

Rapid and major coastal subsidence during the late Miocene in south-central Chile

Alfonso Encinas ^{a,*}, Kenneth L. Finger ^b, Sven N. Nielsen ^c, Alain Lavenu ^d, Luis A. Buatois ^e, Dawn E. Peterson ^b, Jacobus P. Le Roux ^f

^a *Departamento de Ciencias de la Tierra, Universidad de Concepción, Casilla 160-C, Concepción, Chile*

^b *University of California Museum of Paleontology, 1101 Valley Life Sciences Building, Berkeley, CA 94720-4780, United States*

^c *GeoForschungsZentrum Potsdam, Sektion 3.1, Telegrafenberg, 14473 Potsdam, Germany*

^d *Institut de Recherche pour le Développement (LMTG-UMR5563, CNRS/IRD/Université Toulouse 3), 14 Avenue Edouard Belin, 31400 Toulouse, France*

^e *Department of Geological Sciences, University of Saskatchewan, 114 Science Place, Sask., Canada SK S7N 5E2*

^f *Departamento de Geología, Universidad de Chile, Casilla 13518, Correo 21, Santiago, Chile*

Abstract

Sedimentological and paleontological studies, including foraminifera, ostracodes, gastropods, and trace fossils, were carried out on Neogene sedimentary successions and offshore boreholes of south-central Chile (~33°–45°S). Sedimentology shows the occurrence of a thin, shallow marine, basal conglomerate overlain by a succession that includes the following facies: massive sandstones, conglomerates, interbedded siltstones and sandstones showing Bouma cycles, parallel-laminated sandstones, synsedimentary breccias, slides, slumps, diamictites, and massive siltstones. These facies were deposited by gravity flows, with turbidity currents and sandy debris flows as the main modes of deposition. Paleontology indicates the occurrence of trace fossils assigned to the *Zoophycos* ichnofacies and deep-water (~2000 m) benthic foraminifera, ostracodes, and gastropods. Sedimentology and paleontology indicate that deposition took place on a slope apron during a period of rapid and major forearc subsidence. Planktic foraminifera indicate ages ranging from the late Miocene to the early Pliocene (zones N16–N19) for these successions. We attribute this episode of major Neogene subsidence to an important event of subduction erosion that would have removed the underside of the upper continental plate and caused its thinning.

Keywords: Late Miocene; Early Pliocene; South-central Chile; Gravity flows; Foraminifera; Ostracodes; Gastropods; *Zoophycos* ichnofacies; Continental slope; Forearc subsidence; Subduction erosion

1. Introduction

Neogene strata crop out at different localities along the Chilean coastline (e.g., Cecioni, 1980; Ibaraki, 2001; Achurra, 2004; Le Roux et al., 2005) and have been recognized in boreholes drilled on the continental shelf (Mordojovich, 1981; González, 1989). Previous sedimentological studies, mostly conducted on the Navidad Formation (~33°–34°S), generally refer to these units as shallow-marine deposits (Etchart, 1973; Cecioni, 1978;

Cecioni, 1980), though paleontological studies indicate the occurrence of deep-water foraminifera and ostracodes (Osorio, 1978; Martínez-Pardo and Valenzuela, 1979). In addition, the published ages for these units are contradictory (e.g. Tavera, 1968; Martínez-Pardo, 1990; Ibaraki, 1992).

We carried out detailed sedimentological and paleontological studies on the Neogene marine successions of south-central Chile to try to unravel the sedimentary environment, age, and tectonic setting of these deposits. We present the results of these studies and discuss their implications with regard to the dynamics of the forearc.

* Corresponding author. Tel.: +56 41 2203997; fax: +56 41 2246075.
E-mail address: aencinas@udec.cl (A. Encinas).

2. Stratigraphy and sedimentology

Our sedimentological study focuses on the Navidad Formation ($\sim 33^{\circ}$ – 34° S), considered a reference for the marine Neogene of Chile (Cecioni, 1980; Martínez-Pardo, 1990; DeVries and Frassinetti, 2003). We supplement this investigation with less detailed studies of the coeval Ranquil Formation (García, 1968) at Arauco ($\sim 37^{\circ}$ S) and the Lacui Formation (Valenzuela, 1982) at Chiloé ($\sim 42^{\circ}$ S) (Fig. 1).

The Navidad Formation was first described by Darwin (1846), who gave this name to the Neogene marine strata that crop out in the coastal bluffs near the town of Navidad (Fig. 1). More than a century afterward, several authors proposed different stratigraphic schemes for the Neogene successions of the Navidad area (e.g., Etchart, 1973; Cecioni, 1978; Tavera, 1979). In this study, we follow the scheme proposed by Encinas et al. (2006), who divide these strata into the Navidad, Licanheu, Rapel, and La Cueva formations.

The Navidad Formation is approximately 100–200 m thick and overlies Upper Cretaceous marine strata of the Punta Topocalma Formation (Cecioni, 1978) or the Paleozoic granitic basement and underlies the Licanheu Formation (Encinas et al., 2006). It comprises a basal conglomerate overlain by a succession of interbedded siltstone and sandstone with minor conglomerate (Figs. 2 and 3). The unit contains a diverse fossil biota that includes bivalves, gastropods, crabs, ostracodes, foraminifers, shark

teeth, leaf impressions, and pollen (Philippi, 1887; Tavera, 1979; Martínez-Pardo and Valenzuela, 1979; Méon et al., 1994; Troncoso, 1991; Troncoso and Romero, 1993; Finger et al., 2003; Suárez et al., 2006).

2.1. Facies description

The basal conglomerate crops out at only a few localities (Figs. 2 and 3). It is typically a few meters thick but sometimes absent and consists of granitic with minor schistose and mafic basement clasts that are subrounded to angular and range in diameter from a few millimeters to more than a meter. Sometimes it displays alternating beds differentiated by the wide range of clast sizes (from a few centimeters to several meters). The rock is clast supported with an arenitic matrix and contains mollusks, solitary corals, whalebones, shark teeth, and neritic foraminifers (Table 1, samples NV1a and NV3a), as well as wood fragments (some containing the ichnofossil *Teredolites*) and scarce siltstone rip-up clasts. Locally the fossil content is very abundant, forming coquinas dominated by fragments of barnacles, oysters, and echinoderms. In some places, the conglomerate is absent, and a succession of interbedded sandstones and siltstones directly overlies a planar surface carved on the granitic basement.

The most continuous basal succession crops out about 2 km south of Boca Pupuya. It begins with a 15 m thick, clast-supported conglomerate overlying the granitic basement comprising decimeter-sized, subrounded to

