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Quaternary marine terraces, sea-level changes and uplift history of Patagonia, Argentina: comparisons with predictions of the ICE-4G (VM2) model of the global process of glacial isostatic adjustment

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Abstract

Quaternary marine terraces have been investigated along a 1000 km stretch of the coast of Argentinian Patagonia. Fossil mollusc shells, most in living position and collected from raised beaches, were dated using the U-series, ESR and ^{14}C methods. Our analyses show that Holocene sea-level in this region culminated 7000 to 8000 BP at 6–7 m amsl. This beach slightly increases in altitude southward. The last interglacial stage (5e) was identified at 16–17 m amsl whereas the highest and morphologically most distinctive radiometrically dated terrace at 250,000 to 330,000 BP exists at an elevation of 33–35 m amsl. We estimate a constant rate of tectonic uplift of 0.09 m/1000 yr since the middle Pleistocene. Using this estimate of the local rate of tectonic uplift we correct relative sea-level (r.s.l) observations for the Holocene epoch. In turn, the inferred Holocene sea-level histories are compared with those predicted using the ICE-4G (VM2) model of the global process of glacial isostatic adjustment. This model accurately predicts r.s.l history from all sites along the northern part of the east coast of the South American continent (Venezuela, Brazil). However, along the southern part of the coast of Argentinian Patagonia there is evidence of an influence that is not accurately represented in this version of the model of the global process of glacial isostatic adjustment. We suggest that this influence could be connected to the presence of the broad continental shelf that is located offshore of this region, but the influence of significant neotectonic uplift cannot be dismissed.

1. Introduction

There are many locations where long-term tectonism may significantly mask the relative sea-level variations that would otherwise have occurred due solely to the influence of the last deglaciation event of the current ice age (e.g. the Huon Peninsula of Papua New Guinea, Chappell et al., 1996), and others in which this source of contamination is more modest (e.g. Barbados: Matthews, 1973; Fairbanks, 1989). However, there has been relatively little effort to attempt to separate these two effects at any particular location. A notable exception is the analysis recently applied to the coral based record from the Huon Peninsula (Peltier, 1998b). In this paper our intention is to provide a similar detailed analysis of this issue for coastal Patagonia.

The eastern coast of South America is bordered by several morphostructural units, the most important of which is the Brazilian shield, which outcrops along the Brazilian and Uruguayan coast (Urien and Ewing, 1974). South of Rio de la Plata in Buenos Aires, the Pampa plain (Fig. 1a) is interrupted abruptly by a cliff, but south of Rio Colorado there exists a transitional passage to the Patagonian plateau (Zambrano and Urien, 1974). On the Argentinian shelf three main geological provinces can be distinguished. (1) The northernmost province, between Rio de la Plata and 38°S, can be considered as the farthest expression of the Brazilian shield to the south, and in the central part of Buenos Aires province metamorphic and intrusive rocks are found in outcrop (Stipanicic et al., 1971). (2) In the area lying between 38°S and 43°S, which corresponds mainly to a prolongation of the Colorado Basin, folded Paleozoic sedimentary

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sequences occur (Urien and Zambrano, 1972). (3) *The region south of* 44°S contains a variety of Permian and Middle Triassic rocks (Stipanicic et al., 1971). According to Harrington (1972) strong folding phases seem to be restricted to the northern sector of the Rio Colorado Basin. Harrington considers that the sedimentary infill was folded in a single tectonic phase, which took place after late Permian and before Miocene time.

Throughout the middle and late Quaternary period, mollusc-laden marine-terrace sequences accumulated along most of the east coast of the South American continent, from Guyana southward to Brazil, Uruguay and Argentina. In Argentina, which has over 3500 km of coastline facing the Atlantic Ocean, these marine-terrace deposits shape the coastal landscape morphology from the Río de la Plata to Patagonia and Tierra del Fuego at the southernmost tip of the South American continent. Coastal Argentinian Patagonia possesses a number of raised marine shorelines, consisting largely of raised marine deposits at varying elevations. The emerged marine deposits are probably due to plate tectonic processes and, therefore, are ultimately related to the mantle convective circulation. In contrast, the fall of sea-level in the late Holocene that is evident on the basis of these mollusc-laden deposits is probably a consequence of hydro-isostatic tilting following deglaciation (Peltier, 1988).

Concerning the ages and altitudes of Patagonian emerged beaches, which are critical to our ability to separate the tectonic and glacio-isostatic signals, there exist inconsistencies in the previous literature (e.g. Feruglio, 1949, 1950; Radtke, 1989; Rutter et al., 1989, 1990; Codignotto et al., 1992). These discrepancies may be related both to the use of inappropriate dating methods and/or altitude assignments, and have led to uncertainty concerning Quaternary uplift history. The unconsolidated shelly marine deposits found along the Patagonian coast, associated with late Holocene sea-level variations, have been known since the middle of the last century when D'Orbigny (1834–1847) and D'Orbigny (1842–1844) referred to the fossil invertebrates. Darwin (1846) also referred to the marine deposits of south Patagonia, though he provided no systematic description of them. The most complete descriptions of the chronostratigraphy, lithology and paleontology of the Quaternary marine terraces of Patagonia derive from the work of Feruglio (1933, 1949, 1950). Feruglio (1949, 1950) recognized a sequence of six fossil marine deposits, ranging from approximately 8 to 186 m above mean sea-level (amsl). He found these terraces in the same ranges of elevation at widely separated locations along the coast, as follows:

Terrace I:	170–186 m
Terrace II:	105–140 m
Terrace III:	70–80 m
Terrace IV:	35–40 m
Terrace V:	15–18 m
Terrace VI:	8–10 m

Feruglio (1950) attributed the emergence of the Patagonian marine sequences to a marginal uplift effect. However, he did not entirely rule out the possibility that the deposits were formed by a fall of eustatic sea-level.

In terms of age control for the Patagonian raised terraces, radiocarbon ages of mollusc shells reported by Codignotto (1983) were, for the Holocene shells, 1,330 ± 80 to 5,720 ± 105 BP, and for the Pleistocene shells 32,800 ± 140 to > 43,000 BP. Rutter et al. (1989, 1990) provided aminostratigraphy and ESR dating from various locations (San Antonio Oeste, San Blas, Peninsula Valdés, Bahía Bustamante and Puerto Deseado, see Fig. 1a). Their ESR results yielded Last Interglacial ages (OIS 5e) for some locations whereas ages obtained on the basis of aminostratigraphy suggest penultimate interglacial (OIS 7) or older deposits. ESR and Th/U dating by Radtke (1989) showed that Holocene beach risers are at higher elevation in the south of Patagonia than in the north. Further ^{14}C dating by Codignotto et al. (1992) subsequently indicated two rates of Holocene sea-level fall along the coast of Argentina, one of 0.12 mm/yr for the sedimentary basin terrain and another of 1.63 mm/yr for the interbasin regions. Gordillo et al. (1993) suggested, for Tierra del Fuego, a rate of sea-level fall of 1.5-2.0 mm/yr for the Holocene and 2.9 mm/yr in the last millennium. An area in the Straits of Magellan (Fig. 1a) has also been mapped by De Muro and Brambati (1995) who discuss Quaternary shorelines in this region. Schellmann and Radtke (1997) reported, for the Holocene shells from the Camarones and Bustamante sites, radiocarbon ages ranging from 5,380 ± 70 to 6,708 ± 46 yr BP and for Pleistocene shells, ESR ages belonging to Oxygen Isotope Stages 5, 7 and 9 for the entire coast of Patagonia. Schellmann (1998a, b) subsequently provided a detailed study on development of marine deposits from this region.

In this manuscript our intention is to discuss new data obtained by dating of representative terraces, as paleo sea-level markers, along the coastal area of Patagonia and to compare the observations of Holocene sea-level with results of predictions made on the basis of the ICE-4G (VM2) model of Peltier (1996). Furthermore, we will also compare predictions derived from the ICE-4G model for the northern sector of Argentina as well as for Brazil and Venezuela, with observations made for these additional regions. Pleistocene shorelines in Brazil are at 8–10 m amsl, in contrast to their counterparts in Patagonia which are at much higher elevations. Consequently we believe that Patagonia may have been tectonically uplifted. We attempt to determine both the local tectonic uplift rate to infer global Quaternary sea-level changes for the last 300 kyr, comparing the latter inferences to the results obtained by others based upon the analysis of coral based data sets.

