from samples from the Caleta Ayacara (CAYA) and Isla El Manzano (IMAN) sections.

likely close to the depositional basin, as suggested by the abundance of conglomerate and coarse-grained sandstone in the Ayacara Formation. As yet, no Miocene volcanics that could correlate with the marine unit have been recognized in its vicinity. However, Early Miocene–Pliocene plutonic rocks are abundant in the North Patagonian Andes (fig. 1; Duhart 2008) because high exhumation rates that commenced in the Late Miocene led to the erosion of the Miocene volcanics and exposure of the plutonic basement [see below].

U-Pb zircon geochronology can also give us important information regarding paleogeographic reconstructions of the study area during deposition of the Ayacara Formation. Provenance inferences, however, are hampered by the high exhumation
rates that affected the North Patagonian Andes from the Late Miocene onward and gave way to the erosion of the majority of the Meso-Cenozoic volcano-sedimentary cover rocks. Interestingly, there are marked differences in detrital zircon ages between sample CAYA 4, from the lower shallow-marine basal member of the Ayacara Formation, and samples IMAN 14, IMAN 20, CAYA 14, and CAYA 24, all from the upper deep-marine member (table S3, fig. 6). Sample CAYA 4 contains zircons of different ages that include latest Oligocene–Middle Miocene (major group between ~17 and 24 Ma), Early Cretaceous (~121–132 Ma), Early Jurassic (~180 Ma), Carboniferous (~303 Ma), Cambrian (~526 Ma), and Proterozoic (~1100–1200 Ma) grain populations. Carboniferous, Cambrian, and Proterozoic igneous rocks are unknown in the area, but Duhart et al. (2009) found Carboniferous and Proterozoic detrital zircon populations in metamorphic rocks of the Patagonian Andes, which suggests that they are recycled. Jurassic and Cretaceous plutonic and volcanic rocks occur only in the Main Andean Cordillera, the latter solely exposed east of the study area (Duhart 2008; fig. 1). Upper Oligocene–Middle Miocene igneous rocks are abundant in the Main Andean Cordillera, although some small outcrops of Upper Oligocene–Early Miocene volcanic and subvolcanic rocks occur in the Coastal Cordillera (fig. 1). Overall, these data suggest a source area formed by rocks of different ages and located in the Main Andean Cordillera, east of the Ayacara Formation depositional basin.

In marked contrast, samples IMAN 14, IMAN 20, CAYA 14, and CAYA 24, which are from two different sections and belong to a stratigraphically younger interval, contain exclusively Late Oligocene–Early Miocene zircon populations (table S3, fig. 6). A possible explanation for this marked change is that the Miocene transgression that caused deposition of the deep-marine upper member of the Ayacara Formation completely covered the former source area to the east and that sediment provenance was limited to explosive volcanism of a submerged volcanic arc (see Levi et al. 1966). However, most of the conglomerate clasts in the Ayacara Formation are subrounded, indicating some degree of fluvial or alluvial transport (or wave reworking) prior to their marine deposition. In addition, although most clasts are andesitic, they vary in texture, which suggests that they did not come from a single source but rather from the erosion of different parts of the volcanic arc. If this volcanism had been submarine, deposition of angular, compositionally and texturally similar clasts would be expected. Other possibilities for the single Late Oligocene–Early Miocene detrital zircon population is that the entire area was covered by the sea except for an emerged volcanic arc or that an important episode of magmatism created large volcanic ediﬁces that blocked river channels draining Mesozoic-Paleozoic terrains. Another possible scenario is that a signiﬁcant ﬂux of Cenozoic zircons simply diminished the older zircon signal. However, the high exhumation rates that affected the study area during the Neogene have given way to a very different geologic conﬁguration between the Miocene and the present day; thus, all of these possible explanations are speculative.