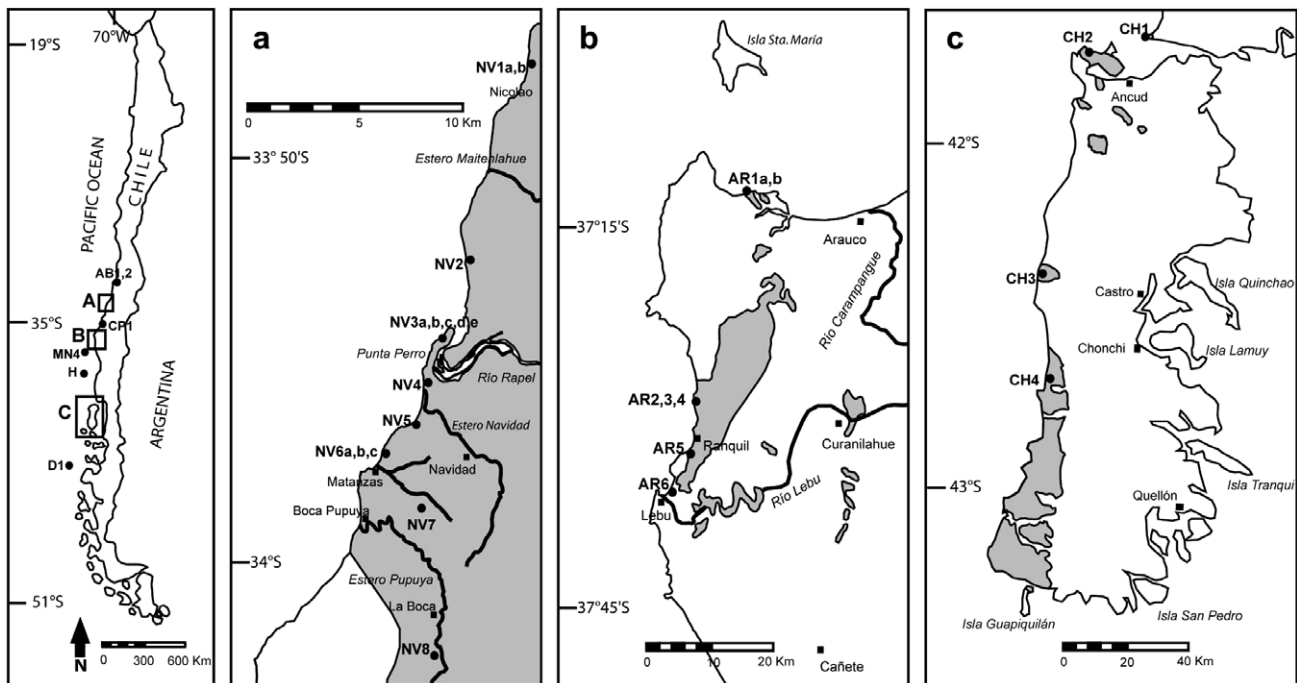
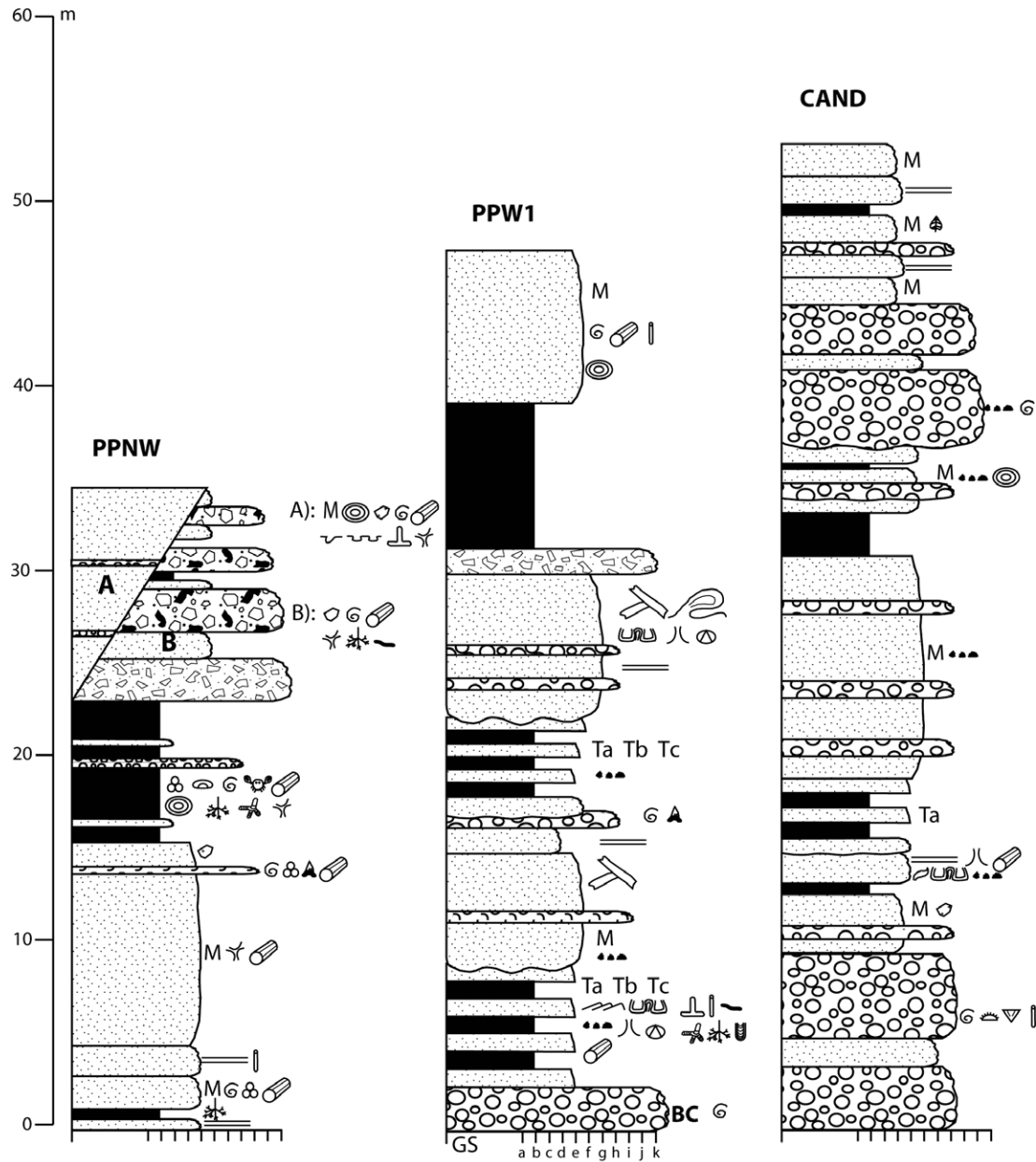


Fig. 1. Location map with plots of foram sample localities cited in Tables 1 and 2. (a–c) Foram sample localities in the Navidad, Arauco, and Chiloé, respectively. Samples labelled with the same name followed by letters (e.g., NV3a, NV3b) were collected from the same or proximal and correlative sections; sample a is lowest. The Navidad, Ranquil, and Lacui formations of the Navidad, Arauco, and Chiloé areas are represented in grey. Extension of the cited formations compiled from Ferraris and Bonilla (1981), Sernageomin (2002), and Lavenu and Encinas (2005).



LEGEND

LT: 1) 2) 3) 4) 5) 6) 7) BC

SS: 8) M 9) 10) Ta-e 11) 12) 13) 14) 15) 16) 17) 18) 19) 20) 21) 22)

F: 23) 24) 25) 26) 27) 28) 29) 30) 31)

TF: 32) 33) 34) 35) 36) 37) 38)

Fig. 2. Representative sections from Punta Perro (sections PPW1 and PPNW) and Nicolao (section CAND), Navidad area (see Fig. 1 for location of sections). GS, grain size: (a) clay, (b) silt, (c) very fine sandstone, (d) fine s., (e) medium s., (f) coarse s., (g) very coarse s., (h) gravel (granules), (i) gravel (pebbles), (j) gravel (cobbles), (k) gravel (boulders). LT, lithology: (1) conglomerate, (2) breccia, (3) diamictite, (4) coquina, (5) sandstone, (6) siltstone, (7) BC, basal conglomerate; SS, sedimentary structures; (8) massive, (9) parallel lamination, (10) Bouma cycles, (11) slides, (12) slumps, (13) convolute lamination, (14) climbing ripples, (15) water-escape structures, (16) load structures, (17) flute casts, (18) sheared flames, (19) rip-up clasts, (20) calcareous concretions, (21) pumice clasts, (22) floating clasts. F, fossils: (23) gastropods, (24) foraminifers, (25) ostracodes, (26) echinoderms, (27) solitary corals, (28) shark teeth, (29) crustaceans, (30) plant fragments, (31) leaves. TF, trace fossils: (32) *Chondrites*, (33) *Zoophycos*, (34) *Ophiomorpha*, (35) *Thalassinoides*, (36) *Skolithos*, (37) *Diplocraterion*, (38) *Planolites*. The top part of the PPNW section shows abrupt lateral facies changes in a few tens of meters, passing from massive sandstones with minor conglomerates into a succession of breccia, diamictites, sandstones, and minor siltstones.



Fig. 3. Shallow marine basal conglomerate (bottom right) overlain by a succession of deep marine interbedded sandstone and siltstone, Navidad Formation. Coastal bluff is approximately 50 m high.

angular granitic clasts in an arenitic matrix. The conglomerate is overlain by a 15 m thick succession of planar cross-bedded, medium- to fine-grained sandstone with abundant thin conglomerate lenses containing well-rounded, centimeter-size clasts. Superjacent is a 24 m thick succession of sandstones similar to those of the underlying unit but with fewer conglomerate lenses and showing planar cross-bedding with foresets dipping in different directions; trace fossils of *Ophiomorpha nodosa*, *Skolithos linearis*, *Thalassinoides* isp., and *Conichnus conicus*; as well as scarce escape traces. At the top there is a 7 m thick succession of fine-grained sandstone with low-angle planar cross-stratification and *Macaronichnus segregatis*. This succession is in fault contact with a succession of interbedded coarse and very fine sandstones showing the ichnogenus *Chondrites*.

The basal conglomerate facies is interpreted as shallow marine and, as the overlying succession confirms, marks the beginning of a marine transgression. A coastal environment is supported by the occurrence of shallow-marine foraminiferal assemblages in some of the samples. The abundant fragments of barnacles, oysters, and echinoderms that compose the coquina beds are interpreted as products of wave reworking in the upper shoreface of a rocky coastal area. Planar surfaces carved on the granitic basement that occur in some areas are interpreted as wave-cut benches. The alternation of beds with clast sizes ranging from millimeters to meters, wood fragments, and scarce siltstone rip-up clasts that occur in some localities suggest these conglomerates were transported seaward from the continental margin. They are interpreted as wave-reworked delta fan conglomerates deposited in a shallow-marine environment. The intercalation of beds that differ widely in clast size suggests pulses of tectonically induced deposition, though an alternation of storm and fair weather conditions is another possibility.

The most complete section at Boca Pupuya reflects the initial tectonic movements that initiated basin subsidence.

The basal succession of conglomerate and cross-bedded sandstone is interpreted as deposited by alluvial fans. The overlying, sandy succession with abundant planar cross-bedding dipping in different directions and the ichnofacies of *Ophiomorpha nodosa*, *Skolithos linearis*, *Thalassinoides* isp., and *Conichnus conicus* indicates an upper-shoreface environment and the beginning of marine transgression. The top beds of low-angle, planar cross-bedded, very well-sorted sandstone with *Macaronichnus segregatis*, which is common of extremely high-energy conditions (Pemberton et al., 2001), indicates a foreshore environment.