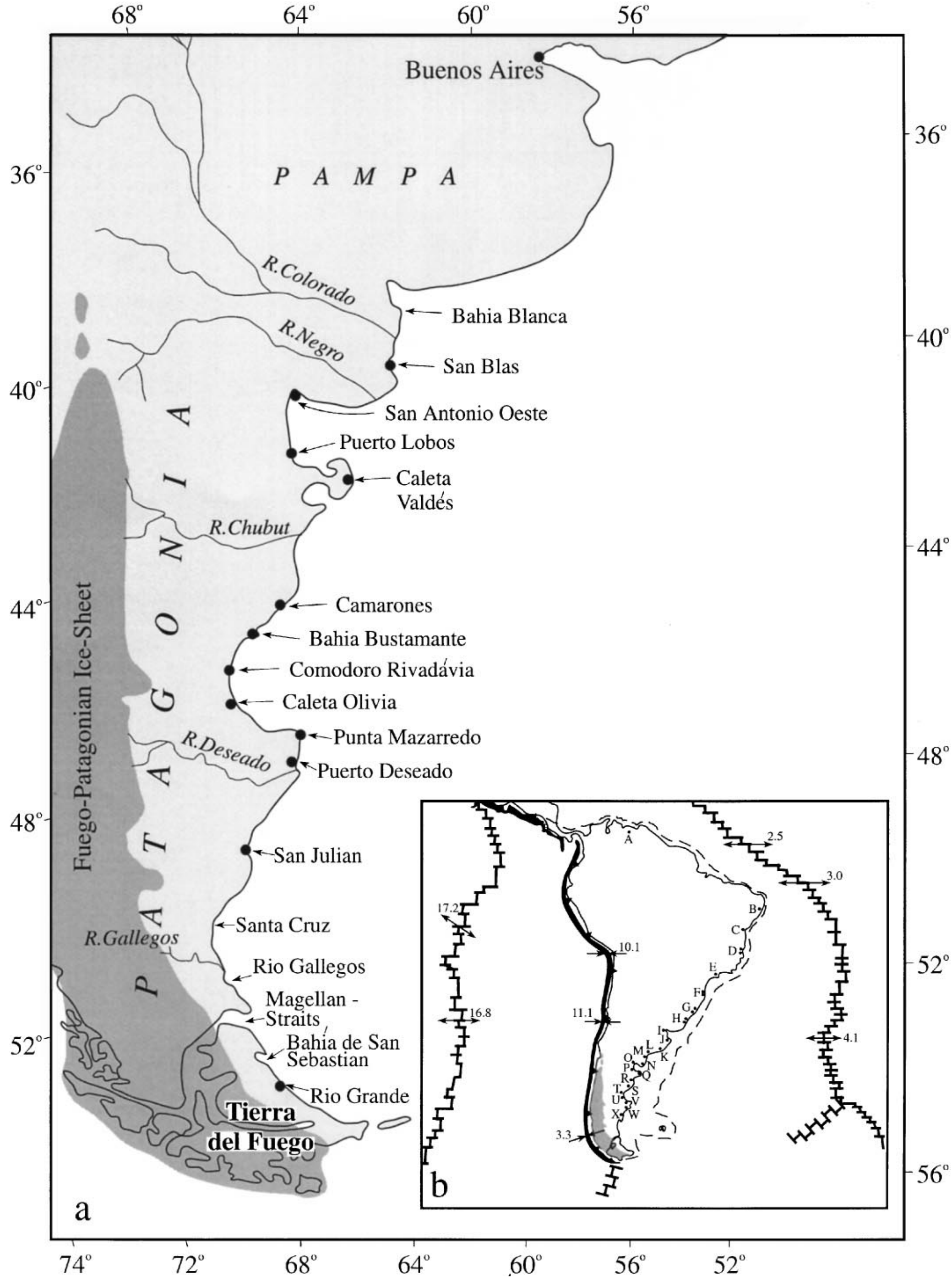

Fig. 1. (a) Location and generalized map of Argentina and Tierra del Fuego (Argentina and Chile) showing sites of main Quaternary marine deposits and raised beaches. The extension of the formerly glaciated region of South America (shaded) defines the limits of Fuego-Patagonian Cordillera ice sheet. Compiled from Polanski (1965); Flint and Fidalgo (1969); Flint (1971); Auer (1974); Hulton et al. (1994) and Clapperton et al. (1995) as well as personal investigations in Tierra del Fuego (K. R.). (b) Map of South America showing the main locations of Quaternary marine deposits and beach risers. The arrows give relative velocities of plate divergence (← →) and convergence (→ ←) in cm/yr. Along the western coast of South America converging arrows indicate the highest velocity where oceanic plate descends into the upper mantle. Because of the narrowing of the continent to the south in Patagonia, the proximity of the east coast of the continent to the subduction zone off the west coast increases. It is presumably this proximity to the subduction zone that is responsible for the tectonic uplift of the coast of Patagonian Argentina. The velocities of divergence and convergence zones are derived from DeMets et al. (1990). A = Venezuela; B = Recife; C = Salvador de Bahía; D = Illhéus and Itahype; E = Rio de Janeiro; F = Ilha Grande and Sao Paulo; G = Santos and Itanhaem; H = Cananela and Iguape; I = Gualeguay; J = Samborombon; K = Mar Chiquita; L = Bahía Blanca; M = Río Colorado; N = San Blas; O = Río Negro; P = Puerto Lobos; Q = Caleta Valdés; R = Camarones; S = Bahía Bustamante; T = Comodoro Rivadávia; U = Caleta Olivia; V = Mazarredo; W = Rio Deseado and X = San Julian, respectively. Note also the dramatic increase of the width of the continental shelf off the east coast of South America between Brazil and the south coast of Argentina. The location of the shelf break is shown as a dashed line.

To this end we have analyzed the morphology, sedimentology and structure of the Holocene and Pleistocene terraces between latitudes 42°S and 54°S. Ten different sites were considered along this 2000–3000 km long (ca. 1300 km of arc length) segment of coast. The individual sites are Puerto Lobos, Caleta Valdés, Camarones, Bahía Bustamante, Comodoro Rivadávia, Caleta Olivia, Punta Mazarredo, Puerto Deseado, San Julian and Rio Grande (Fig. 1a). Neither the Puerto Lobos location (in northern Patagonia) nor the Rio Grande location (in southernmost Tierra del Fuego) yielded any suitable shell samples, so the data reported here are derived from the subset of the eight remaining locations. Furthermore, at the Bahía Bustamante and Puerto Deseado sites there exist no representative Pleistocene marine deposits so that only Holocene information is available from them.

At each of the sites from which useful information is available we also observed a variety of features including (a) terraces containing re-cycled materials, including polished whole and broken shells of various size derivative of unknown sources and initial elevations; (b) storm-beach deposits, generally characterized by the presence of "out-washed" shells in sediment in proximity to polished fossil shells, sand, coarse gravel and boulders. Such features are generated by intense storm-induced wave action behind or at the inner margin of the beach, above the level reached by normal high spring tides or by ordinary waves; (c) ridges indicated by mound features that are found above the water level on the foreshore, and are characterized by fractured shells of various size. They may also occasionally contain whole shells; however, these are polished; (d) bars which are low ridges of sand or shingle that were laid down by marine aggradation in shallow water adjacent to a coastline and are easily recognized by the presence of shells that were washed ashore. We found that the Holocene ridges and storm-beach deposits were always at higher elevation than the terraces of the same age on which our study focused exclusively as sources of material for dating. Study of the additional features containing reworked material was nevertheless useful in helping us to establish criteria for sampling. We will therefore occasionally comment on such features when they aid the interpretation.

The estimation of the present elevation of each of the investigated terraces was based upon the average of three independent measurements, each of which was taken with respect to a new null at local sea-level. The reproducibility of these measurements was such as to imply an accuracy of the Holocene terrace elevations of $\pm$ 0.5 m, a consequence of the short distance between these Holocene terraces and the present position of the beach. For terraces of the Last Interglacial (OIS 5e) the reproducibility degraded such as to imply a reduced accuracy of 0.5–1 m (on windy days this increased further to 1.5 m), and for terraces formed during OIS 9 the reproducibility degraded still further to 1–2 m. For these older Pleistocene terraces the increased error in the height estimations are attributed to the much increased distance between the location of the terrace and the present location of the beach where the null-adjustment is made. (Note: the altitude measurements provided for the mussel samples extracted from individual terraces henceforth relate to the shell-bearing horizon in the terrace). All of these elevation measurements have been corrected to local mean sea-level by interpolation using the published tide tables for locations within the study area (Armada Argentina Servicio de Hidrográfica Naval, 1993/1995), and the results of these measurements are expressed as lower elevation bounds in Table 1, column 4 for each feature. According to tide tables, within this study area the tidal range varies from 2–3 m in the north (e.g., Caleta Valdés) to 6–7 m in the south (e.g., San Julian).

2. Sampling procedures, sample preparation and analytical methods

Shelly marine deposits differ in terms of their indicative meaning concerning sea-level from coral reefs as well as in their physico-chemical construction. Reefs are constructed primarily of the calcareous skeletal remains and secretions of corals and algae, which post-mortem form compact features. However, the mollusc habitat consists of hard or sandy substrate and individual samples are embedded in sediment. Therefore in sampling of molluscan fauna extra attention has to be paid to terrace sedimentology and stratigraphy, since shells can be easily washed out and displaced by storms, spring tides and surges. As first investigated by Fray and Ewing in 1963, the Argentinian shelf habitat comprises a rich variety of species, many of which are confined to the littoral and littoral to intertidal zone. In this article we will focus entirely on the use of mussels as indicators of former sea-level, using species which presently occupy the intertidal zone. Fossil mussels are usually found in sediment which, following exposure by tectonic uplift or falling sea-levels, is subject to subaerial weathering.

At most locations we have found that the altitudes of paleo shorelines measured by us differ from those of Feruglio (1950), Rutter et al. (1989), Radtke (1989) and Codignotto et al. (1992). An altitude range of 8–10 m amsl has consistently been reported for Holocene features. These elevations are precisely the values originally quoted by Feruglio (1950), as previously discussed, but these have apparently never been confirmed using modern methods of direct measurement. The approximate elevations of these features originally quoted by Feruglio were most probably obtained from topographic maps and it appears to us that later investigators have simply adopted these same elevations rather than measuring them directly since, for their purposes, more exact measurements were not important. An alternative explanation is that these elevations correspond to the "storm-beach" deposits or "ridges", which are generally found at higher elevations than the marine terraces themselves. These features were generally easy to recognize during our field expeditions, but they were considered unreliable sea-level markers and, therefore, are irrelevant for paleo sea-level analysis. Since tidal range along the Argentinian coast increases to the south (see above), it may be possible that wind- and wave-induced sea-level fluctuations (which are characteristically seasonal and most extreme along this stretch of coast that faces the prevailing westerly winds) are responsible for the formation of the higher elevation features. Especially in view of these complexities it will be clear that the value of any elevation measurement can only be assessed by answering the following questions: (i) what was the position of the altimeter employed to determine terrace elevation, i.e. was its position on top of the terrace or level with the shell-bearing horizon, (ii) how reproducible were the measurements, (iii) how accurate was the instrument (our measurements were made using the Swiss manufactured THOMMEN altimeter, which has a precision of ± 1 m), (iv) was the measurement properly corrected for the modern ocean tide in order to reduce it to msl. All of these issues were addressed in the course of our measurements. Failure to adequately address them might account for the discrepancy between our altitude measurements and those usually assigned to the same features by our colleagues (e.g. Feruglio, 1949, 1950; Radtke, 1989; Rutter et al., 1989, 1990; Codignotto et al., 1992).

In all of the localities that we have studied, there exists one terrace which marks the maximum of Holocene transgression. This may be easily distinguished from terraces at lower elevations (regressive features), because of its form as a wave-cut platform or because of its superior preservation. The intensity of eolian and coastal erosion often made it impossible to identify some regressive marine features at some locations. Higher paleo shorelines, interpreted as those formed during the Last Interglacial high stand (OIS 5e), are abundant at all locations in the study area at 6–17 m amsl. In some locations (e.g., Bahía Bustamante), however, the shells in the Pleistocene shorelines are strongly polished and reworked and, therefore, of little value for our purposes. Based on enhanced weathering of mussel shells, superior degree of sediment consolidation as well as increased degree of weathering of the terraces themselves, the most distinctive terrace is the terrace sequence which appears at approximately the same elevation (33–35 m) wherever it is found. In collecting shell samples to be used for the definition of relative sea-level, deep holes were excavated into the top of the marine terrace, far from the paleo-cliff, since this feature normally contains re-worked materials. Furthermore, because of eolian and weathering processes, which are characteristic of Patagonia, colluvium from the higher terrace has in most locations topped the inner (landward) edge of the lower one. We have therefore paid extra attention so as to avoid sampling such features. At some locations the sediment covering the shell-bearing horizon consisted of reworked material, including shells. We always removed this layer of sediment prior to sampling. In the Holocene marine terraces, fossil shells were found approximately 60 cm below the surface. In contrast, Pleistocene horizons bearing molluscs occurred at 100–120 cm depth. Shell-bearing horizons, subsequently dated as belonging to stage 9, were found at 150–200 cm depth. In addition to the mollusc samples we collected, additional samples from similar features were provided by our colleagues Dr. Codignotto and Dr. Schellmann. During a second expedition several of these sites were re-sampled and the ages of the additional samples that we obtained were found to be concordant with those provided by these colleagues.

To prevent contamination, the initial steps taken in the preparation of fossil shell samples involved removal of growth edges, detritus material and secondary carbonate from the shell surfaces. The shells were then cleaned with a diamond saw and dentist's drill, and finally leached in dilute HCL for about 2 min to remove any remaining surface contamination. Since the external and immediately underlying internal layers of the fossil specimens will clearly be more contaminated in radio-isotopes (Th-excess, detrital thorium and continental uranium) than the materials in the deepest innermost portion, a significant fraction of clean shell exteriors and interiors must also be removed to further diminish contamination. For this purpose we employed two procedures to extract the most pristine innermost portions of the mollusc shells. First we employed "one-step-dissolution" or "leaching" with dilute acid prior to analysis (a process which leads to the loss of 80–90% of the total sample weight). Second we employed "sequential leaching" in which, after removing the growth edges and all other evident foreign materials to a depth of about 1 mm from both sides of the shells, the shells were submerged in 60 ml diluted acid (1 N HCl = conc. $\sim 3\%$) for 3–4 min. Following removal from the acid bath and oven drying, weight loss can be calculated for the first sub-sample. To the first sub-sample were then added ^{228}Th and ^{232}U as a spike for measuring uranium and thorium concentrations by isotope dilution. This procedure was then repeated for the second and all other sub-samples. In this way, samples could be dissolved into six sub-samples (at least in the case of *Mercenaria*). The layers successively removed by such sequential leaching are as a rule parallel to the shell surfaces. For both the "sequential leaching" and "one-step dissolution" analyses, the Th/U, ESR and ^{14}C ages to be presented in what follows will refer to the innermost layer of each specimen. Note that the greater the initial thickness of the mollusc shells the higher should be the accuracy of the age determined. For terraces older than stages 5 and 7, because of their higher degree of alteration, we exclusively employed the hinge of the fossil shells for Th/U analyses since only this part of the shells remained aragonitic.