An important issue regarding paleogeographic and tectonic reconstructions in the study area during the Miocene is the relationship between the Ayacara Formation in the western Main Andean Cordillera and the Cenozoic marine deposits of the Rio Foyel and La Cascada Formations that crop out only ~70 km east in the eastern Andes (ﬁg. 1). Ramos (1982) related the Rio Foyel Formation and equivalent deposits (including the La Cascada Formation) to a Paciﬁc transgression and correlated these strata with the Navidad-equivalent deposits along Chile’s Paciﬁc coast. However, the uncertainty over the age of these marine deposits has complicated their correlation. The Rio Foyel Formation was ascribed to the Eocene on the basis of molluscs (Chiesa and Camacho 2001), to the Early Oligocene on the basis of 87Sr/86Sr (Griffin et al. 2004) and K/Ar dating (minimum age on crosscutting dikes by Giacosa et al. 2001), to the early Middle Oligocene–Early Miocene on the basis of foraminifers (Bertels 1980), and to the Late Oligocene–Early Miocene on the basis of palynomorphs (Barreda et al. 2003) and foraminifers (Malumíán et al. 2008). The molluscan fauna of the La Cascada Formation, on the other hand, has affinities with the Eocene and Miocene faunas of Chile and Argentina according to Thiele et al. (1978), but their poor preservation renders their identiﬁcation at the species level doubtful. Recently, Encinas et al. (2012a) carried out LA-MC-ICPMS U-Pb geochronology on Río Foyel and La Cascada detrital zircons and obtained four maximum ages between ~22 and ~17 Ma. In addition, recent studies of calcareous nannofossils from the Río Foyel Formation indicated an Early-Middle Miocene age (Bechis et al. 2012). These results imply a temporal correlation between the marine deposits from the eastern and western flanks of the North Patagonian Andes. However, it is unclear whether both successions were deposited during the same marine transgression, as paleontologic evidence is contradictory. On the basis of foraminifera and molluscs, Bertels
mann et al. (2011) and Ramos (1982) ascribed a Pa-
succession at Cerro Plataforma. However, Feld-
origin for the Rı´o Foyel Formation and a correlative
(1980) and Griffin et al. (2002) favored an Atlantic

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served fossils, is needed to resolve this issue.
served, which hinders their identification and
nificant problem is that most fossils from the east-
noi gastropod of Atlantic origin,
ern part of the Andean Cordillera are poorly pre-
ginally, Nielsen et al. (2005) proposed an Atlantic-
sion culminating in deepwater deposition of its up-

Possible Causes of Basin Subsidence. The sedimen-
tologic, ichnologic, and micropaleontologic data
this study allow reconstruction of the deposi-
tional environment in which the Ayacara Forma-
tion accumulated. The basal member is interpreted
as a fan-delta succession in which conglomeratic fa-
cies were deposited by episodic debris and hyper-
concentrated flows and laminated sandstone reflects
wave reworking during “normal” marine sedimenta-
tion. The upper member is interpreted as a deep-
marine succession where intermittent turbidity flow deposits alternate with hemipelagic sedimenta-
tion. The presence of lower-middle bathyal ben-
thetic foraminifera in mixed-depth assemblages, as
well as the high planktic-to-benthic ratios, further
supports this interpretation (e.g., Van der Zwaan
1999). The occurrence of the Zoophycos ichnofacies
in fine-grained strata is also consistent with the mi-
cropaleontologic data (Frey and Pemberton 1984).
An abundance of gravel and sandstone versus mud
in these deposits, the predominance of basal Bouma
divisions, and the common occurrence of convolute
lamination in the turbiditic facies are characteristic
of tectonically active marine basins with narrow
shelves and high sedimentation rates (see Reading
and Richards 1994). Most of the sediment was likely
shed into the basin from a nearby volcanic arc, as
indicated by the predominance of volcanic clasts and
minerals and the abundance of interbedded tuffs.
Gravel and sand were probably transported short dis-
tances to the marine basin via alluvial fans, as in-
dicated by the subrounded clasts. Deposition of wa-
terlain tuffs would have resulted from ashfalls or
pyroclastic flows during or immediately following
explosive volcanism.