2.2. Interbedded siltstone and sandstone, with minor conglomerate

Overlying the basal conglomerate, and in some places resting directly on the basement, is an interval of interbedded sandstone and mudstone with minor conglomerate, in which several facies can be distinguished: massive sandstones, conglomerates, interbedded sandstones and mudstones, parallel-laminated sandstones, siltstones, and disrupted deposits (Fig. 2). Massive sandstones and interbedded sandstones and mudstones are the most common facies in this interval.

2.2.1. Facies 1, massive sandstone

This facies consists of light yellowish-brown, medium- to coarse-grained massive sandstone. Beds are generally more than 1 m thick and laterally continuous but in some places pinch out or form very inclined and irregular contacts with the underlying beds. They can have either erosive or non-erosive basal contacts and sometimes exhibit load structures and large but poorly defined flutes. They lack well-defined grain-size variations and graded bedding. The sandstones occasionally include intercalations of conglomerate and fossiliferous stringers, parallel-laminated sandstones, thin siltstone layers, rip-up siltstone clasts (more common at the base of the unit), and highly bioturbated horizons (Fig. 4). Also present but not as common are pumice clasts (sometimes grouped in spherical balls), water escape marks, sheared flames, armored mud balls, and floating granules and clasts that range to more than 1 m in diameter. Ovoid calcareous concretions, which often include fossils, are common and usually coalesce into concretionary beds with irregular forms parallel to the bedding. Fossils are generally scarce and consist of wood fragments that are locally abundant and can reach more than 1 m in length, mollusks, and shark teeth. Fossils are generally well preserved, including some large and delicate bivalves. The ichnotaxa *Thalassinoides paradoxicus*, *Ophiomorpha* isp., and *Skolithos linearis* are common and in some places extraordinarily abundant (Fig. 5).

Shanmugam (2000) discusses the intense debate among sedimentologists regarding the mode of deposition of massive sandstone. We follow this author's interpretation that massive sandstones are deposited by sandy debris flows



Fig. 4. Massive sandstone with abundant siltstone rip-up clasts at the base, Navidad Formation. Some clasts are floating within the sandstone (right of the photo).

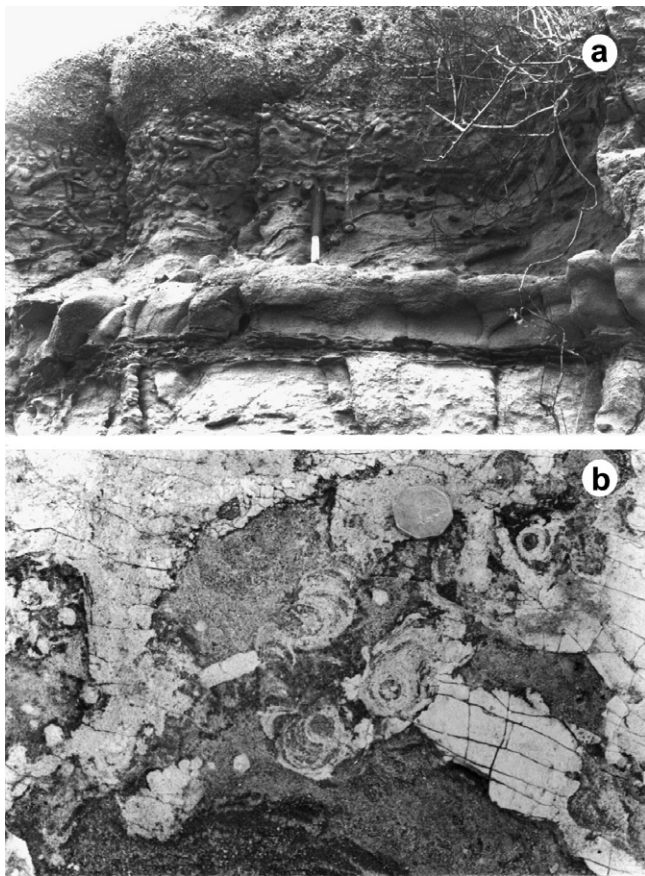


Fig. 5. Selected trace fossils from the Navidad Formation. (a) Abundant *Thalassinoides* isp. in medium- to coarse-grained massive sandstones (above hammer). (b) *Zoophycos* isp. in very fine-grained sandstones. Coin is 1.6 cm in diameter.

meters to meters thick, locally showing abrupt lateral variations in bedding thickness (Fig. 6). They can be coarsening- or fining-upward or ungraded. Clasts are subangular to subrounded and range from millimeters to meters in size. The conglomerates commonly include mollusks that are

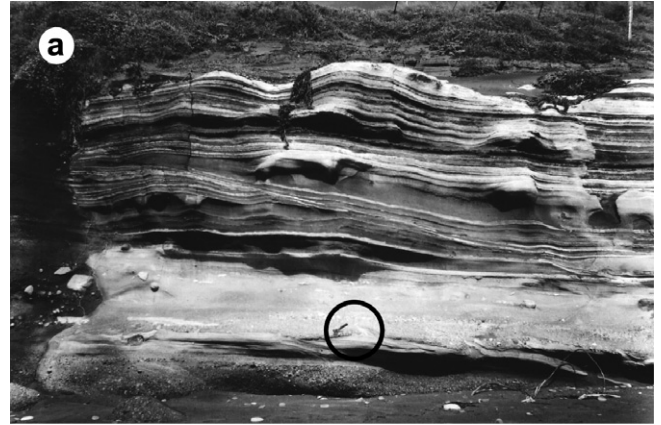


Fig. 6. (a) Interbedded sandstones (darker beds) and siltstones (lighter beds) in upper half of figure and massive conglomerate in lower half, Navidad Formation. Hammer for scale and location of photo b is encircled. (b) Detail of massive conglomerates. Interbedded is parallel-laminated sandstone with sheared flames. Knife is 10 cm long.

usually well preserved and sometimes show a preferred orientation. Siltstone and sandstone rip-up clasts are common and can range above 1 m in diameter. Basal contacts can be sharp or deeply erosive and sometimes show load structures. This facies locally shows intercalations of siltstones that are sometimes disrupted and fragmented, and it generally grades vertically and laterally into massive sandstone facies.

We interpret this facies as debris flow deposits on the basis of the poor rounding and wide size range of clasts, as well as the presence of matrix-supported conglomerates, inverse grading, and large rip-up clasts (Nemec and Steel, 1984). The presence of intercalated siltstone and rip-up clast layers indicate bed amalgamation.

2.2.3. Facies 3, thinly interbedded sandstone and mudstone

These beds are composed of medium- to coarse-grained sandstone alternating with siltstone or very fine-grained sandstone and range from centimeters to decimeters in thickness (Fig. 6). Sandstones predominate and form partial Bouma cycles in which Ta (massive), Tb (parallel-laminated), and Tab are more abundant and Tbc (parallel-laminated and asymmetric ripples) is scarcer. Rip-up clasts

are common; less common are load structures, dish and pillar structures, pumice clasts, convolute lamination, and climbing ripples. The basal contacts of these successions are usually sharp but sometimes channeled. The ichnotaxa *Chondrites* isp., *Zoophycos* isp., *Lophoctenium* isp., *Diplocraterion parallelum*, and *Planolites* isp. are common in siltstones and very fine-grained sandstones (Fig. 5).

This facies is interpreted as deposited by turbidity currents (Bouma, 1962; Lowe, 1982; Walker, 1992). The abundance of basal Bouma cycle divisions (Ta, Tb, Tab) and the predominance of sandstone indicate they are proximal turbidites. The channeled basal contacts are interpreted as formed by turbulent erosion prior to deposition.

2.2.4. Facies 4, parallel-laminated sandstone

This facies consists of yellow to white, medium- to coarse-grained sandstone with upper flow regime parallel lamination. Successions are up to 4 m thick and consist of centimeter to decimeter thick beds. There are scarce intercalations of massive sandstone, which are generally coarser grained and show erosive contacts, and thin siltstone layers that are typically disrupted and transformed into rip-up clasts. Well-rounded pumice clasts, burnt wood fragments, and water-escape structures up to 1 m high are very common. Leaf casts are extremely abundant in some of the fine-grained sandstone intercalations.

We interpret this facies as deposited by turbidity currents (Bouma, 1962; Lowe, 1982; Walker, 1992). The unit is composed of amalgamated beds formed predominantly by Tb Bouma divisions. Abundant pumice clasts, burnt wood fragments, and fossil leaves suggest the onset of catastrophic pyroclastic flows that would have been channeled along paleorivers originating on the flanks of volcanoes (see Carey, 1991). These flows emptying into the marine realm produced increased sedimentation rates on the platform, causing turbidity currents down the continental slope (see Fisher and Smith, 1991). The large fluid escape structures offer evidence of rapid deposition.

2.2.5. Facies 5, siltstone

Siltstone, mudstone, and very fine sandstone occur in intervals ranging in thickness from one to several meters. The beds are usually massive, but some outcrops show thin bedding. Locally there are thin intercalations of coarse sandstone and fossiliferous microconglomerate. Some dark grey beds contain abundant foraminifers, ostracodes, crabs, gastropods, and bivalves. White tuffaceous siltstone and mudstone, with fewer or no fossils, also occur. Trace fossils of *Chondrites* isp. and *Zoophycos* isp. are very common.