The uranium and thorium concentrations in the individual shells were ascertained using alpha spectrometry; the analytical procedures for separating and purifying were those described by Frank et al. (1994). Purified radioisotopes were electroplated onto Platinum discs and counted using a silicon surface barrier alpha detector. Radiocarbon ages were measured by beta counting. The ESR measurements were made with a Bruker EMS 104 spectrometer. All analyses were performed at the Heidelberg Academy of Sciences at the Institute for Environmental Physics at the Ruprecht-Karls-University in Heidelberg. A brief introduction to the procedure employed for ESR age determination is provided in Appendix A. For further detailed discussion of the ESR technique of age determination the interested reader should see Schellmann and Radtke (1999); Schellmann and Radtke (1997); Barabas et al. (1992a); Grün (1990, 1989); Barabas (1989) and Radtke et al. (1985). Th/U ages are calculated using the equation of Kaufman and Broecker (1965) which has the form:

$$\left(\frac{^{230}\mathrm{Th}}{^{234}\mathrm{U}}\right)_t = \frac{(1-\mathrm{e}^{\lambda_{230t}})}{(^{234}\mathrm{U}/^{238}\mathrm{U})_t} + \left(1 - \frac{1}{(^{234}\mathrm{U}/^{238}\mathrm{U})_t}\right) \times \frac{\lambda_{230}}{\lambda_{230}-\lambda_{234}}\,(1-\mathrm{e}^{-(\lambda_{230}-\lambda_{234})t})$$

in which λ_{230} is the decay constant of ^{230}Th of 9.217×10^{-6} per year, and λ_{234} is the decay constant of ^{234}U which is 2.79×10^{-6} per year (Ivanovich and Harmon, 1982) and t is the age of the sample.

The Th/U ages will be quoted together with a standard deviation determined on the basis of laboratory precision. Calculated ages and radiochemical data for all of the samples analyzed are documented in Table 1. Sketches of the sedimentary section at each sample site along with an overall description of the samples analyzed are presented in Figs. 2–7.

3. Site descriptions and dating results

3.1. *Caleta Valdés (42°15′ − 42°30′S)*

Caleta Valdés is located on the east coast of the Valdés Peninsula, about 100 km southeast of Puerto Lobos; it has a series of northwest–southeast trending paleo shorelines, here denoted by Roman numerals I, II, III, IV and V in descending order of elevation (Figs. 2a and b). We excavated holes at several locations in search of appropriate shell materials, and from the Pleistocene units collected *Mytilus* shells in living position that were relatively well preserved. The Holocene terrace (shoreline V), which covers the northern part of this site, was sampled by removal of approximately 80 cm of reworked sediment (Fig. 2c). This horizon is 4.5–5 m amsl. Mollusc sample 5-0/1, which was collected from this elevation gave a radiocarbon date of 4070 ± 90 kyr (Note: henceforth radiocarbon dates quoted here are in calendar years) and a Th/U age of 5000 ± 2000 BP. Samples vi and vii from this location were provided by Dr. Codignotto of the University of Buenos Aires. Shell vi was Th/U dated to 4000 ± 2000 BP whereas mollusc vii gave a Th/U age of 28000 ± 10000 BP. These ages were also reported by Codignotto et al. (1992) (not on the same shells) using radiocarbon, who obtained an age of 4010 ± 90 kyr for sample vi and an age of 5233 ± 90 kyr for shell vii. The discrepancy between the radiocarbon and Th/U ages for sample vii is most likely due to the fact that the shell was not sufficiently cleaned for Th/U analysis (^{230}Th/^{232}Th < 10, see also Table 1) and so the Th/U age determination is disregarded. An age of 4000 to 5000 yr BP is entirely plausible for this fossil marine terrace. Having studied the sedimentology of the Holocene terrace the question arose as to the origin of the 80–90 cm of reworked material, which topped the sampled horizon. This material may have been derived from a higher transgressive phase of the sea, been reworked and deposited at the lower elevation. Since radiocarbon ages of approximately 7000 BP at a terrace elevation of approximately 5 m amsl have been reported for locations north of the study area, i.e. in the Province of Buenos Aires (cf. Aguirre and Whately, 1995), which were considered to relate to a transgressive episode, we therefore assume that the sites that we have sampled at Caleta Valdés correspond to such regressive stages.

To sample terraces I and II (Figs. 2a and b), 120 cm deep pits were excavated 4–5 km inland. Shell 1-0/1 collected from terrace II (17 ± 0.5 m amsl) was leached in three sub-samples and the final sample (innermost portion of the shell) gave a Th/U age of 115,000 ± 5000 yr BP. The shell 3-0/1 taken from terrace I was also analyzed in three sub-samples. The shell was split in two for Th/U and ESR dating before totally dissolving the innermost part, and the dating gave concordant ages of 126,000 ± 10,000 BP and 137,000 ± 14,000 BP. In both terraces the consolidated shell-bearing horizon also contained clayey/silty sand (see Fig. 2d).

For terrace I, Rutter et al. (1989) previously reported D/I ratios (aspartic acid and leucine ratios) in the range 0.5–0.6 on *Mytilus* which suggests that this terrace represented OIS 7 (corresponding to an age of 190–230 kyrs before present). This apparent disparity is discussed further below.

3.2. *Camarones (44°40′–44°59′S)*

The city of Camarones is located approximately 250 km south of the Valdés site and approximately

Table 1
Radiochemical data and ages of fossil molluscan fauna from the Atlantic coast of Argentina. Isotopic ratios are activity ratios and the errors shown are based on counting statistic, $\pm 1\sigma$)*

Location	Sample ID	Lab. no.	Altitude (m amsl)	^{238}U (ppm)	$^{234}U/^{238}U$	$^{234}Th/^{230}U$	$^{230}Th/^{232}Th$	^{14}C age (cal. yrs)	Th/U age(kyrs)	ESR age(kyrs)
Caleta Valdés	vi	3983	4–5	0.07 ± 0.01	1.13 ± 0.08	0.03 ± 0.02	> 100	$4{,}070 \pm 90$	4 ± 2	
	vii	3984	4–5	0.03 ± 0.01	1.38 ± 0.06	0.20 ± 0.06	< 10	$5{,}233 \pm 90$	28 ± 10*	
	5–0/1	6564	5 ± 0.5	0.09 ± 0.05	1.08 ± 0.05	0.05 ± 0.01	15	$4{,}070 \pm 90$	5 ± 2	
	1–0/1	5689	16–17	1.56 ± 0.04	1.40 ± 0.05	0.68 ± 0.02	> 100		115 ± 5	
	3–0/1	5691	17–18	1.23 ± 0.05	1.27 ± 0.07	0.71 ± 0.03	> 100		126 ± 10	132 ± 14
Camarones	2–0/1	5536	5–6	0.57 ± 0.04	1.32 ± 0.04	0.06 ± 0.02	41		7 ± 2	
	3–0/1	6589	16–17	1.42 ± 0.03	1.47 ± 0.05	0.69 ± 0.02	> 100		117 ± 5	
	3–0/2	6590	16–17	0.97 ± 0.02	1.38 ± 0.04	0.68 ± 0.03	12		115 ± 9	110 ± 8
	3–0/3	6625	16–17	1.04 ± 0.05	1.43 ± 0.01	0.67 ± 0.05	15		112 ± 13	114 ± 9
	D2417/Pa31a	4625	16–17	2.03 ± 0.04	1.28 ± 0.03	0.68 ± 0.02	> 100		117 ± 6	
	4–0/2	6572	33–34	3.83 ± 0.08	1.36 ± 0.04	1.02 ± 0.03	> 100		$309 + 50{,}000/-35{,}000$	
	4-0/3	6573	33–34	3.23 ± 0.09	1.50 ± 0.05	1.07 ± 0.03	> 100		354 ± 45	
	4–0/4	6574	33–34	5.46 ± 0.06	1.43 ± 0.02	1.05 ± 0.02	> 100		338 ± 34	
Bustamante	1–0/1	5536	5–6	0.75 ± 0.04	1.50 ± 0.01	0.06 ± 0.01	17		7 ± 1	
	D2444/Pa43a	4599	5–6	0.40 ± 0.03	1.10 ± 0.03	0.07 ± 0.01	33		8 ± 2	
Comodoro	RT. SII	4420	ca. 3 ± 0.5	0.41 ± 0.04	1.31 ± 0.06	0.02 ± 0.01	41		3 ± 1	
	RT. NII	4419	ca. 5 ± 0.5	0.25 ± 0.04	1.11 ± 0.04	0.06 ± 0.01	41	$6{,}441 \pm 70$	7 ± 1	
	RT 2–0/2	221.2	ca. 4 ± 0.5	0.05 ± 0.01	1.11 ± 0.01	0.05 ± 0.01	> 100	$2{,}959 \pm 55$	$5{,}870 \pm 66$	
	RT 2–0/3	222	ca. 4 ± 0.5	0.09 ± 0.01	1.12 ± 0.01	0.03 ± 0.02	> 100	$2{,}959 \pm 40$	$3{,}030 \pm 27$	
Caleta olivia	7–1/1	4411	ca. 5 ± 0.5	0.10 ± 0.05	1.07 ± 0.08	0.03 ± 0.01	82		5 ± 1	
	D2619/Pa 72	4809	ca. 7.00	0.26 ± 0.01	1.19 ± 0.07	0.07 ± 0.01	82		8 ± 2	
	1–2/1	6117	16–17	1.30 ± 0.03	1.28 ± 0.04	0.69 ± 0.02	50		120 ± 6	132 ± 14
	1–2/2	6265	16–17	0.78 ± 0.02	1.40 ± 0.04	0.75 ± 0.02	36		137 ± 7	
	1–3/1	5535	16–17	2.92 ± 0.24	1.40 ± 0.05	0.74 ± 0.06	83		133 ± 15	
	1–3/3_b	6082	16–17	1.30 ± 0.04	1.18 ± 0.05	0.75 ± 0.03	> 100		143 ± 11	
Mazarredo	1–2/1_b	6226	ca. 6 ± 0.5	0.14 ± 0.01	1.43 ± 0.15	0.07 ± 0.01	15		8 ± 2	
	2–0/1	6227	ca. 5 ± 0.5	0.26 ± 0.02	1.39 ± 0.13	0.04 ± 0.01	82		5 ± 1	
	Maz 24	3992	ca. 5 ± 0.5	0.03 ± 0.01	1.17 ± 0.46	0.07 ± 0.01	82	$4{,}050 \pm 90$	8 ± 2	
	Maz 27	3994	ca. 6 ± 0.5	0.06 ± 0.02	1.20 ± 0.03	0.01 ± 0.01	17	$5{,}233 \pm 90$	1 ± 1	
	5–1/1	6627	16–17	1.00 ± 0.04	1.47 ± 0.09	0.64 ± 0.04	> 100		104 ± 10	
	5–1/3	6641	16–17	1.45 ± 0.05	1.13 ± 0.06	0.77 ± 0.03	97		137 ± 7	
	5–2/3	6642	16–17	1.10 ± 0.05	1.31 ± 0.07	0.65 ± 0.04	57		108 ± 10	
P. Deseado	2–0/1a	6561	6–7	0.30 ± 0.04	1.12 ± 0.05	0.06 ± 0.01	81		8 ± 1	
San Julian	2–0/1	6560	7–8	0.03 ± 0.01	1.17 ± 0.46	0.07 ± 0.04	81		8 ± 2	
	1–0/1	5675	16–17	0.95 ± 0.04	1.26 ± 0.08	0.67 ± 0.03	> 100		115 ± 9	
	1–0/2	6140	16–17	0.65 ± 0.03	1.12 ± 0.07	0.72 ± 0.04	> 100		134 ± 14	99 ± 9
	1–0/3	6271	16–17	1.23 ± 0.02	1.36 ± 0.03	0.72 ± 0.02	> 100		128 ± 6	
	4–0/2	6608	33–34	1.85 ± 0.06	1.48 ± 0.06	1.06 ± 0.04	> 100		342 ± 54	
	4-0/3	6609	33–34	1.22 ± 0.06	1.61 ± 0.09	1.18 ± 0.06	> 100		$\geqslant 355$	
	4–0/4c	6743	33–34	1.01 ± 0.04	1.51 ± 0.08	0.97 ± 0.06	78		240 ± 38	
Caleta Olívia	1–1/1	6089	14–15	1.50 ± 0.06	1.32 ± 0.05	0.95 ± 0.03	> 100		239 ± 28	
	1–1/2	6090	14–15	0.90 ± 0.03	1.31 ± 0.05	0.80 ± 0.03	> 100		158 ± 13	
	5–0/1	5633	14–15	1.41 ± 0.04	1.33 ± 0.05	0.89 ± 0.03	> 100		200 ± 19	
	5–0/2	5801	14–15	1.48 ± 0.03	1.31 ± 0.03	0.90 ± 0.02	> 100		208 ± 12	222 ± 36
	5–0/3	6360	14–15	1.60 ± 0.04	1.36 ± 0.05	0.99 ± 0.03	65		274 ± 30	

*We used a radiocarbon age of $5{,}000 \pm 100$ for this mussel sample (see text for Valdés site).