The Ayacara Formation represents a transgres-
sion culminating in deepwater deposition of its up-
per member. Because this member yielded some
species of benthic foraminifera that indicate lower-
middle bathyal depths (table S1) and maximum eu-
static sea-level changes during the Cenozoic are on
the order of a few hundred meters (e.g., Haq et al.
1987), tectonic subsidence must have been the pri-
mary cause of this transgression. Forsythe and Nel-
son (1985) hypothesized that the Cenozoic Golfo
de Penas-Tatua basin, farther south near the Chile
Triple Junction (46°–48°S), could have formed as a
pull-apart basin in response to the northward dis-
placement of the Chiloé block along the LOFZ.
This prompted some authors to speculate that the
Ayacara Formation and other supposedly correla-
tive units between 42°S and 47°S were deposited in
a similar tectonic context (Pankhurst et al. 1992;
Hervé 1994; Rojas et al. 1994; Hervé et al. 1995),
but they provided little structural or chronological
 evidence that relates dextral strike-slip motion on
the fault to coeval downwarping of these basins and
subsequent sedimentation therein (see Cembrano
et al. 2002). More importantly, correlation of the
Ayacara Formation with marine Miocene deposits
on Chiloé Island and in the Central Valley at 42°S
indicates that subsidence affected a wide area, ~100
km from the present west coast of Chiloé to the
western Main Andean Cordillera. Furthermore,
Miocene subsidence not only affected the margin
at this latitude but almost the entire length of the
Chilean forearc (see Encinas et al. 2008). Therefore,
Miocene subsidence along the Chilean margin
must have been caused by a more regional event
than localized activity along the LOFZ or other tec-
tonic discontinuities. Subsidence cannot be related
to the load imposed by a growing accretionary
wedge, as the actual prism in south-central Chile
is too small and probably formed within the last 2
m.yr. (Bangs and Cande 1997). In addition, subsi-
dence in the study region affected an area many
tens of kilometers landward of the prism where the
upper plate is unaffected by accretionary loading.
Muñoz et al. (2000) and Jordan et al. (2001) de-
scribed an episode of moderate extension that af-
fected the Chilean margin between ~33°S and 43°S
during the Late Oligocene–Early Miocene and re-
sulted in extensional basins filled with thick suc-
cessions of volcanic and sedimentary rocks from
the present Chilean coast to the retroarc. Muñoz
et al. (2000) ascribed extension to negative rollback
of the subducting slab caused by the transient
steepening of the subduction angle that they related
to a change from slower, more oblique South
America–Farallón convergence to more rapid, near-
normal South America–Nazca convergence at ~25–
24 Ma (e.g., Pardo-Casas and Molnar 1987; Somoza
1998). Muñoz et al. (2000) also related forearc Mio-
ocene marine sedimentation in south-central Chile to this extensional event. However, their assumption is based principally on the idea that Miocene marine rocks on northwestern Chiloé Island (fig. 1) are interbedded with Late Oligocene–Early Miocene volcanic rocks of the Complejo Volcánico de Ancud, but García (1968) and Encinas et al. (2009) demonstrated that these marine rocks unconformably overlie the volcanics and are therefore younger. In addition, most data indicate that extension ceased in this region at ∼20 Ma and was followed by contractional basin inversion (Munoz et al. 2000; Jordan et al. 2001; Kay et al. 2006). Although a small overlap exists, U-Pb and foraminiferal ages obtained in this study indicate that the Ayacara Formation commenced deposition coincident with the end of the regional extensional event. Some uncertainty remains, because a few radiometric ages for some of the volcano-sedimentary successions believed to have been deposited under extensional tectonics are younger than 20 Ma [see Flynn et al. 2008 and references therein].

On the basis of Von Huene and Scholl’s (1991) study of convergent margins, Encinas et al. (2008, 2012b) related deep-marine Miocene sedimentation and major forearc subsidence in south-central Chile to subduction erosion, where the subducting plate erodes the underside of the overriding plate, causing its thinning and margin subsidence. Melnick and Echtler (2006) proposed that a low-relief Andes resulted in a sediment-starved trench that, in addition to high plate-convergence rates, triggered subduction erosion of the south-central Chilean margin. Subsidence affected the entire width of the forearc south of 38° as well as the western flank of the North Patagonian Andes at ∼42°S, in contrast to other regions of Chile, where it was restricted to the present coastal area. Yet, subduction erosion–driven subsidence is normally described for offshore and coastal regions (Clift and Vannucchi 2004). South-central Chile’s anomalously thinned crust, which was probably caused by the Oligocene–Early Miocene extensional tectonic event (Jordan et al. 2001), could have facilitated tectonic erosion–driven subsidence in this area (Encinas et al. 2012b). Flattening of the slab, on the other hand, has been proposed to cause basal erosion in forearc areas from the trench to the depth of magma generation [see Kukowski et al. 2006 and references therein]. Munoz et al. [2000] signaled that transient steepening of the subduction angle could have caused negative rollback in the south-central Chilean margin at ∼25 Ma. We speculate that subsequent slab flattening in combination with a sediment-starved trench, high convergence rates, and an anomalously thinned crust could have caused tectonic erosion and the related subsidence of the entire forearc and western flank of the present North Patagonian Andes.