This facies is interpreted as having been generated by distal turbidity currents Tde(h) and hemipelagic deposition Te(t) (Bouma, 1962; Lowe, 1982; Walker, 1992). The intercalation of coarse-grained sandstone and microconglomerate indicates sporadic deposition by more proximal turbidity currents and debris flows. Tuffaceous siltstones indicate an explosive volcanic provenance.

2.2.6. Facies 6, disrupted deposits

These deposits can be subdivided into four types: breccia, slides, slumps and diamictites. Breccia occurs as small lenses within sandstone and as bedded intervals several meters thick. The latter successions are associated with synsedimentary faults and form fining- and thinning-upward cycles that display a reduction in dip angle toward the top of the succession. These beds contain clasts derived from Facies 3 (interbedded mudstone and sandstone) that range from a few centimeters to more than a meter in diameter (Fig. 7). Small breccia lenses within sandstone are formed by fragments of the underlying beds.

Synsedimentary slides of siltstone and mudstone, and less commonly sandstone beds, are broken up into meter-sized segments with variable dips that sometimes reveal syndepositional thrusts and occasionally pass into intraformational breccias. They usually form thin successions in which the bottom beds are deformed but become undisrupted upward. Some beds occur within sandstones and show abundant sand injections entering bedding planes, as well as local deformation and brecciation.

Slumped beds of siltstone and less frequently sandstone occur either as locally folded, isolated strata within horizontal beds of similar lithology or as laterally continuous beds, a few meters thick, composed of intensely folded strata of similar lithology to those of the underlying and overlying beds (Fig. 8). Underlying strata usually have sheared flame structures below the contact with the slumped beds.

The diamictites comprise mixtures of angular fragments of sandstone and siltstone ripped up from the underlying units, angular to subrounded granitic clasts ranging in size from granules to boulders, and scarce tree trunks up to more than 1 m long. These clasts are mixed together within a sandy matrix (Fig. 9). They sometimes form repetitive fining-upward successions that are a few meters thick with erosional basal contact, which are overlain by diamictites bearing abundant rip-up clasts that grade upward into massive sandstones.



Fig. 7. Sedimentary breccia, Navidad Formation. Base of succession is at bottom part of photograph. Note large clast of interbedded siltstone and sandstone.

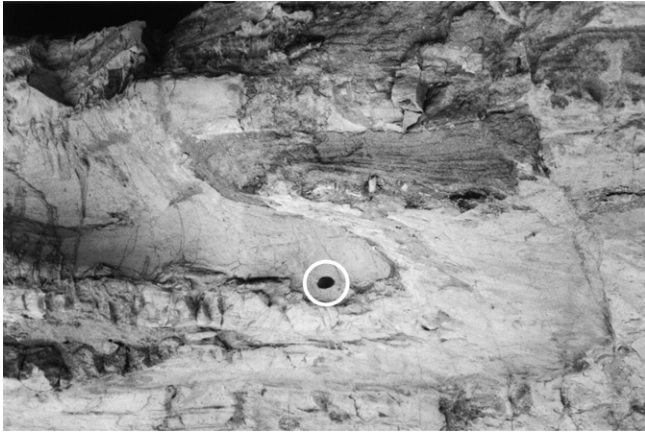


Fig. 8. Slump formed in interbedded siltstones and sandstones, Navidad Formation. Encircled coin for scale is 2.7 cm in diameter.



Fig. 9. Diamictites formed by a mixture of coarse-grained sandstones, siltstone intraclasts, and granite clasts, Navidad Formation.

Breccia, slides, and slumps consist of sediments formerly deposited by turbidity currents and sandy debris flows that were subsequently disrupted and displaced downslope. Such movements probably were triggered by depositional overloading or synsedimentary faults in an unstable setting (Bell and Suárez, 1995). Diamictites, slides, and breccia within sandstones are interpreted as having been scoured from underlying beds by sandy debris flows that would have transported huge granitic blocks and tree trunks as well.

With respect to the depositional architecture of these six facies described, no clear upward grain size and strata thickness variations are detected. Most beds are continuous with planar contacts, though some pinch out or are channel-shaped. In some places, they have highly inclined basal contacts or show abrupt lateral changes in facies. In some localities, it is possible to observe the presence of firm grounds with abundant specimens of *Thalassinoides* isp., filled with material derived from the overlying bed.

3. Paleontology

Foraminifers, ostracodes, and gastropods, primarily collected from the Navidad, Arauco, and Chiloé areas, were studied to determine the ages and depositional environments of the sedimentary successions.

3.1. Foraminifers and ostracodes

Foraminifers and ostracodes were extracted from siltstones and mollusk-bearing lenses within sandstones and conglomerates at Las Cruces (33°30'S), Navidad (~34°S), Concepción (~36°30'S), Arauco (~37°S), and Chiloé (~42°S). We examined core samples from hydrocarbon exploration wells drilled on the continental shelf by ENAP near Mocha Island (Mocha Norte #4 well), Valdivia (H well), and Taitao Peninsula (Darwin #1 well). Microfossil species of greatest utility in the interpretations are listed in Tables 1 and 2, and the sample locations are plotted in Fig. 1.

Most of the samples that we examined, particularly those that are relatively well preserved, yield associations of planktic foraminifers that indicate ages within the Tortonian (late Miocene) to Zanclean (early Pliocene) interval, represented by planktic foraminiferal zones N16–N19 (zonation of Blow, 1969, updated by Berggren et al., 1995). Similar results were obtained for the Neogene successions of the three ENAP wells.

Approximately 100 ostracode species have been identified. The vast majority are endemic to the Miocene of the southeast Pacific, though some also occur in the Caribbean and southwest Atlantic. Among the fauna are species originally described from the Miocene and lower Pliocene of Trinidad, the Miocene of Brazil, and the upper Oligocene and lower Miocene of Argentina and Tierra del Fuego.

The occurrence of a Late Cretaceous globotruncanid and *Catapsydrax dissimilis* (middle Eocene–early Miocene) in some study samples indicates reworking of older sediments. Ostracode species originally described from pre-Tortonian deposits elsewhere and recognized in our samples at Chiloé Island may be the result of this phenomenon.

Most of the samples examined yield mixed associations of littoral, neritic, and bathyal species of foraminifers and ostracodes. The deepest dwelling taxa indicate minimum depths of deposition at upper middle bathyal (500–1500 m), lower middle bathyal (1500–2000 m), and lower bathyal (2000–4000 m) depths (Table 1). Only three sample localities, all from the Navidad area, yield assemblages devoid of bathyal indicators: NV1a, NV3a, and NV3b (Table 1, Fig. 1). NV1a and NV3a correspond to the basal conglomerate deposited in shallow water. In contrast, NV3b is a succession of sandstone and siltstone with sedimentary facies and trace fossils indicative of deep water, but the recovered assemblage is too weak for depth analysis. Samples CPI and AR6 were recovered from siltstone or sandstone facies that directly overlie the basement. Sample

Table 2
Planktic foraminifera collected from coastal successions and offshore boreholes of south-central Chile (~33°–45°S)

Biostratigraphic Data

AREA	SAMPLE	PLANKTIC FORAM MARKERS													PLANKTIC FORAM ZONES	AGE			
		Miocene-Pliocene																	
		<i>Globigerina apertura</i>	<i>Globigerina pseudobesa</i>	<i>Globigerina venezuelana</i>	<i>Globoquadrina dehiscens</i>	<i>Globorotalia punctulata</i>	<i>Globorotalia spheriomiozea</i>	<i>Globorotalia menardii</i> s.l.	<i>Globorotalia cibaoensis</i>	<i>Globorotalia conoidea</i>	<i>Globorotalia conomiozea</i>	<i>Neogloboquadrina continuosa</i>	<i>Neogloboquadrina acostaeensis</i>	<i>Neogloboquadrina humerosa</i>			<i>Neogloboquadrina pachyderma</i>	<i>Orbulina universa</i>	Paleoc.
LAS CRUCES	AB1 (4)																	indeterminate	Pliocene?
	AB2		X														X	N9-N19	Langhian-Zanclean
NAVIDAD	NV1a (1)																	indeterminate	indeterminate
	NV1b	X	X	X	X	X												N19	Zanclean
	NV2	X	X	X	X												X	N16-N19	Tortonian-Zanclean
	NV3a (1), (3)																	indeterminate	indeterminate
	NV3b (3)																	indeterminate	indeterminate
	NV3c	X	X		X													N16-19	Tortonian-Zanclean
	NV3d		X	X	X	X											X	N19	Zanclean
	NV3e (3)																	indeterminate	indeterminate
	NV4				X													N4b-N19	Miocene-Pliocene
	NV5	X	X	X	X	X	X											N19	Zanclean
	NV6a			X							X	X		X				N16	Tortonian
	NV6b				X						X			X				N16	Tortonian
	NV6c																	indeterminate	indeterminate
NV7																	indeterminate	indeterminate	
NV8			X	X													N4b-N19	Miocene-Pliocene	
CONCEPCIÓN	CP1 (2)		X														N19 or older	Miocene	
ARAUCO	AR1a		X	X		X											N19	Zanclean	
	AR1b	X	X	X	X	X				X	X		X				N19	Zanclean	
	AR2		X	X	X					X	?						N13-N16	Miocene	
	AR3		X	X	X												N13-N19	Miocene-Pliocene	
	AR4		X			X										X	N19	Zanclean	
	AR5			X	X						X			X			N16	Tortonian	
AR6 (2)		X														N16-N19	Tortonian-Zanclean		
MOCHA ISLAND	MN4 (163-343 m) (4)																N19?	Zanclean?	
	MN4 (352-595 m)			X					X	X							N17-N18	Tortonian-Messinian	
	MN4 (595-1118 m)															X	P1	Danian	
VALDIVIA	H (216-883.9 m)			X		X											N19	Zanclean	
CHILOÉ	CH1	X	X		X					X	X						N16	Tortonian	
	CH2	X	X	X	X					?	X						N16	Tortonian	
	CH3			X	X												N4b-N19	Miocene-Pliocene	
	CH4					X											N19	Zanclean	
TAITAO PENINSULA	D1 (270-1356 m)						X	X									N19	Zanclean	