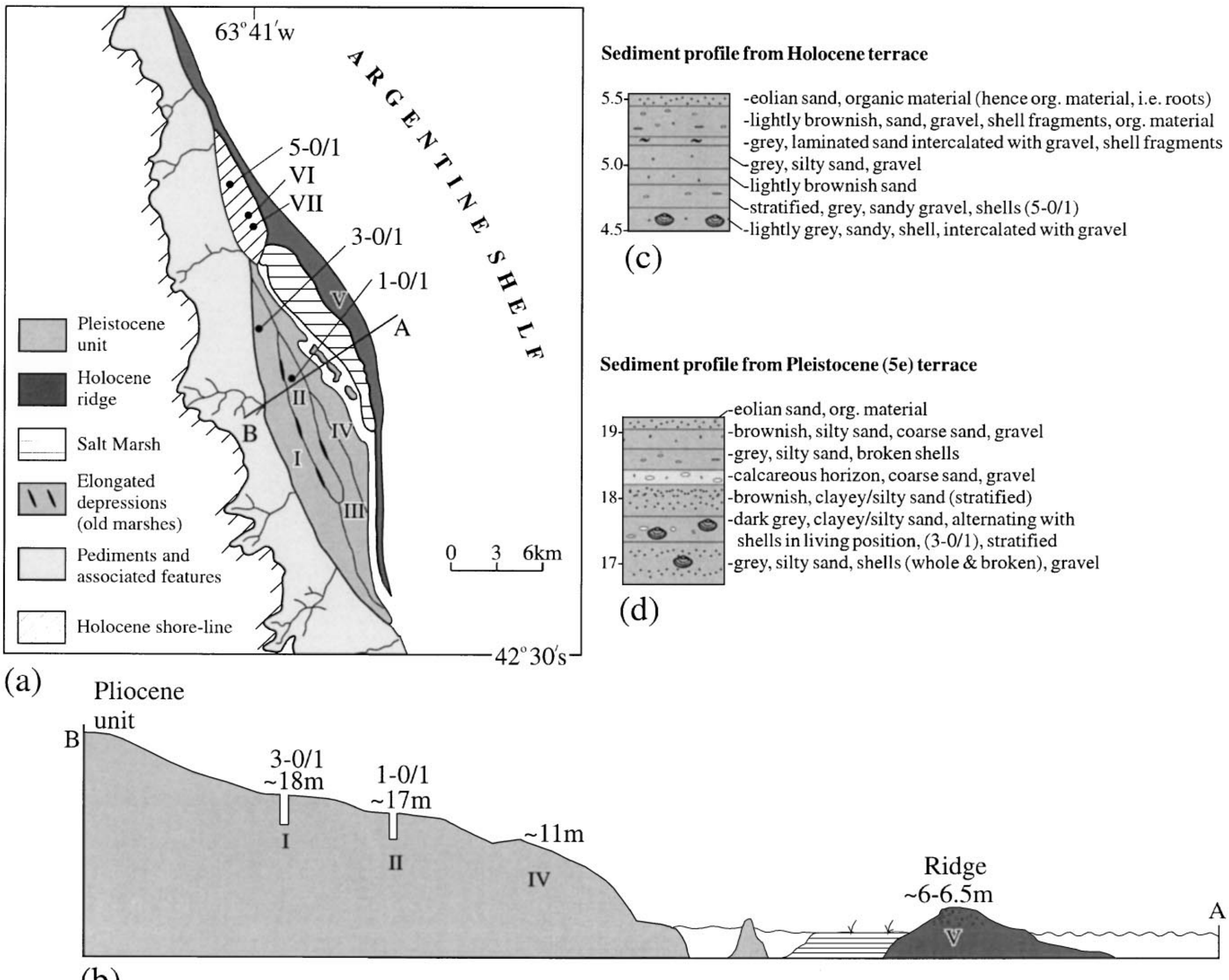

Fig. 2. (a) Geological and site location map of the Quaternary marine shorelines at the Caleta Valdés location (modified from Fasano et al., 1984). (b) Geological cross-section showing morphostratigraphy of the paleo shorelines at the Valdés site. (c) holocene terrace sedimentology. (d) pleistocene terrace sedimentology.

130 km north of the city of Comodoro Rivadávia. There are three terraces, which do not occur in staircase fashion at a single location, but rather are distributed along the coast over a distance of 25 km. At a site 5 km north of the city of Camarones, we sampled an elevated terrace at 5.5–6 m containing fossil molluscs in a dark grey sandy horizon (Figs. 3a and b). On the basis of both the planar morphology and the well stratified sediment of this terrace as well as the results of Th/U dating (see Table 1) we suggest that this terrace was formed during a sea-level highstand at approximately 7000 BP. The ca. 16 m high Pleistocene terrace at this site close to Camarones (Fig. 3a), does not contain any appropriate molluscan fauna. Additional Pleistocene terraces were sampled, one located 12 km south of the city at an elevation of approximately + 17 m amsl and another located 6–7 km south of the city at an elevation of 33–34 m amsl, which was exposed by road-cut.

Based on the well stratified nature of the terrace sedimentology, the Pleistocene wave-cut platform (Fig. 14a) at 17 m amsl, which was sampled approximately 12 km south of the city of Camarones was found, in terms of its lateral continuity and planarity, to be the best-defined terrace in this area. Molluscan fauna, in living position, were found confined to a horizon 20–25 cm thick (Fig. 3c). *Mercenaria* shells in living position extracted from approximately 120 cm depth were found to have a Th/U age of 117,000 ± 5000 BP for sample 3-0/1, Th/U 115,000 ± 9,000 and ESR 110,000 ± 8000 BP for 3–0/2 and, for sample 3–0/3, Th/U 112,000 ± 13,000 and ESR 114,000 ± 9000 BP (Table 1). Therefore this paleo shoreline, which is

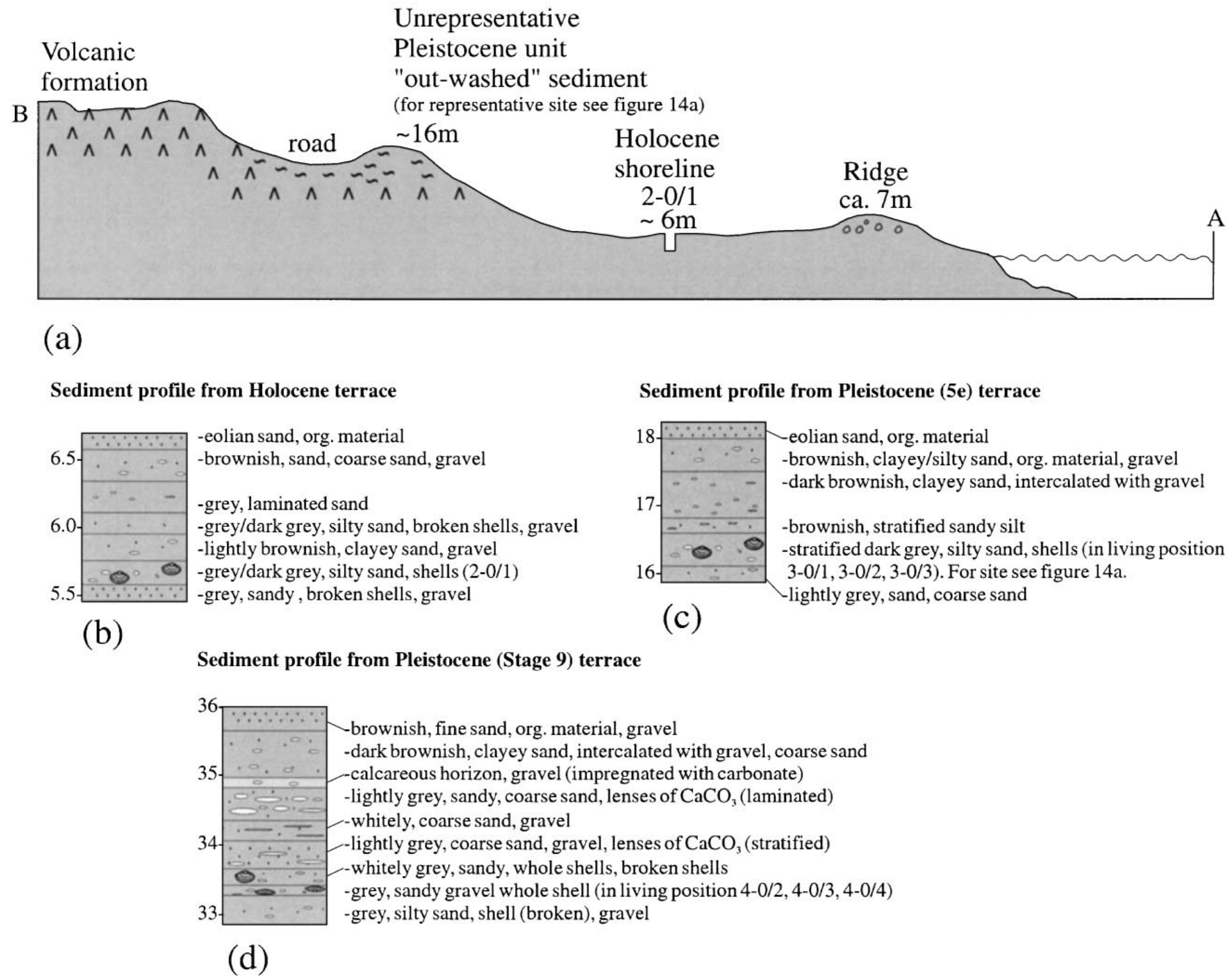

Fig. 3. (a) Geologic cross-section at the Camarones location showing morphostratigraphy of the Quaternary paleo shorelines, (b) Sedimentologic profile of Holocene terrace. (c) sedimentologic profile of Pleistocene terrace (OIS 5e). For morphology of terrace see Fig. 14(a). (d) Sedimentologic profile of Pleistocene terrace (stage 9).

exceptionally well characterized by its paleo sea-cliff, appears to have been formed during the Last Interglacial (OIS 5e). From a terrace at approximately the same elevation at the Camarones site, a mollusc shell (D2417/Pa31a-genus unknown) provided to us by Dr. Schellmann, yielded a Th/U age of 117,000 $\pm$ 6000 BP.