On the basis of studies of the northeast Japan margin and of numerical models of the interaction between the subducting plate and the overriding continent, Regalla et al. (2011) and Martinod et al. (2012) independently proposed that increasing convergence rates can induce changes in slab geometry and drive upper-plate subsidence. Considering that a threefold increase in trench-normal convergence rate between the Nazca and South American plates took place at ∼25 Ma, this could be another cause of forearc subsidence.

In summary, we consider subduction erosion to be the most probable cause of forearc subsidence in the North Patagonian Andes. We do not discard as possible causes the extensional event described by Munoz et al. (2000) and Jordan et al. (2001) or phenomena related to the interaction between the oceanic and continental plates, such as changes in the convergence rate. Additional research is needed to better understand the vertical motion of forearcs in convergent margins.

**Implications for Andean Deformation.** Ayacara Formation marine strata deposited at bathyal depths during the Early-Middle Miocene are tilted, folded, faulted, and exposed in the western flank of the North Patagonian Andes. Thus, after deposition the unit uplifted, emerged, and contracted. Considering that the Ayacara Formation was deposited in a deep-marine environment, Early-Middle Miocene is the maximum age for the deformation and uplift of the western part of the Main Andean Cordillera in the study area. This implies that at least the western part of the mountain chain formed rather recently, which is a significant contribution of this study. Our data agree with other studies of the tectonic evolution of the Andean Cordillera in the area ∼41°–42°S, although some uncertainties remain. The Early Miocene age (∼22–17 Ma) of the Rio Foyel marine deposits and equivalent units (Encinas et al. 2012a) is also the maximum age for the uplift and tectonic deformation of the eastern part of the Main Andean Cordillera. However, it is not yet clear whether contraction and mountain building were contemporaneous with marine deposition in this area. Orts et al. (2012) described progressive unconformities in the lower Miocene (∼18 Ma) marine succession of the Cerro Plataforma and in the probably coeval Late Oligocene–Middle Miocene? upper strata of the continental Nirihuau Formation, which they interpreted as evidence of ongoing contraction during sedimenta-
tion. This agrees with the ~20 Ma age that most authors consider the start of contractional basin inversion in south-central Chile and Argentina (Muñoz et al. 2000; Jordan et al. 2001; Kay et al. 2006). However, Bechis et al. (2011) cited growth strata that they interpreted as synextensional in the Troncoso Formation, a mostly continental sedimentary unit that overlies the Río Foyel Formation and was dated as 16.6 ± 0.5 Ma (U-Pb on zircons, Bechis et al. 2012). Another uncertainty is the Pacific or Atlantic origin for the Río Foyel Formation and equivalent units [see above], which is difficult to resolve because of the uncertain identifications of its poorly preserved fossils and because its outcrops are absent to the west, along the central North Patagonian Andes.

We envisage two possible scenarios for the tec-tono-sedimentary evolution of the Andean Cordillera in this region at ~22–16 Ma. The first is that a narrow and subdued mountain range rose as a consequence of contractional inversion of basins formed during a previous extensional phase [Muñoz et al. 2000; Jordan et al. 2001; Kay et al. 2006]. This incipient cordillera separated two distinct Pacific and Atlantic marine transgressions or presented narrow seaways that permitted either a limited connection between the two seas or the incursion of the Pacific to the east of this range. In the second scenario, the previously described Oligocene–Early Miocene extensional tectonic phase continued in this region with marine deposition. This implies that the Andean Cordillera had not yet formed, which likely facilitated either the transoceanic connection or the incursion of the Pacific in the Río Foyel area. An emerging volcanic arc would have formed a series of islands between the eastern and western flanks of the present cordillera. According to most published data, a marine connection between the Atlantic and the Pacific in the first tectonic scenario is more likely.