Table contains most important planktic markers, foraminiferal zones of Blow (1969; updated by Berggren et al., 1995), and ages determined for each sample. Location of samples is indicated in Fig. 1. (4) Ages suggested by benthic foraminifera. See Table 1 caption for additional information about samples and numbers.

NV1b was collected from a massive siltstone that overlies an only 0.5 m thick basal conglomerate bearing neritic foraminifera (sample NV1a) (Table 1).

The upper-depth limits of several benthic foraminifera in the Chilean Miocene are based on the depth distributions of similar taxa currently living along the Pacific margin

of central South America (Bandy and Rodolfo, 1964; Resig, 1981). Among the lower middle and lower bathyal indicators in the Chilean Miocene are species of *Bathysiphon*, *Melonis*, *Osangularia*, *Pleurostomella*, *Siphonodosaria*, and *Sphaeroidina* that are similar or identical to those Van Morkhoven et al. (1986) classify as cosmopolitan deepwater taxa. We consider paleodepth indicated by benthic foraminifer species as fairly reliable for the following reasons: (1) The modern pattern of ocean circulation began in the middle Miocene in response to expansion of the antarctic icecap. The introduction of cold, more oxygenated waters into the ocean basins led to a turnover in benthic foraminifera and the onset of most of the species that exist today, which became bathymetrically restricted as deep-ocean water became increasingly stratified (Kennet, 1982; Lipps, 1993). Therefore, post-middle Miocene benthic foraminifera are considered especially useful in paleodepth interpretations (Lipps, 1993); (2) the presence in our samples of several lower middle and lower bathyal species that live in different oceans of the world with different physical and chemical parameters (temperature, pH, dissolved oxygen, substrate, currents, food availability) indicate that the water depth must be the most important factor controlling their distribution, which makes paleodepth interpretations based on these cosmopolitan species highly reliable; and (3) upward transport of bathyal benthic foraminifers is dismissed because wind-driven upwelling currents off the west coasts of continents take place at maximum water depths of 100–200 m (Suess and Thiede, 1983; Colling, 2004), and they are not in contact with the bottom boundary except very near the coast (Suess and Thiede, 1983). Therefore, this kind of current cannot transport middle to lower bathyal (>500 m) benthic foraminifers, some of which dwell within the substrate. In addition, if upwelling were capable of transporting deep marine benthic foraminifers upslope, the extant fauna restricted to deep water would be found at shallower depths in the regions where these currents occur.

As with the foraminifera, many of the Chilean Miocene ostracode genera are extant in the southeast Pacific, southwest Atlantic, Caribbean, and Southern oceans, and their modern depth distributions suggest that most of our Miocene assemblages are mixed depth associations resulting from downslope displacement and deposition at bathyal depths. Psychrospheric species, which Benson (1972a,b, 1975) distinguishes as belonging to a major global fauna characterizing cold water masses that occur at a depth greater than or equal to 500 m (Kennet, 1982), are recognized in the majority of assemblages studied (Table 1).

3.2. Gastropods

Gastropods were recovered from outcrops of the Navidad Formation and its equivalents, the Ranquil and Lacui formations. These units have been long regarded as shallow-water deposits due to their coarse-grained sediments and mollusk fauna, particularly the common gastropods

Lamprodomina, *Olivancillaria*, and *Testallium* (Watters and Fleming, 1972; Vermeij and DeVries, 1997; Nielsen, 2004). The gastropod fauna includes many indicators of nearby rocky shores (Nielsen et al., 2004), as well as many exclusively or predominantly warm-water genera such as *Nerita*, *Strombus*, *Sinum*, *Distorsio*, *Ficus*, *Zonaria*, *Olivancillaria*, *Terebra*, *Architectonica*, and *Heliacus* (see Covacevich and Frassinetti, 1980; Nielsen et al., 2004).

Most of the mollusks first described from these units by Sowerby (1846), Hupé (1854), and Philippi (1887) were found in coarse-grained sandstones. However, the sandstones are intercalated with siltstones that contain an entirely different and lesser known fauna that has few species in common with those of the sandstones (Nielsen et al., 2003, 2004).

We evaluated 200 gastropod species representing 110 genera from several localities in the Navidad, Arauco, and Chiloé areas and recognized three primary associations. First, an association typical of rocky coasts is found in conglomerates and very coarse sandstone beds and includes the genera *Fissurella*, *Nerita*, *Zonaria*, and *Austrocominella*. Second, an association typical of shallow-water sandy substrates occurs in fossiliferous lenses contained in medium-grained massive sandstones and includes the genera *Astele*, *Ameranella*, *Sassia*, *Aeneator*, *Nassarius* and *Austrotoma*. Most of the species described from the Navidad Formation belong to this association. Third, a much less common, deepwater association includes calliostomatids (*Otukaia*; Nielsen et al., 2004), aporrhoids (*Struthiochenopus*; Nielsen, 2005a), naticids (*Falsilunatia*), cassids (*Dalium*), volutes (*Adelomelon*; Nielsen and Frassinetti, 2007), ptychtractids (*Exilia*, Nielsen, 2005b), cancelariids, conids (*Borsonia*, *Borsonella*), and turrids (*Bathytoma*, *Cochlespira*, *Nihonia*). This association is found exclusively in grey siltstone beds that contain abundant bathyal foraminifera and ostracodes restricted to the cold, deepwater masses. Neither shallow- nor warm-water taxa occur in the deepwater mudstones. However, an anomalous association, found in a thick siltstone bed at Punta Perro (Fig. 1), includes deep-marine foraminifers and the warm-water gastropod *Xenophora* (Nielsen and DeVries, 2002). Although this stratum also yields several shallow-dwelling indicators, its fauna probably is a composite of parautochthonous and allochthonous taxa.

Associations 1 and 2 are shallow water assemblages originally living at or near rocky and sandy coasts, respectively. Because they are found in massive sandstones and conglomerates (Facies 1 and 2) interbedded with siltstones containing deep marine foraminifers and icnofacies, we interpret these fossil associations as subsequently transported downslope by debris flows and ultimately deposited at bathyal depths without significant mixing with other sediments or mollusks. In contrast, Association 3 is a deepwater assemblage contained within siltstone beds (Facies 5). Fossils in this association appear to have been deposited in situ or experienced little transport, as evidenced by the fine-grained sediments and the abundance of deep-marine

foraminifers and trace fossils. Yet it is not clear if the coastal Associations 1 and 2 are contemporaneous with this deepwater fauna or were reworked from significantly older strata. Correlation of some shallow-water mollusk species from the Navidad and equivalent formations with those from confidently dated sequences in southern Peru (DeVries and Frassinetti, 2003; Nielsen et al., 2003; Finger et al., 2007) suggests a late Oligocene–early middle Miocene age for at least part of the shallow-water component.

4. Depositional setting

The combination of sedimentological and paleontological data obtained from this study allows a reconstruction of the depositional environment of the studied successions. The basal conglomerate and coquina are interpreted as coastal facies that mark the beginning of a marine transgression. The succession overlying the basal conglomerate appears to have been deposited primarily by gravity flows alternating with the settling of fine particles from suspension. The main depositional modes of transport were debris flows and turbidity currents. Gravity flows and the mixed-depth thanatocenosis of fossils, including foraminifers, ostracodes, gastropods, and shark teeth (Suárez et al., 2006), indicate downslope transport of sediments and deposition in deep water.