Mytilus shells were also collected in living position from a terrace at 33–34 m elevation. This shell deposit was capped by unconsolidated sediment up to 300 cm thick (Fig. 3d). A part of this extremely thick sediment was displaced by road-cut. Three samples were dated: shell 4–0/2 gave Th/U 309,000 + 50,000/-35,000 BP, shell 4-0/3 gave Th/U 354,000 $\pm$ 45,000 BP and shell 4–0/4 gave Th/U 338,000 $\pm$ 34,000 BP. All of these three shells had relatively high uranium concentrations, but they also have higher thorium isotope concentration (Table 1) (for a discussion of such samples see Appendix A). As a Th/U age of 354 kyr is well beyond the limit of this dating method, or at best within the critical range, this age will be disregarded in further considerations of paleo sea-level. From the Camarones site radiocarbon ages of 6708 $\pm$ 46 BP and 6662 $\pm$ 59 BP were reported for two mussel shells, found in living position, by Schellmann and Radtke (1997). For marine terraces at elevations ranging from 16 to 30 m, they reported ages such as to imply that these terraces represented OIS 5, 7 and 9.

3.3. *Comodoro Rivadávia* (*45°34′–45°47′S*)

Close to the city of Comodoro Rivadávia, Holocene marine terraces are easily recognized both to the north and to the south. One site is approximately 20 km to the north and the other 5 km to the south. At the northern site *Mercenaria* sample RT.SII was extracted in living position from a sandy terrace horizon (Fig. 4b) at 3 m amsl, and was Th/U dated, giving an age

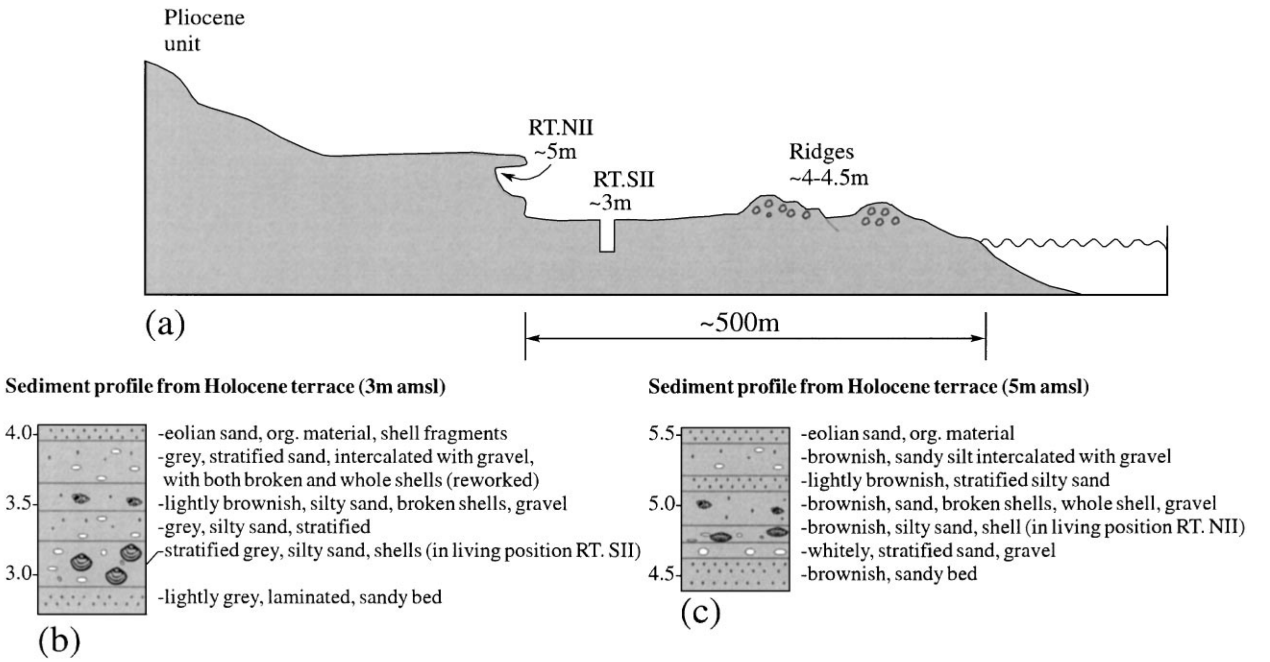

Fig. 4. (a) Morphostratigraphic cross-section of successive Holocene terraces at Comodoro Rivadávia location showing positions of molluscan fauna. (b) Sedimentologic profile of marine terrace at 3 m and (c) at 5 m amsl. For a photograph of the site see Fig. 8(b).

of 3000 ± 1000 BP. The *Mercenaria* sample RT.NII, taken from a higher (+ 5 m) terrace at the same location (Fig. 4a), was Th/U dated to 7000 ± 1000 BP and radiocarbon dated to 6441 ± 70 kyr. The morphology of both features is shown in Fig. 4a (see also Figs. 14b and c), and the sedimentology of the terraces is depicted in Figs. 4b and c.

Five kilometers south of Comodoro Rivadávia, at the Rada Tilly location, the mollusc samples RT 2–0/2 and RT 2–0/3 were collected from a terrace at 4 m amsl (different sites) and dated by thermal ionization mass spectrometry (TIMS), giving ages of Th/U 5870 ± 66 BP for shell RT 2-0/2 and Th/U 3030 ± 27 BP for shell RT 2-0/3. Radiocarbon dating of these samples yielded ages of 2959 ± 55 BP for RT 2-0/2 and 2959 ± 40 BP for RT 2-0/3. Consequently, we adopt an average age of 3000 to 3500 yr BP for both paleo-beach horizons. At the Rada Tilly site we also investigated a feature at approximately 6 m amsl. This deposit was replete with out-washed fossil mussel shells in a very loose sediment consisting of sand, boulders and no matrix material (Fig. 14d). Since these mollusc shells were assumed not to be in their original habitant, but rather deposited (e.g. by spring tides or surges) and therefore not representative, no sample was taken. For a sequence of five Holocene paleo shorelines at approximately 1 to 5 m amsl, a detailed investigation was previously conducted 20 km north of Comodoro, at the Bahía Solano site, by Codignotto et al. (1990). Their reported radiocarbon ages ranged from 1045 ± 60 to 6310 ± 100 BP.

3.4. *Caleta Olivia* (*46°29′–46°48′S*)

The city of Caleta Olivia is located 75 km south of the Comodoro Rivadávia site. At this location we investigated marine terraces at approximately 7, 14 and 16–17 m amsl, respectively (see Figs. 5a and c). These terrace deposits are approximately 10 km north of the city of Caleta Olívia and were found to consist of silt, sand, gravel and abundant marine fossil remains, but not all exposures (e.g. the 7 m terrace in Fig. 5c) yielded representative mollusc shells. Approximately 6 km south of the city of Caleta Olivia, at Bahía Langara (Fig. 5b), a rich shell-bearing layer of grey colored silty sand (Fig. 5d) at 5 m amsl was Th/U dated to 5000 ± 1000 BP (sample 7-1/1). A radiocarbon age of 5380 BP has been reported by Schellmann (1995) from a beach horizon at the Caleta Olivia location. This would confirm our result that the 5 m level at the Caleta Olivia site corresponded to a marine stillstand around 5000 BP. Specimen D 2619/Pa 72, taken from a terrace at approximately + 7 m elevation, was also made available to us by Dr. Schellmann; it had a Th/U age of 8000 ± 2000 BP. From approximately the same elevation, a similar radiocarbon age of 8090 BP was reported by Schellmann and Radtke (1997). We relate both dates to the maximum Holocene transgression.

Also recognized at this location was a terrace at approximately 14 m amsl, which was overlapped by a feature at 16–17 m amsl (Fig. 5c). A section through this terrace was freshly exposed by road construction. The

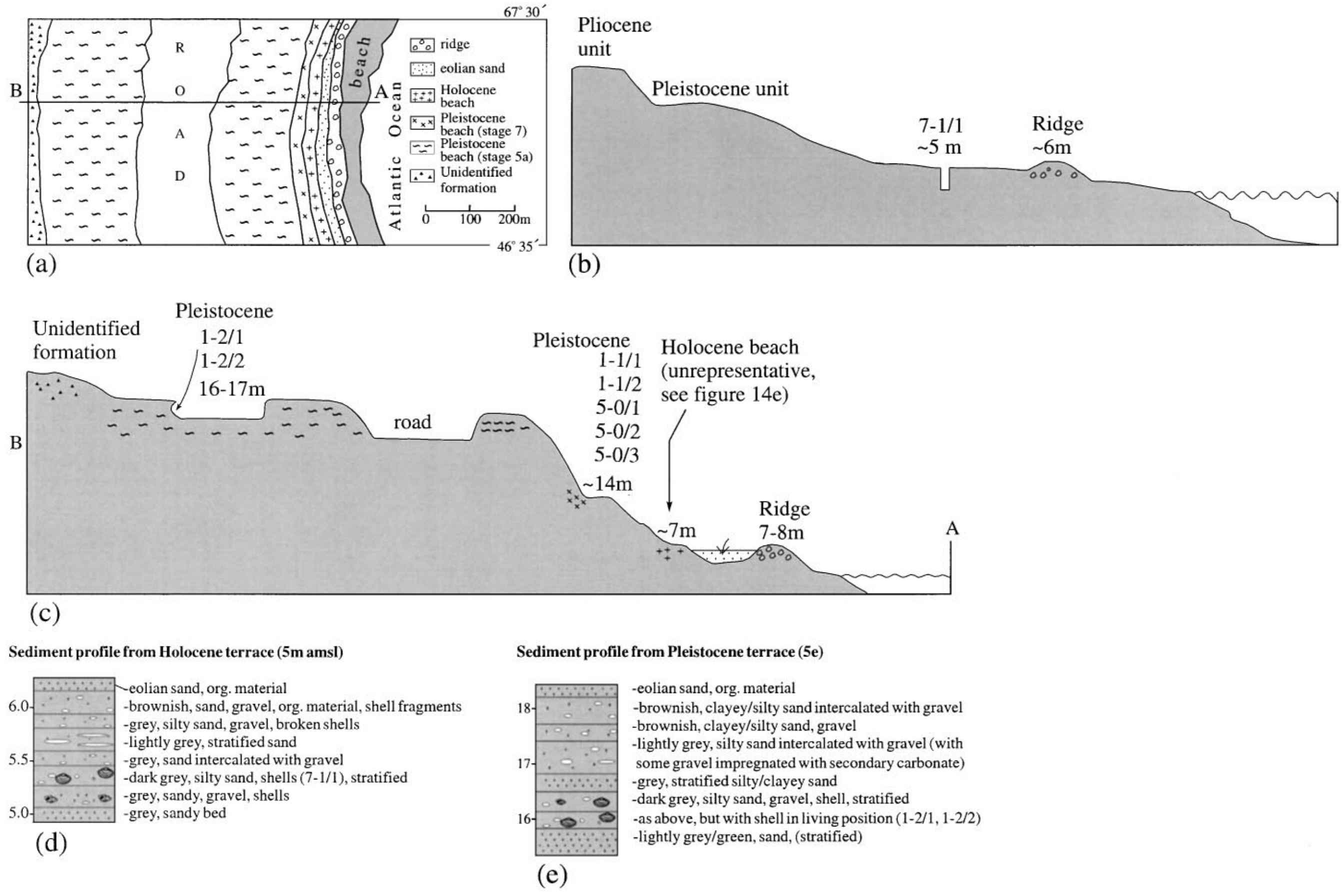

Fig. 5. (a) Geological structure of Quaternary marine deposits at the Caleta Olivia location. (b) Geological cross-section of Quaternary marine deposit at the Bahía Langara location. (c) Morphostratigraphic cross-section through the Quaternary paleo shorelines. Fig. (d) presents the sedimentologic profile taken from a Holocene terrace at Bahía Langara, i.e., 6 km south of the Caleta Olivia location. (e) Is the sedimentologic profile for terrace at ca. 16 m amsl.

overlapping nature of the terrace suggested an unconformity in the terrace sediments, and when examining the remnants of the shelly terrace material, discussion arose concerning the indurated degree of consolidation of the sediment. Since specimens were not collected in their original habitats, but taken from materials cut by bulldozer, they appear separately in Table 1 and will not be used in what follows. Nevertheless we collected five fossil mollusc shells from this section and analyzed them both by sequential leaching and by one-step leaching. The results obtained for these specimens were as follows: sample (1-1/1) Th/U 239,000 $\pm$ 28,000 BP, (1-1/2) Th/U 158,000 $\pm$ 13,000 BP, (5-0/1) Th/U 200,000 $\pm$ 19,000 BP, 5-0/2 Th/U 208,000 $\pm$ 12,000 BP (ESR dated 222,000 $\pm$ 36,000 BP) and (5-0/3) Th/U 274,000 $\pm$ 30,000 BP (see also Table 1). If we reject the last age (274,000 $\pm$ 30,000 yr) as an outlier, then the average age is 201,000 $\pm$ 18,000 BP. These results may therefore provide paleoclimatic information correlated to OIS 7.