The angular unconformity between the Nirihuau Formation and the overlying Collón-Curá Formation clearly evidences tectonic Andean uplift and contractional tectonics in the eastern flank of the North Patagonian Andes during the Middle to Late Miocene (Bechis and Cristallini 2005). Carbon and oxygen isotope studies indicate significant uplift of the southern Patagonian Andes at ~16.5 Ma [Błesniuk et al. 2005]. During the Late Miocene, deformation probably ceased in the eastern Main Andean Cordillera and was displaced to the western part (Folguera and Ramos 2002), where it has been related to the dextral strike-slip LOFZ (Cembrano et al. 2002, Thomson 2002, Adriasola et al. 2006). Evidence for this important tectonic discontinuity comes from structural analysis and ⁴⁰Ar/³⁹Ar geochronology from mylonites at 41°–46°S, which show a Late Miocene–Pliocene dextral strike-slip event, oblique-slip, and contractional deformation indicative of bulk transpressional deformation (Cembrano et al. 2000, 2002). These data agree with zircon and apatite fission-track thermochronology that supports denudation of the southern Andes at the same latitudes during the Late Miocene–Pliocene, related to dextral transpressional activity in the LOFZ combined with intense erosion by Plio-Pleistocene glaciation (Thomson 2002; Adriasola et al. 2006). Enhanced exhumation gave way to erosion of the majority of the Meso-Cenozoic volcano-sedimentary cover and widespread exposure of the plutonics that form the North Patagonian Batholith [Adriasola et al. 2006]. Cembrano et al. (2002) and Thomson (2002) proposed a crustal-scale transpressional pop-up structure defining the main range of the Patagonian Cordillera, principally based on the spatial distribution of geologic units, fission-track data that show deeper crustal levels in the central part of the Main Andean Cordillera, more abundant Meso-Cenozoic cover rocks in the borders of this orogen, and the oblique-slip and contractional shear zones in the eastern and western parts of the main Patagonian Batholith. This pop-up structure consists of several blocks characterized by different amounts of cooling and denudation and separated by major oblique- or reverse-slip faults that link at depth below the main trace of the Liquine-Ofqui fault (Thomson 2002).

Subsurface geologic data from the submerged Longitudinal Depression west of the Main Andean Cordillera agrees with this model. The ENAP Puerto Montt 1 well records Miocene marine strata covered by ~2000 m of Plio-Pleistocene deposits (Elgueta et al. 2000). The 4010-m well did not reach basement even though it was only ~20 km west of the western Main Andean Cordillera, where deformed Miocene marine strata of the Ayacara Formation crop out and Meso-Cenozoic plutonic rocks are widely exposed. Additional information from ENAP seismograms from the Golfo de Ancud, between Chiloé Island and the area where the Ayacara Formation crops out, shows subhorizontal strata, presumably Cenozoic, in the western Longitudinal Depression that become folded to the east (González 1983). Thus, in agreement with the Cembrano et al. (2002) and Thomson (2002) pop-up model, we infer a significant tectonic discontinuity between the Longitudinal Depression and the Main Andean Cordillera that separates an uplifted, strongly deformed domain to the east from a downward, mildly deformed domain to the west (fig. 7). Flex-
ural subsidence induced by the major uplift of the Main Andean Cordillera, in combination with significant glacial erosion, gave way to 2000 m of Plio-Pleistocene synorogenic deposits in the eastern part of the Longitudinal Depression (fig. 7).