Deep-water deposition also is suggested by the occurrence of two distinct trace fossil assemblages (Fig. 5). The first occurs in fine-grained sandstone and siltstone beds and includes *Chondrites* isp., *Zoophycos* isp., *Lophoctenium* isp., *Diplocraterion parallelum*, and *Planolites* isp. It is dominated by feeding structures (Fodinichnia) that mainly represent the activity of deposit-feeding organisms. This association is ascribed to the *Zoophycos* ichnofacies, characterized by low oxygen levels associated with high organic detritus in quiet-water settings, and is typical for slope and apron settings (Frey and Pemberton, 1984; Buatois et al., 2002). The second trace fossil assemblage occurs mostly in massive, medium- to coarse-grained sandstone and includes *Thalassinoides paradoxicus*, *Ophiomorpha* isp., and *Skolithos linearis*. It is dominated by dwelling structures (Domichnia) and mostly records the activity of suspension feeders. This assemblage characterizes the *Skolithos* ichnofacies, which is typical of high-energy shallow-marine environments, but also has been widely reported in deep-water settings, where it reflects local environmental conditions such as high energy, sandy substrate, high levels of oxygen, and an abundance of suspended organic particles (e.g., Crimes, 1977; Crimes et al., 1981; Buatois and Lopez-Angrián, 1992; Uchman, 1995). In a deep-water environment, the *Zoophycos* ichnofacies would predominate during calm, low sedimentation intervals, whereas the *Skolithos* ichnofacies reflects short-term, high-energy conditions associated with the sudden deposition of thick packages of massive sand. The presence of firm grounds in strata of the Navidad Formation and abundant bioturbation in massive sandstone and even in

brecciated successions indicates long, nondepositional intervals between periods of sudden and rapid deposition.

We infer that sedimentation took place on a slope apron because the deposits have (1) lower bathyal foraminifers; (2) *Zoophycos* ichnofacies; (3) reduced sediment thickness typical of slope aprons; (4) rare systematic vertical and lateral trends in facies organizations that characterize slope environments rather than more organized submarine fan settings (Lomas, 1999); (5) abrupt lateral facies changes; (6) disrupted, small-scale deposits that suggest downslope movement of unstable material formerly deposited by turbidity currents and debris flows, rather than the large slumps and debris flows originating from major scarp failures (Bell and Suárez, 1995); and (7) low rates of sedimentation indicated by intensive bioturbation and the presence of firm grounds, which suggest that bypassing of sediments was an important process in the basin.

Abundant plant debris and some freshwater gastropods suggest a deltaic source area. Some layers have an abundance of rounded pumice clasts, which indicates episodic explosive volcanism in the arc domain to the east. The abundance of sandstone and conglomerates and presence of locally abundant basement clasts up to several meters in diameter, as well as large tree trunks, suggest that the depositional area was close to the continent and the continental shelf was rather narrow. The basal conglomerate/coquina, which was deposited in a nearshore environment, is generally not more than a few meters thick and overlain by deep-water deposits, which in some places directly overlie the basement, indicating rapid subsidence of the basin.

5. Age

The age of the Navidad Formation and its correlative units, the Ranquil and Lacui formations, has been a matter of debate for several decades among paleontologists. Tavera (1968, 1979) assigned a Burdigalian (early Miocene) age to the Navidad Formation on the basis of correlations with mollusks from the Patagonian of Argentina. Several workers study foraminifers from this unit and obtain different ages that include early Miocene (Dremel, in Herm, 1969), middle Miocene (Martínez-Pardo and Valenzuela, 1979), early–late Miocene (Martínez-Pardo, 1990), and late Miocene (Martínez-Pardo and Osorio, 1964; Cecioni, 1970; Ibaraki, 1992). In his study of ostracodes, Osorio (1978) also assigns a probable late Miocene age to the Navidad Formation. The Ranquil Formation has been ascribed to the Miocene by Tavera (1942) and García (1968) on the basis of mollusks and foraminifers, respectively. The Lacui Formation has been assigned to the late middle Miocene on the basis of foraminifers (Sernageomin, 1998).

A thorough study of planktonic foraminifers recovered from onshore exposures and ENAP core samples for this work indicate that the age of the Navidad Formation and equivalent onshore and offshore strata from south-central Chile range between planktonic foraminifer zones N16 and N19 (late Miocene–early Pliocene) (Table 2, Fig. 1).

The occurrence of late Oligocene–early middle Miocene gastropods (DeVries and Frassinetti, 2003) and late Oligocene–early Miocene shark teeth (Suárez et al., 2006) suggests that these fossils were reworked from an older stratigraphic unit that has not yet been identified in the field.

6. Discussion

6.1. Miocene subsidence

The results of the sedimentological, paleontological, and ichnological studies on Neogene exposures and offshore boreholes of south-central Chile indicate the occurrence of sedimentary successions deposited at bathyal depths during the late Miocene–early Pliocene. Studies elsewhere in Chile also indicate the presence of Neogene deep-marine successions that crop out along the Coastal Cordillera and Central Valley of this country. We cite, from north to south, the published antecedents that refer to these deposits.

Chile's northernmost Neogene marine strata are located about 80 km south of Iquique (21°S), attributed to the late Miocene–Pliocene on the basis of diatoms (Padilla and Elgueta, 1992). Near Antofagasta, at Caleta Herradura de Mejillones (~23°S), Ibaraki (2001) correlates planktic foraminifera within zones N7–N17 (early–late Miocene), and benthic foraminifera indicate bathyal to outer shelf depths for the succession (Scott Ishman, pers. com., 2004). At Caldera (27°S) and Carrizalillo (29°S), Neogene successions are attributed to the middle Miocene–Pliocene (Achurra, 2004; Gómez, 2003; Le Roux et al., 2005), and benthic foraminifera indicate some of the deposits accumulated at bathyal depths. It is interesting to note that in the cited areas, middle Miocene ages are based on Sr isotope data that were obtained only in the lowermost part of the successions, whereas planktic foraminifera and Sr isotopes indicate late Miocene–Pliocene ages for the rest of the succession. Farther south, Neogene strata crop out in the Central Valley and the western flanks of the Main Andean Cordillera between Temuco and Puerto Montt (38°30'–41°30'S). At Temuco, foraminifera recovered from the ENAP Labranza#1 borehole allow Osorio and Elgueta (1990) to determine that this area reached a water depth greater than 2000 m during the middle–early late Miocene interval. However, a revision of the planktic foraminifera listed by these authors shows the occurrence of *Globigerina druryi* in some intervals, indicating a late Miocene age (zones N16–N17). In the Valdivia and Osorno-Llanquihue basins (~39°30'–41°30'S), a marine transgression initially resulted in the development of estuarine peat swamps during the late Oligocene–early Miocene, followed by deep marine embayments during the middle Miocene, according to Le Roux and Elgueta (2000) and Elgueta et al. (in press). However, foraminifera listed for this area (Martínez-Pardo and Zúñiga, 1976; Martínez-Pardo and Pino, 1979; Marchant and Pineda, 1988; Marchant, 1990) include *Globige-*

rina pachyderma, which has its first appearance in the late Miocene at approximately 11 Ma, and *Melonis pompilioides*, a cosmopolitan lower bathyal (>2000 m) indicator. Farther south, Tavera et al. (1985) and Frassinetti (2001, 2004) cite the occurrence of marine deposits with a molluscan fauna similar to that of the Navidad Formation and turbiditic facies between Chiloé and Taitao (~42°–47°S), and Forsythe et al. (1985) find marine successions with late Miocene planktic foraminifera at Golfo de Penas (47°–48°S).

Our data and these aforementioned studies indicate that the coastal area from north to south-central Chile was subject to a significant Neogene marine transgression. The sea extended as far as the present eastern Coastal Cordillera and Intermediate Depression between 38°30'S and 41°30'S (Osorio and Elgueta, 1990; Elgueta et al., 2000). The marine transgression took place after a regressive period that lasted for most of the Oligocene (García, 1968). Because the water depth of some of these successions was at least 2000 m, and maximum eustatic sea-level changes are just on the order of a few hundred meters (Haq et al., 1987), we attribute this marine transgression to a major event of forearc subsidence.

Although constrained within the Neogene, the beginning of subsidence appears to have a different age depending on the area. For example, subsidence commenced during the early Miocene in Antofagasta, according to Ibaraki (2001), but our dating, which is based in a thorough sampling of planktic foraminifera (some recovered from strata directly overlying the basement), reveals that it did not begin in the Navidad, Arauco, and Chiloé areas until the late Miocene–early Pliocene. It is possible that subsidence was not coeval along the coast of Chile. However, the occurrence of reworked middle Eocene–early Miocene foraminifera, late Oligocene–early Miocene shark teeth (Suárez et al., 2006), and late Oligocene–early middle Miocene gastropods (DeVries and Frassinetti, 2003) in the successions of south-central Chile opens the possibility that subsidence commenced in the late Oligocene–early Miocene in this area as well. Perhaps deposition took place initially during the late Oligocene–early Miocene, and the marine strata were subsequently eroded and their fossils reworked and incorporated into the late Miocene–early Pliocene deposits. Whatever the case, and the age of commencement of the subsidence in the forearc, the data cited indicate that it is constrained to the Neogene for all localities along Chile and that most of these areas experienced a significant amount of subsidence during the late Miocene.

Evidence for major subsidence during the Miocene has been reported elsewhere around the Pacific, including New Zealand (Buret et al., 1997), Costa Rica (Vannucchi et al., 2001), Guatemala (Vannucchi et al., 2004), Japan (Von Huene et al., 1982), and Peru (Von Huene and Suess, 1988). The occurrence of this important and widespread event of subsidence around the Pacific suggests that it was due to a common cause.