A Pleistocene terrace at 16–17 m amsl (Fig. 14f) was found at the same site and is characterized by a mollusc bearing horizon 20 cm thick (Fig. 15a). An excavation in the top of this prominent terrace (2–3 m deep, 20–30 m in size) was made for gravel production, which allowed the sedimentology to be investigated (Fig. 15a) . By lateral removal of sediment from its surface, mussel samples 1-2/1 and 1-2/2, in living position, were collected. The Th/U age of sample 1-2/1 was determined to be 120,000 $\pm$ 6000 BP by alpha-spectrometry and 115,000 $\pm$ 3000 BP by TIMS. An ESR age for this specimen was also determined to be 132,000 $\pm$ 14,000 BP. Sample 1-2/2 yielded a Th/U age of 137,000 $\pm$ 7000 BP by alpha spectrometry. This terrace can be correlated, based on terrace stratigraphy and sedimentology (Fig. 5e) as well as apparent age, with a further marine terrace at approximately 17 m amsl (not shown), which is 11 km north of the city of Caleta Olívia. This site was exposed by road-cut, and *Mercenaria* samples 1-3/1 and 1-3/3_b were taken and these yielded ages of Th/U 133,000 $\pm$ 15,000 and 143,000 $\pm$ 11,000 BP, respectively.

3.5. Punta Mazarredo (47°03′–47°07′S)

The Mazarredo site is located approximately 130–150 km southeast of the city of Caleta Olivia. A pair of wave-cut platforms at approximately 6 m amsl and 16–17 m amsl were found at this location (Figs. 6 and 15b). In addition, a regressive beach deposit at 5 m amsl, partly eroded by a gulch to 1 m depth, was also recognized. The terrace at 6 m amsl consists of well-stratified silty sand containing molluscs in living position (Fig. 6b). A fossil *Mytilus* sample (1-2/1_b) from this horizon gave a Th/U age of 8000 $\pm$ 2000 yr BP. The shell-bearing horizon of the regressive terrace at 5 m amsl contained broken shells in addition to whole shells, which indicates reoccupation. Having removed the reworked sediment from the top of this terrace we extracted *Mytilus* sample 2-0/1 and Th/U dated this to obtain an age of 5000 $\pm$ 1000 yr BP. Two other specimens from the terraces at 5 and 6 m amsl, which were originally radiocarbon dated to 4050 $\pm$ 90 yr (sample ID Maz 24) and 5233 $\pm$ 90 yr (sample ID Maz 27), were made available to us by Dr. Codignotto. Th/U dating yielded an age of 8000 $\pm$ 2000 BP for sample Maz 24 and $\leqslant$ 1000 BP for Maz 27. We interpret this lower Th/U age of Maz 24 as deriving from an incorrect evaluation of the thorium peak in the spectrometer or some other effect in the alpha spectrometer. The fact that Th/U and radiocarbon dating have not been performed on the same mollusc shells may also have contributed to the apparent discordance. However, we assign an age of 4000 to 5000 yr BP for this raised beach.

The wave-cut platform at 16–17 m amsl was also sampled and the two specimens of *Mytilus* collected from this site gave Th/U ages of 104,000 $\pm$ 10,000 BP and 137,000 $\pm$ 7,000 BP. *Mytilus* sample 5-2/3 (in living position) was taken from a separate excavated hole on the same terrace (see Fig. 6a) and dated by Th/U assay as 108,000 $\pm$ 10,000 BP. The morphology of the investigated terraces is presented in Figs. 6a and 15b, whereas the terrace sedimentology is described in Figs. 6b and c.

3.6. San Julian (49°00′–49°27′S)

We recognized up to three marine terraces in this area spread over a distance of 35 km. One km north of the city of San Julian, in the Bay of San Julian, a beach deposit at 7–8 m amsl was found to consist of broken mollusc shells and polished whole shells (Fig. 15c). This feature was interpreted as being a "storm-beach" deposit and, therefore, irrelevant for our purposes, so no samples were taken. 2 km south of the city we found a pair of wave-cut platforms at elevations of 7–7.5 m and 16–17 m (Figs. 15d and 7a and b). A gravel pit (15–30 m in size and 2 m deep) had been excavated on the top of the lower terrace (Figs. 7b and 15e); from this a *Mytilus* sample 2-0/1 was collected from a horizon at 40 cm depth. This sample gave a Th/U age of 8000 $\pm$ 2000 BP. This age correlates well with others derived for Holocene raised beaches in southern Patagonia. Mussel sample (Y-180) taken "in situ" by Auer (1974) from a depth of 30 cm below a shoreline accumulation at the San Julian site yielded a radiocarbon

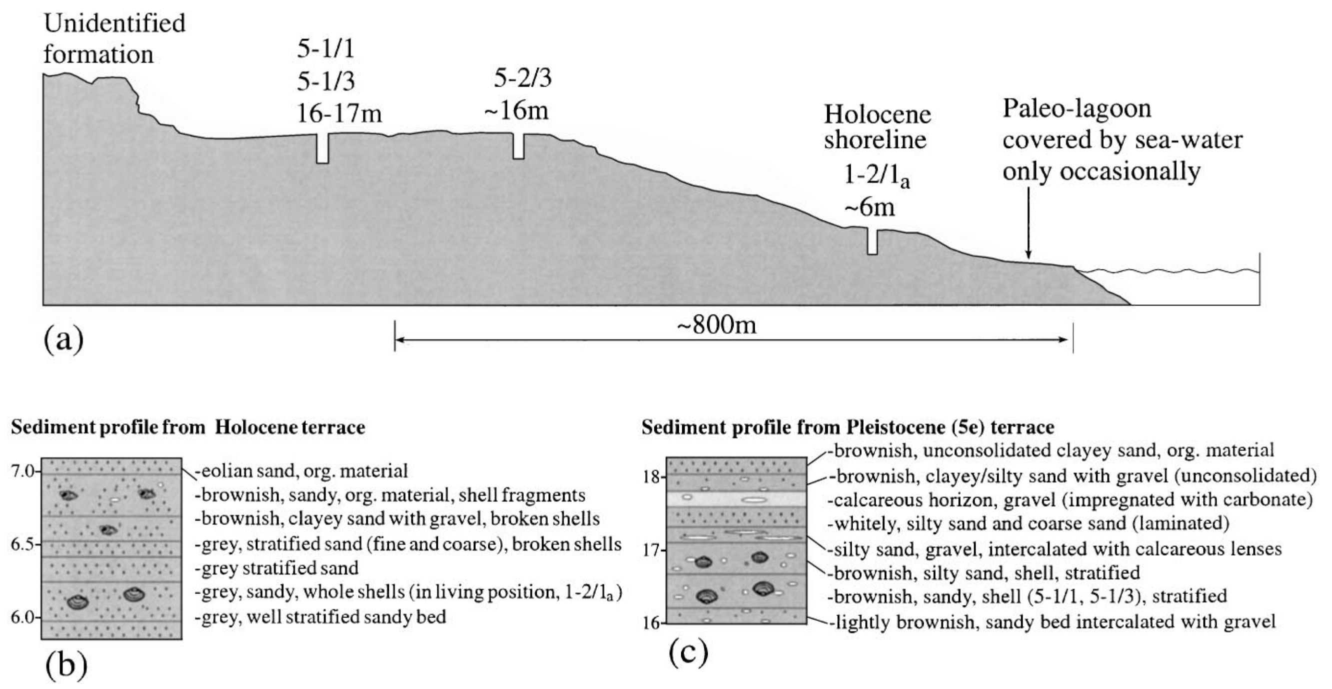

Fig. 6. (a) Morphostratigraphic cross-section through the Quaternary paleo shorelines at the Punta Mazarredo location. (b) and (c) presenting the sedimentologic profiles of terraces at 6 m and 16 m amsl.

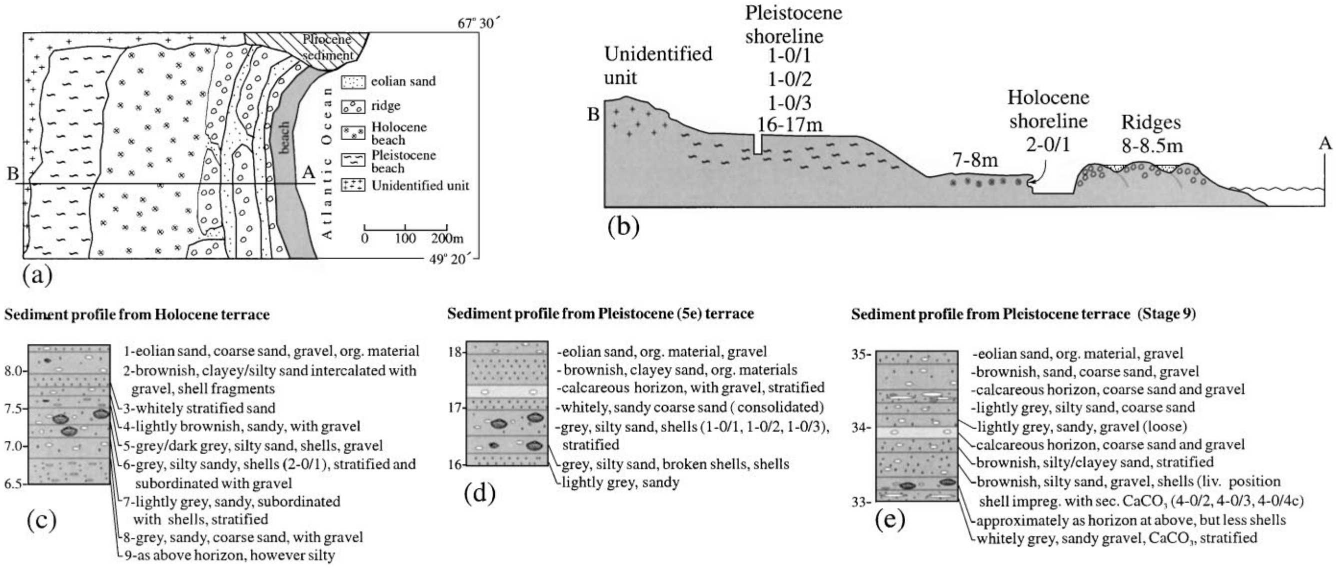

Fig. 7. (a) Geological structure of the Quaternary marine deposits at the Bahía San Julian location. (b) Morphostratigraphic cross-section through the Quaternary paleo shorelines. Figs. (c)–(e) presenting the sedimentologic profiles of beach deposits at + 7, 16–17 and 33–34 m amsl.

age of 8110 ± 100 BP. This provides independent evidence that our Th/U age determined for sample 2-0/1 is valid.