With regard to the tectonic activity of the LOFZ in the study area, the age of this major discontinuity is uncertain, and it is not yet clear whether the activity of this fault zone is restricted to the Late Miocene to present that most age data infer (e.g., Cembrano et al. 2000, 2002; Thomson 2002; Adriasola et al. 2006) or whether it began prior to the Late Cre-
taceous and was reactivated at various times in response to an evolving plate-kinematic framework (Cembrano et al. 2000). The cause of the transpressional tectonics associated with the LOFZ has also been debated. Some studies speculated that oblique subduction was the driving mechanism (Hervé 1977; Beck 1988). However, Forsythe and Nelson (1985) considered the angle of oblique convergence in the region (∼25°) to probably be small to solely account for the strike-slip activity of the LOFZ, and they alternatively proposed the indenter effect of the Chile Ridge at the southern termination of the fault system. Cembrano et al. (2000) deduced that oblique subduction combined with strong intraplate coupling were the key factors in the long-term motion along the LOFZ. According to them, ridge collision is a second-order mechanism that has favored dextral displacement along the southern portion of the LOFZ only since the Pliocene. Whereas plate motions and convergence angle have not changed substantially during most of the Cenozoic, partitioning of the Nazca–South America slip vector into forearc shortening and intra-arc dextral-strike displacement may have been continuous for most of this period (Cembrano et al. 2000). On the basis of the lack of apparent denudation between 42°S and 46°S before ∼16 Ma that was indicated by fission-track data, Thomson (2002) discarded transpressional tectonics in this area before the Middle Miocene. Because the Chile Ridge first collided with the Pacific margin of the South American plate at ∼14 Ma and very rapid cooling and denudation rates in the southern Andes between ∼7 and 2 Ma are concurrent with the collision of several short segments of this spreading center, Thomson (2002) concluded that the indenting of the Chile Ridge acted as a major driving force in sustaining the late Cenozoic dextral transpressional tectonism. Similar conclusions were reached farther north [41°–42°S] by Adriasola et al. (2006), also on the basis of fission-track data.

Our data indicate that the onset of major regional subsidence and deep-marine sedimentation in the Ayacara area (i.e., very close to the main trace of the LOFZ) occurred during the Early-Middle Miocene, which implies that transpression in the area started afterward, in accordance with conclusions based on fission-track data (Thomson 2002; Adriasola et al. 2006). Therefore, motion along the LOFZ is probably linked to the arrival of the Chile Ridge and either did not commence before ∼14 Ma or was dominated earlier by horizontal or transtensional strike-slip tectonism, as suggested by Thomson (2002).

Finally, integrating our data with published information enables us to propose a model (summarized in fig. 7) for the Neogene tectono-sedimentary evolution of the study area. First, regional subsidence, probably related to subduction erosion, affected the entire forearc and at least the western part of the Main Andean Cordillera during the Early-Middle Miocene. Subsidence initiated a significant transgression from the west, and deepwater deposition ensued. An incipient growing Cordillera probably blocked the Atlantic transgression that reached the eastern flank of the Andes in the Río Foyel area from reaching the Pacific ingression, although narrow seaways probably connected the two oceans. Second, during the Late Miocene tectonic activity along the LOFZ led to the initial development of a transpressional pop-up structure that uplifted the Main Andean Cordillera and resulted in deformation and emergence of the Miocene marine Ayacara Formation strata. It is not clear whether subsidence and deep-marine sedimentation was still active in the forearc to the west, as the age of the marine deposits in that area is under revision [see above]. Third, transpressional deformation, uplift, and exhumation in the Main Andean Cordillera have continued from the Pliocene to the present day. Uplift and emergence of the Neogene marine successions in the Coastal Cordillera and western Longitudinal Depression also occurred during this interval. Their uplift has been attributed to underplating of sediments because their strata show little deformation and no evidence of significant tectonic shortening (Encinas et al. 2008). In the eastern part of the Longitudinal Depression, on the other hand, flexural subsidence, ascribed to the uplift of the Main Andean Cordillera and its intense denudation, gave way to deposition of ∼2000 m of Plio-Pleistocene synorogenic deposits.

Conclusions

First, marine deposits in the western flank of the North Patagonian Andes at ∼42°S, termed the Ayacara Formation, were deposited in a deep-marine environment during the Early-Middle Miocene. Second, correlation of the Ayacara Formation with coeval deep-marine deposits in the present Coastal Cordillera and Longitudinal Depression indicate that marine deposition was caused by a major regional event of tectonic subsidence probably related to subduction erosion that affected the entire forearc and the western flank of the Andean Cordillera. Third, an Early-Middle Miocene age for the Ayacara Formation is a reliable maximum for the onset of deformation and uplift of the western part of the North Patagonian Andes, in agreement with
previous studies based on microstructural analysis and fission-track thermochronology. Fourth, major regional subsidence and deep-marine sedimentation of the Ayacara Formation close to the main trace of the LOFZ during the Early-Middle Miocene implies that transpressional tectonics associated with this fault system commenced after the Middle Miocene, in accordance with fission-track data reported by others.

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