6.2. Possible causes of subsidence

What are the causes of this major Neogene subsidence that affected the Chilean forearc? Important subsidence of convergent margins can be produced by three main processes, according to Von Huene and Scholl (1991): (1) depression of the lower oceanic plate, (2) thinning of the upper plate by extension and seaward sliding of the margin, and (3) subduction erosion. Depression of the lower oceanic plate can be caused by loading of a growing accretionary prism or an increase in the bulk density of the subducting plate. Accretionary prisms, however, are very small or absent along most of the Chilean margin (Bangs and Cande, 1997; Von Huene et al., 1997, 1999), and they could not have affected areas that are located many tens of kilometers landward of the present accretionary prism (Von Huene and Scholl, 1991). Plate reconstructions (e.g., Pardo-Casas and Molnar, 1987), in contrast, show that the age of the subducting plate, which is proportional to its bulk density, is relatively young (maximum 48 Ma) and progressively decreases due to the western motion of the South American plate. Crustal thinning by massive-scale sliding at convergent margins should generate regional-scale extension faulting and the translation of rotating blocks from the continental margin to the trench (Von

Huene and Lallemand, 1990; Von Huene and Scholl, 1991). However, tectonic analysis of the Navidad area (Laveno and Encinas, 2005) and offshore seismic lines do not show such large faulting (Mordojovich, 1981; González, 1989; Bangs and Cande, 1997; Laursen and Normark, 2003).

Subduction erosion is the process by which the subducting plate removes the rock and sedimentary bodies from an ocean margin. It is divided into frontal and basal erosion (Von Huene and Lallemand, 1990). Frontal erosion loosens and removes rock and sediment masses located at the toe of the landward trench slope, causing a landward migration of the trench axis. Basal erosion subcrustally removes the underside of the upper plate, causing its thinning and the subsidence of the margin (Von Huene and Scholl, 1991). This process has been considered the cause of Neogene subsidence of several Pacific margins, such as those of Peru (Von Huene and Suess, 1988), Japan (Von Huene and Lallemand, 1990), New Zealand (Buret et al., 1997), Costa Rica (Vannucchi et al., 2001), and Guatemala (Vannucchi et al., 2004). The most important arguments in favor of subduction erosion-induced subsidence for some of these margins are based on seismic lines and deep marine drilling and include (Von Huene and Scholl, 1991) (Fig. 10) (1) the absence of large accretionary wedges; (2) absence of

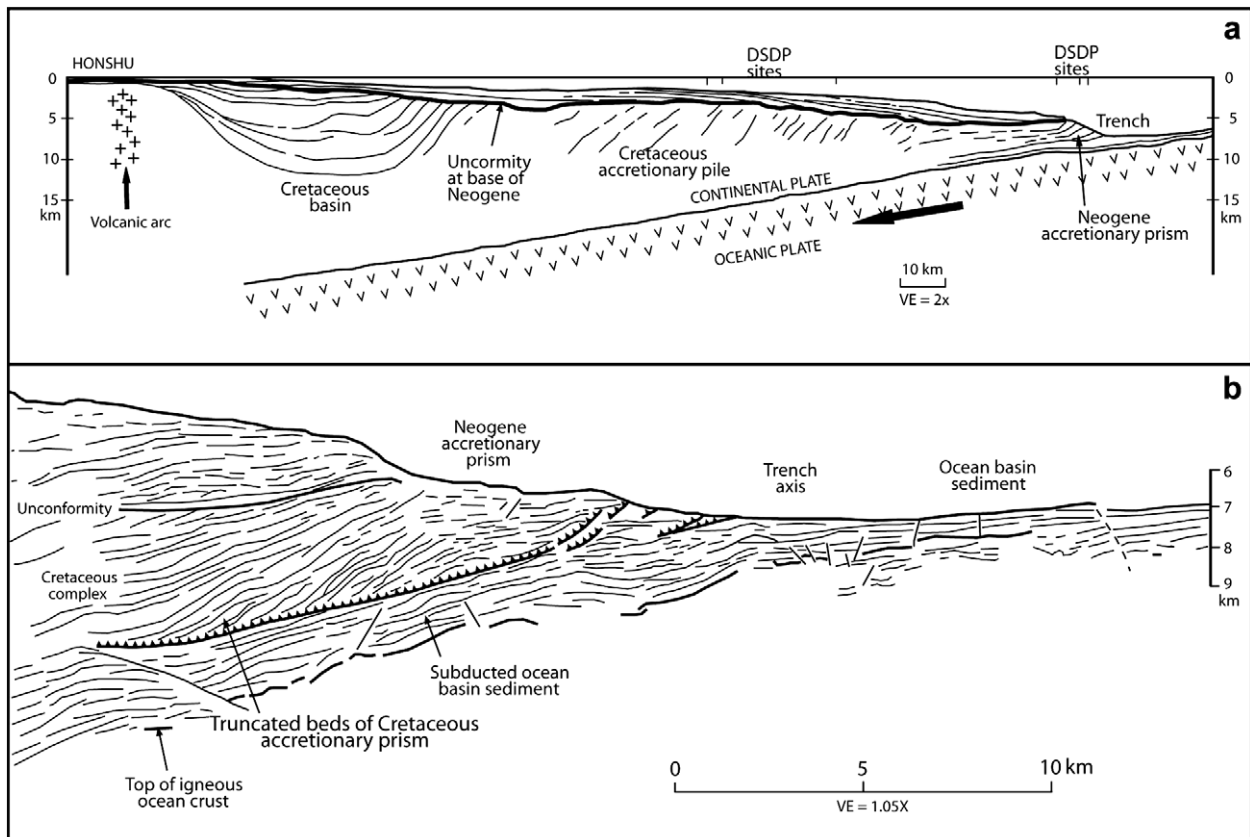


Fig. 10. (a) Geologic section across the Japan margin showing the presence of an important unconformity separating Neogene from pre-Neogene strata. The unconformity, originally carved near sea-level, is presently located at a depth of 4–5 km in the proximity of the trench. Unconformity major subsidence interpreted as produced by subcrustal tectonic erosion of the upper plate. (b) Detail of the Japan Trench, where the structural or depositional fabric of older sedimentary rocks has been truncated above the subducting oceanic plate. Truncation is also ascribed to subduction erosion along the base of the upper plate. Figures modified from Von Huene and Scholl (1991).

significant normal faults in the forearc; (3) presence of subsided discordances, separating Neogene from pre-Neogene successions, that were carved near sea level and presently are at a depth of more than 4000 m; and (4) occurrence of structural or depositional fabrics in the basal continental plate that are truncated at the contact with the subducting oceanic plate.

Considering these arguments, we regard subduction erosion as the most probable cause for the major subsidence of the Chilean margin during the Neogene (Fig. 11). This cause would agree with the small size of the accretionary wedges along most of its length (Bangs and Cande, 1997; Von Huene et al., 1997, 1999) and the minor deformation shown by Neogene successions (Mordojovich, 1981; González, 1989; Bangs and Cande, 1997; Laursen and Normark, 2003; Lavenu and Encinas, 2005). Other causes that imply a major extensional tectonic event, such as an increment in the roll-back of the subduction hinge relative to the advance of the overriding plate, are dismissed because they would give way to the onset of widespread extension that would affect not only the continental margin but also the Main Andean Cordillera. On the contrary, several researchers cite the onset of a major tectonic compressive phase in the Main Cordillera and sub-Andean ranges during the late Miocene (e.g., Jordan and Alonso, 1987; Ramos, 1989; Noblet et al., 1996; Gregory-Wodzicki et al., 1998). Tectonic erosion could account for major subsidence in the continental margin coeval with important uplift and compressive deformation in the Main Andean Cordillera. Eroding continental margins are characterized

by thin trench sediments (Underwood and Moore, 1995; Vanneste and Larter, 2002), which gives way to high shear stresses along the plate interface in the subduction zone due to the decrease in the lubricating effect that water-rich sediments produce, resulting in intense Andean uplift (Lamb and Davis, 2003). Compressive deformation would have little effect in the margin because the forearc constitutes a strong, cold, rigid geotectonic element that transfers, with minimal internal deformation, the nonseismic component of the convergence to the rheologically weakened regions of the orogen (Tassara, 2003, 2005).

After deposition of deep marine successions, the margin was uplifted during the Plio-Pleistocene, and marine deposits emerged, locally elevated to a few hundred meters above sea level. In some areas, as in Navidad, Pliocene shallow-marine deposits overlie bathyal successions, indicating that the basin remained at shelf depths prior its definitive emergence (Encinas et al., 2003). Considering that Neogene strata show very little deformation and no evidence of significant tectonic shortening (Lavenu and Encinas, 2005), we consider underplating of sediments the most probable cause for margin uplift. By this mechanism, subducted material is accreted at the base of the overriding plate at an intermediate position between the magmatic arc and the trench, leading to thickening and uplift of the continental plate without additional shortening (Cloos, 1989; Underwood and Moore, 1995). A similar explanation has been proposed to account for coastal uplift of the Coastal domain in northern (Von Huene et al., 1999; Hartley et al., 2000) and south-central Chile (Lohrmann et al., 2001).