Having excavated a pit at the top of the higher wave-cut platform (16–17 m amsl) and investigated the terrace sedimentology (Fig. 7d) *Mytilus* samples 1-0/1, 1-0/2 and 1-0/3 were collected from a depth of approximately 120 cm and Th/U dated. The sequentially analyzed shell 1–0/1 yielded, for the innermost portion, an age of 115,000 ± 9000 BP, Sample 1-0/2 yielded a Th/U age of 134,000 ± 14,000 and an ESR age of 99,000 ± 9000 BP. Specimen 1-0/3 gave a Th/U age of 128,000 ± 6000 BP.

35 km north of the city of San Julian a broad, planar raised terrace surface (Fig. 15f) occurs at an elevation of approximately 32–34 m amsl. This marine terrace forms the base of the only North–South running road in Patagonia. From a 130–150 cm hole excavated in the terrace surface, we extracted three *Mytilus* shells, in living position, and obtained the following Th/U ages: 342,000 ± 54,000 BP, $\geqslant$ 355,000 BP and 240,000 ± 38,000, respectively. As a Th/U age $\geqslant$ 355,000 yr is well beyond the range of Th/U dating technique, we reject this age and an average age for the remaining two specimens is 291,000 ± 46,000 years BP.

4. Pleistocene terrace ages and the implied history of tectonic uplift

In the previous subsections we described the Quaternary marine terraces that were sampled to constrain paleo shoreline age and elevation and presented a set of internally consistent measurements for the ages of the various features. These aspects were illustrated in Figs. 2–7. From the previous section it is clear that terraces corresponding to OIS 5 appear in Patagonia at 16–17 m amsl. However, to the north in Brazil comparable features are found at approximately 8–10 m amsl (see Fig. 11b) (Clark and Bloom, 1979b; Barbosa et al., 1986); we believe on this basis that the beach risers in Patagonia have been tectonically uplifted. The coral reef at 7–8 m amsl of Ouvéa, Loyalty Island, New Caledonia has a Th/U age of 117,000 ± 6000 BP (Marshall and Launay, 1978), corresponding to a time when sea-level stood slightly higher than present. By assuming an Eemian eustatic sea-level that stood 5–6 m higher than present (Chappell, 1983; Chappell and Shackleton, 1986; Shackleton, 1987; Chappell et al., 1996) we would expect, on relatively stable coastlines, shorelines of Last Interglacial age to be deposited at 7–10 m amsl on most continental margins (cf. Clark and Bloom, 1979b). If Atlantic Patagonia were tectonically stable the Eemian paleo shorelines found at 16–17 m amsl ought to be at approximately 7–8 m, as in New Caledonia, or at approximately 7–10 m, as in Brazil. Thus the shorelines in our study area must have been tectonically uplifted by approximately 10 m since the Last Interglacial. Here, we infer the Pleistocene uplift history of the study area based on uplift calculations for each location using a subset of Th/U data in which there is either agreement with ESR ages or internal consistency according to the "multi-sample" procedure or based on the use of data with the

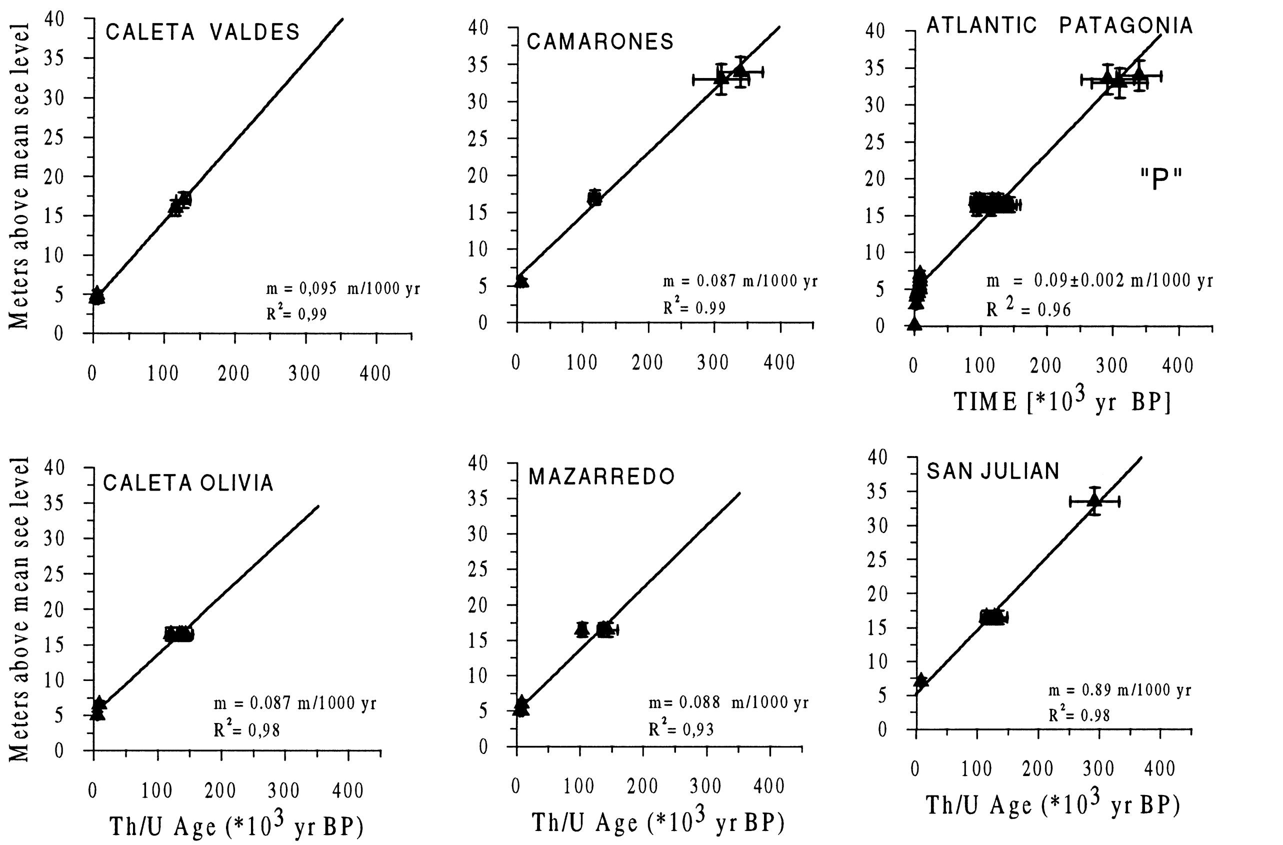

Fig. 8. Height–age plots based upon the radiometric data for each site of study area. To infer the Quaternary uplift history we calculated the elevation of each of the beach horizon above present sea-level as a function of its radiometric age. The slope (m) of linear trends represents a constant uplift rate for each individual location along the Atlantic Coast of Patagonia in meters per 1000 yr. Diagram "P" has been generated by combining all data from all sites together to obtain an uplift rate for the entire coast of Argentinian Patagonia. The vertical and horizontal bars denote estimated errors in altitude and standard errors of ages, respectively.

^{230}Th/^{232}Th ratio > 30. In support of the general validity of ^{230}Th measurements, ^{230}Th/^{232}Th activity ratios of $\geqslant 25$ were reported by Rutter and Catto (1995) indicating that contamination of shells by extraneous ^{230}Th cannot be responsible for the ^{230}Th content (for details see Appendix A). The age–height data for all terraces is plotted in Fig. 8. The inferred Pleistocene uplift rate was found to be consistent for all locations since the middle Pleistocene and derived uplift rates were found to be between 0.088 mm/yr (e.g. at the Mazarredo location) and 0.095 mm/yr, for example, at the location of Caleta Valdés (Fig. 8). In constructing curve "P", also presented in Fig. 8, we used all data to compute a regional tectonic uplift rate for the entire study area. Based on these analyses we have obtained a relative regional land uplift rate of 0.09 ± 0.002 mm/yr for the past 300 kyr. The long-term uplift rate for this passive plate boundary region is small and entirely plausible on a priori grounds.

From a preliminary examination of Fig. 8, we conclude firstly that terraces of the same age are found at approximately the same elevation. Secondly the climax of the transgression of the sea during OIS 5e (120,000–130,000 yr BP) is now found in every location at 16–17 m elevation. Thirdly, mollusc-bearing horizons deposited at approximately 300,000 yr BP now have risen to 33–35 m amsl. Taken together, these data provide no evidence of any significant short-timescale tectonic contamination of the sea-level signal in the study area. If such tectonic (perhaps coseismic) events had occurred between each period of terrace formation (such as on the Huon Peninsula of Papua New Guinea, e.g. Ota et al., 1993), it would be impossible to follow these age–height-equivalent features over a distance of more than 1000 km along the coast. Successive paleo shorelines represent periods of relative tectonic stability during which the terraces were deposited or eroded. Construction occurred while the continental margin was being slowly uplifted and while eustatic sea-level was changing due to glaciation and deglaciation and while north–south tilting was also occurring.

Of particular interest is the absence of paleo shorelines deposited during OIS 7 (approximately 210,000 yr BP). The absence of terraces of this age may be understood on

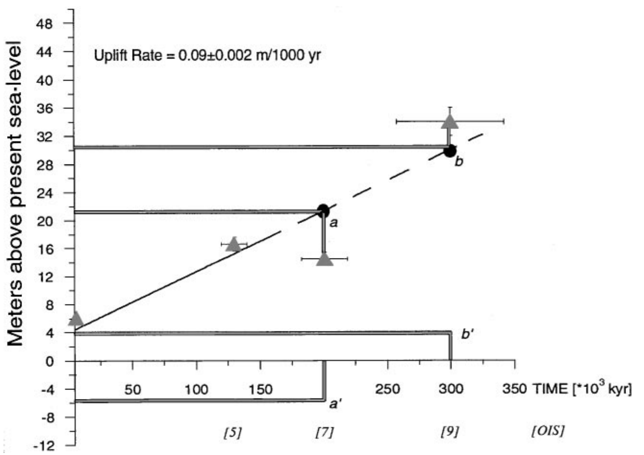

Fig. 9. The model employed to check the paleo sea-levels for stages 7 and 9. The solid line is the linear fit for stages 1 and 5. From these ages we calculated an uplift rate of 0.09 mm/yr^1 and extrapolated it back to stage 9 (dashed line). Points (a) and (b) in the extrapolated part of the curve indicate where the terraces of stages 7 and 9 should appear if both of these sea-level high stands were similar to the present sea-level. These terraces should appear at elevations of $+21$ and $+30$ m, respectively. The difference between the predictions of this model and the observations, shown as points (a′) and (b′) is evident. See text for more details.

the basis of precisely this slow rate of tectonic uplift of 0.09 ± 0.002 mm/yr since this uplift rate would be insufficient to have isolated the terrace from reoccupation during the subsequent transgressive OIS 5e when sea-level rose by considerably more than in the previous (OIS 7) interglacial. Because the time interval between the climax of OIS 7 at 200,000–220,000 yr BP and the culmination of the subsequent interglacial at 120,000–130,000 yr BP was 80,000–90,000 yr, the terrace occupied during OIS 7 could have been uplifted during this interval only by 7 to 8 m (85 kyr $\times 0.09$ m/kyr = 8 m). Since our data indicates (see Table 2) that the position of sea-level was approximately 6 m higher than present during the Last Interglacial (see also Chappell et al., 1996; Bard et al., 1996), the terrace occupied during the Penultimate Interglacial (OIS 7) was at precisely the same elevation as that at which the 5e terrace was deposited. It is likely, therefore, that the terraces of OIS 7 were exposed to the sea in OIS 5e and that their shell materials were reworked and overtopped during the higher eustatic sea-level stand of OIS 5e.