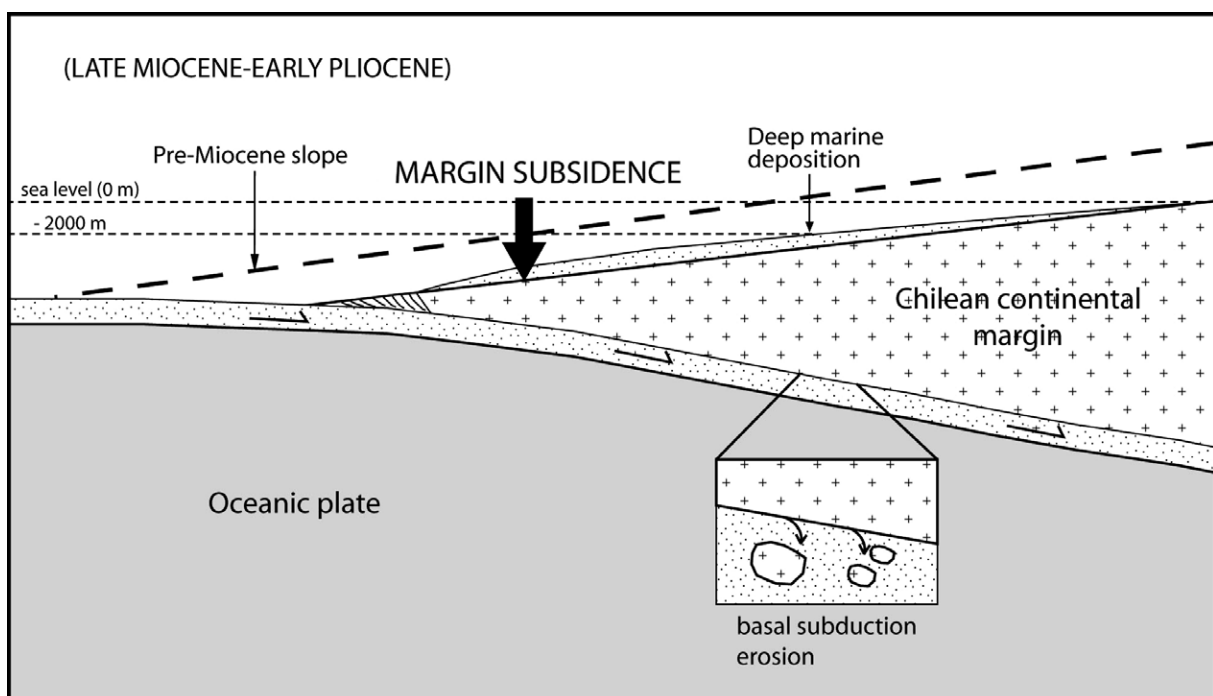


Fig. 11. Diagram (not to scale) illustrating forearc subsidence produced by subduction erosion in south-central Chile during the late Miocene. Thick dotted line demarcates top of continental crust prior to its subsidence.

6.3. Relationship between tectonic erosion and magmatism

The proposed event of subduction erosion during the Neogene is coincident with major eastward displacements of the volcanic arc that took place in various parts of Chile during this period (Cande and Leslie, 1986; Scheuber et al., 2000; Kay et al., 2005). Various authors (i.e., Gill, 1981; Jarrard, 1986; Tatsumi and Eggins, 1995) propose a fixed vertical distance of approximately 110 km between the volcanic arc and the subducting slab. Assuming a constant slab angle, the distance between the trench and the volcanic front should remain constant. Volcanic front displacements therefore are ascribed to tectonic erosion. One of the first authors to consider the possibility of significant truncation of the continental margin was Rutland (1971), who invoked this explanation for the occurrence of a belt of arc-igneous Jurassic rocks along the coast of northern Chile. This belt lies approximately 200 km west of the modern arc, so the same amount of continental crust had to be removed by subduction erosion (Rutland, 1971).

Arc displacement during the Neogene is evident in northern Chile (20°–26°S), where a new volcanic arc developed about 30–100 km east of the former one at approximately 25 Ma (Scheuber et al., 2000) and shifted farther east around 9 Ma (Wörner et al., 2000). Eastward arc migration also took place in central (33°–36.5°S) (Kurtz et al., 1997; Kay et al., 2005) and southern (47°–53°S) Chile (Cande and Leslie, 1986) during the Neogene. More complex is the situation between approximately 39°S and 47°S, where apparently no major systematic change in the location of the magmatic belt has taken place since the Late Jurassic (Pankhurst et al., 1999). We therefore find contradictory arguments for and against subduction erosion in the cited latitudes during the Neogene: major forearc subsidence and the lack of arc displacement. There are two possible explanations for this apparent contradiction, namely, either subsidence of this area was not caused by subduction erosion or subduction erosion took place in this area but the arc did not migrate. Major subsidence also occurred along central and northern Chile, as well as in Peru and other Pacific margins, where there is evidence to suggest that subduction erosion caused this subsidence (Von Huene and Suess, 1988; Von Huene and Lallemand, 1990; Buret et al., 1997; Vannucchi et al., 2001, 2004). Therefore, we consider it unlikely that this area subsided for a different reason. In addition, seismic reflection shows the occurrence of a small accretionary wedge in south-central Chile (~38°–40°S), which could have accumulated in the last 1–2 m.y. and reveals that nonaccretion and possibly erosion could have taken place in this margin prior to the increase in sedimentation rates during the Pleistocene glaciations (Bangs and Cande, 1997). Furthermore, interpretation of single-channel seismic data by Von Huene et al. (1985) suggests truncation of the continental margin in the Isla Mocha region (~39°). A possible explanation for the lack of displacement of the arc in this area during the Neogene is that magmatism emplacement was con-

trolled by the strike-slip Liquiñe Ofqui fault zone, as suggested by Hervé (1994) and Cembrano et al. (1996), who cite the deformed Mio–Pliocene plutonic rocks emplaced into a highly deformed wall rock along this fault zone as evidence of syntectonic intrusion.

Coeval with major shifts of the volcanic front in the Neogene are important changes in magma geochemistry that have been related to crustal contamination, as well as the generation of adakitic rocks and formation of important mineral deposits (Trumbull et al., 1999; Kay and Mpodozis, 2002; Kay et al., 2005). Significant epithermal, porphyry gold, and porphyry copper deposits were generated between 26° and 34°S (e.g., Clark et al., 1983; Maksaeu et al., 1984). Of particular importance are the giant upper Miocene–lower Pliocene porphyry copper deposits located between 32° and 34°S, notably those at Los Pelambres (~10 Ma) (Sillitoe, 1973; Mathur et al., 2001) and Río Blanco-Los Bronces and El Teniente (6.46–4.37 Ma) (Deckart et al., 2003; Maksaeu et al., 2003).

Crustal contamination–related changes in the chemistry of magmas and generation of porphyry copper deposits usually have been associated with crustal thickening and uplift related to tectonic deformation (e.g., Kay et al., 1991; Skewes and Holmgren, 1993). In contrast, adakitic rocks have been ascribed to melting related to flat-slab subduction (Gutscher et al., 2000; Reich et al., 2003). However, Stern (1991) and Stern and Skewes (2003) attribute adakite formation, magmatic crustal contamination, and ore generation in the Los Pelambres, Río Blanco-Los Bronces, and El Teniente deposits to the contamination of magmas by subducted sediments and continental crust transported into the mantle by subduction erosion. These authors associate subduction erosion with the southward migration of the aseismic Juan Fernández Ridge and the resultant decrease in subduction angle. However, according to Kay and Mpodozis (2002), the El Teniente copper deposit is located south of the area reasonably affected by this ridge, which suggests that its influence was not determinant. Considering that the timing of the cited magmatic changes in central Chile is approximately coincident with the onset of major forearc subsidence, we speculate that tectonic erosion of the margin could have played an important role in the crustal contamination of magmas, the generation of adakites, and the formation of economically important mineral deposits.

7. Conclusions

New sedimentological and paleontological studies of Neogene sedimentary strata in coastal outcrops and offshore boreholes of south-central Chile (~33°–45°S) shed new light on the geologic history of the region's forearc.

Sedimentological studies show the occurrence of a thin basal conglomerate, deposited in a shallow-marine environment, overlain by a succession of sandstone, siltstone, and minor conglomerate deposited by gravity flows, with turbidity currents and sandy debris flows as the main

modes of deposition. Ichnological studies indicate the presence of abundant *Chondrites* isp. and *Zoophycos* isp., typical of slope settings. Paleontological studies reveal bathymetric mixing of littoral, neritic, and bathyal species of foraminifers, ostracodes, and gastropods, which indicates downslope transport and deposition at minimum water depths of approximately 2000 m. Planktonic foraminifera indicate that deposition of these successions took place during the Tortonian (late Miocene, N16) to Zanclean (early Pliocene, N19).

The cited sedimentological and paleontological evidence reveals that these marine successions were deposited on a slope apron after a period of rapid and major forearc subsidence. We ascribe the cause of this subsidence to an important event of subduction erosion that would have removed the underside of the upper continental plate and caused its thinning.

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