The Th/U and ESR dates for the reworked mollusc shells found in our Sangamonian (OIS 5e) terraces provide direct support for this scenario. Shell debris, consisting of fractured and whole but strongly polished mollusc shells, occurred in most locations and in more or less the same horizon and elevation where terraces of OIS 5e were found (16–17 m amsl). ESR results reported by Rutter et al. (1990) from a location south of the Camarones site (the Bahía Bustamante location, see Fig. 1a) gave ages ranging from 116,000 yr to $\geqslant 195{,}000$ yr BP. For other sites, for example at Caleta Valdés, their D/L ratios of aspartic acid suggested pre-last interglacial ages while ESR gave ages for the last interglacial. Analysis of samples collected from the Bustamante site (not discussed in detail) confirm this scenario. Again, this apparent incongruity may be understood as a consequence of the low tectonic uplift rate and the tendency for reoccupation of OIS 7 terraces by later sea-level positions. This accords with the observations made by Leonard and Wehmiller (1992) and with the results of Radtke (1987) for Coquimbo Bay in Chile which suggested that OIS 7, if present, is found on the lower terrace together with the materials of OIS 5e. At Sumba Island, Indonesia, where there is also a low tectonic uplift rate (< 0.5 m/ky), paleo shorelines belonging to OIS 7 are also missing (Bard et al., 1996). Bard et al. (1996), in discussing the impact of an increase of the uplift rate in their numerical model, observed that this paleo-shoreline would outcrop if the uplift rate exceeded a rate of 0.1 m/kyr for a significant period, i.e. between OIS 7 and OIS 5e.

By knowing the rate of uplift and having absolute age estimates for the paleo shorelines as well as for present terrace elevations, it is possible to reconstruct the Quaternary sea-level in the study area, as has been demonstrated for Barbados (Matthews, 1973); Papua New Guinea (Chappell and Shackleton, 1986) and Sumba Island, Indonesia (Bard et al., 1996). Although Substage 5e sea-level was 6 ± 2 m higher than present sea-level (Matthews, 1973; Ku et al., 1974; Marshall and Thom, 1976; Bloom and Yonekura, 1985 and Chappell et al., 1996), there is less certainty concerning the position of earlier sea-levels. The coral-based sea-level record from the Huon Peninsula, Papua New Guinea, indicates a paleo sea-level of approximately -5 ± 8 m to -6 ± 8 m for OIS 7, and one of approximately $+4 \pm 2$ for OIS 9 (Chappell and Shackleton, 1986), which is in accord with results obtained by Bard et al. (1996).

A comparison between calculated paleo sea-levels from this study with those in Chappell (1983), Chappell and Shackleton (1986), Shackleton (1987) and Chappell et al. (1996) is outlined in Table 1. For the Last Interglacial (OIS 5e) we have 15 Th/U age data from terraces at 16–17 m elevation (Table 1). These ages range from 104,000 BP to 143,000 BP and taken at face value suggest a range of different Sangamonian sea-levels. This interpretation might be valid for coral-based records, but, in general, not for the fossil molluscs. However, the scatter of such data is common not only for fossil shells, but can also occur in some circumstances for coral-based records. Seven dates, ranging from 142 ± 21 to 93 ± 14 kyr BP, for example, were reported for a Sangamonian terrace by Pirazzoli et al. (1993). Nevertheless, we believe that mollusc dates in this study represent the same sea-level high-stand, namely the culmination of the Last Interglacial

(OIS 5e), as they were all collected from terraces which are currently at 16–17 m amsl. Because of open system effects some of these samples have younger ages as is reasonable. Considering the frequency of a given age in our data-set, the culmination of the Sangamonian sea-level highstand in Patagonia occurred 120–130 kyr BP. An average age for 15 samples is 130,000 yr BP. This age compares with estimates based on TIMS measurements on corals from Sumba Island, which range from 119,000 to 132,000 yr BP (Bard et al., 1996). Furthermore, based upon corals, an age of 120 kyr was assigned by Chappell et al. (1996) for the climax of the Last Interglacial (comparable to OIS 5e in deep sea cores).

We will therefore employ the average value of $130{,}000 \pm 10{,}000$ yr for OIS 5e, although the meaningfulness of such averages, which have also been performed by others (e.g. Fairbanks and Matthews, 1978) is open to question. For the pre-5e Interglacial (OIS 7), because of the relative consistency among Th/U ages and of the concordance between Th/U and ESR results, we employ the data presented in Table 1 from the Caleta Olivia location (14–15 m amsl). We have four representative ages for this paleo beach deposit, which range between 158 and 239 kyr (158, 200, 208 and 239 kyr) averaging $201{,}000 \pm 18{,}000$ BP. For terraces at 33 to 35 m elevation there are five Th/U data, ranging from 240 ± 35 to 342 ± 54 kyr (240, 308, 309 and 342 kyr). For the purpose of calculating eustatic sea-level we will employ a mean of $300{,}000 \pm 42{,}000$ yr from all these ages for stage 9.

To determine eustatic sea-level in the study area we compute $S = (H - U)$ with $U = U^* \times t$ where S is sea-level, H is the height of the terrace (amsl), U^* is the calculated uplift rate, U is the total tectonic uplift in the study area and t is the age of the sample. The results for paleo sea-level relative to the present sea-level for each stage are listed in Table 2. As evident from Table 2, we infer an eustatic sea-level for the Last Interglacial of $+6 \pm 2$ m amsl (the calculated sea-levels range from $+5.4$ to $+6.4$ m due to the range of terrace elevations of 16–17 m). This value is surprisingly close to previous estimates of eustatic sea-level during the Last Interglacial reported for many regions, for example based on coral records for Indonesia (Bard et al., 1996); for Barbados (Fairbanks and Matthews, 1978); for Papua New Guinea (Chappell and Shackleton, 1986, Chappell et al., 1996) and for Australia (Bloom and Yonekura, 1985). The results suggest a paleo sea-level of -4 ± 2 m for the Penultimate Interglacial (OIS 7) and of approximately $+6 \pm 2$ m for OIS 9 (due to the range of terrace elevations of 33–35 m, the calculated sea-levels are $+7.4$ to $+5.4$ m). In our error estimates of ± 2 m for all paleo sea-levels we have included the influence of tidal fluctuations, uncertainties in the relationship of sample elevation and former mean sea-level, and uncertainties between individual terraces in sequences as well as overall errors due to height measurements as the square root of the sum of the squares of the errors. These values, considering all mentioned errors and uncertainties in the data, are comparable to the eustatic sea-level data reported by other investigators (see Table 2).

To verify our interpretations of the pre OIS 5e terraces we employed our simple model (Fig. 9) to predict the elevations of these terraces. For this purpose we used the Holocene results as well as the data from the Last Interglacial terraces as these ages seem most reliable and are less complicated to interpret than those derived from the terraces of OIS 7 and 9. The uplift rate is fixed by these two datasets (solid line on Fig. 9) and this linear fit may be extrapolated to make a prediction of the elevation expected for the terraces belonging to the OIS 9 (dashed line on Fig. 9). If our calculated uplift rate based upon the Holocene and OIS 5 data is correct, terraces of OIS 7 and of OIS 9 should appear close to the extrapolated linear fit (points a and b in Fig. 9). If during OIS 7 the position of sea- level were similar to present sea-level, the paleo shorelines occupied by the sea at that time would now

Table 2
Magnitudes of calculated Pleistocene paleo sea-levels based on coral data, of Chappell (1983), Chappell and Shackleton (1986), Shackleton (1987), Chappell et al. (1996) and based upon mollusc shells for Atlantic Patagonia (this paper)

Isotope Stage or substage	Age ($\times 10^3$ yr BP) (after authors listed above)	r.s.l. (m amsl)[a] (after authors listed above)	Age ($\times 10^3$ yr BP)[b] (this report)	r.s.l. (m amsl)[c] (this report)
5e	124	$+6 \pm 2$	130	$+6 \pm 2$
7	200	-5 ± 8 to -6 ± 8 [d]	201	-4 ± 2
9	320	$+4 \pm 2$	300	$+6 \pm 2$

[a]Relative sea–level derived from the deep sea records combined with the New Guinea coral terrace records (cf. Chappell and Shackleton, 1986).

[b]For relative sea-level determinations we employed an average of 130 kyr for OIS 5e, derived from 15 representative samples ranging from 104 to 143 kyr. For OIS 7 a mean value of 201 kyr from four age measurements was used and for OIS 9 an average of 300 kyr derived from four representative ages was used.

[c]Paleo sea-level derived from Patagonian marine mollusc shells, this paper. Errors estimations included errors in depth of water in which shells grew, altitude estimates (or altimeter accuracy) and uncertainty in the tide fluctuations as r.m.s. values.

Error estimates for stages 5e and 7 terraces: $[(0.5)^2 + (1.0)^2 + (1.0)^2]^{1/2} = 1.5$,

Error estimates for stage 9 terraces: $[(0.5)^2 + (2.0)^2 + (1.0)^2]^{1/2} = 2.2$.

[d]r.s.l. was measured using the curves derived by Chappell and Shackleton (1986).

appear at approximately + 21 m elevation (according to the model). Similarly the terraces of OIS 9 should now appear at approximately + 31 m elevation. In the field we actually observed the terraces occupied during OIS 7 at approximately + 14 m amsl rather than at + 21 m. Similarly at the Caleta Olivia site the beach risers of stage 9 were not found at + 31 m but rather at approximately + 34 m amsl. Hence we believe that sea-level must have been lower approximately 210,000 yr BP (during OIS 7) since the land occupied by the sea at that time does not appear at + 21 m but rather at a much lower elevation, due to the small uplift rate for this region. Also, during its occupation of the beach associated with stage 9, sea-level must have been higher than present since terraces were not found at + 31 m but rather at higher elevation. Thus from these differences we infer the position of eustatic sea-level for each of OIS 7 and 9. As evident in the model (points a' and b' in Fig. 9) the model inference of paleo sea-level for OIS 7 is approximately − 5 m whereas it is approximately + 4 m for the position of eustatic sea-level during OIS 9. On the other hand, we have calculated eustatic sea-level of -4 ± 2 m for stage 7 and of $+6.4 \pm 2$ m for stage 9. The results obtained from both methods of inference are much the same, with the small differences lying well within the limits of error. This inference, despite large error-bars on the ages derived for stages 7 and 9, suggests that the uplift rate we have inferred is correct.

5. Holocene sea-level ages and uplift history

As previously discussed, raised Holocene marine terraces were found all along the coast at different elevations ranging from 3 to 7 m amsl. At some sites there were up to three representative terraces recognized (e.g. the Comodoro Rivadávia site, Table 1). At other sites many such Holocene features consisted mainly of "storm beach" deposits or ridges and were therefore not useful. On this selective basis we have 15 ages that may be employed to define a Holocene sea-level curve, of which we employ only 12 data with a $^{230}Th/^{232}Th$ ratio > 30 (see Table 1). A plot of these data from all of the investigated sites is shown in Fig. 10a. The plot includes two age determinations from Codignotto et al. (1990), both of which have a ^{14}C age of approximately 1000 yr for mollusc shells collected at approximately 2 m amsl at Bahía Solano (ca. 20 km north of Comodoro Rivadávia). Based upon the data from along this coast there are several sea-level highstands, which appeared at 7 to 2 m amsl with ages of 8000 BP, 7000 BP, 5000 BP, 4000 BP, 3000 BP and 1000 BP, respectively. We interpret these successive ages to reflect spasmodic Holocene regressive phases of Patagonian sea-level. However, we believe that the 7000–8000 BP terraces mark the peak of the Holocene highstand, and that sites younger in age represent regressive sea-levels. Radiocarbon ages of 6000–8000 BP from northern Argentina (i.e. from the Province of Buenos Aires) were also assigned to the peak of the climatic optimum during the Holocene by Codignotto and Aguirre (1993).

From the data in Fig. 10a paleo-shorelines occupied at approximately 8000 yr BP occur at different elevations along more than 1000 km of the coastal study area (Fig. 10a). These features are lower in northern Patagonia and increase in elevation southward. This trend is also valid for the other beaches investigated and is not an error of the altimeter measurements. This southward

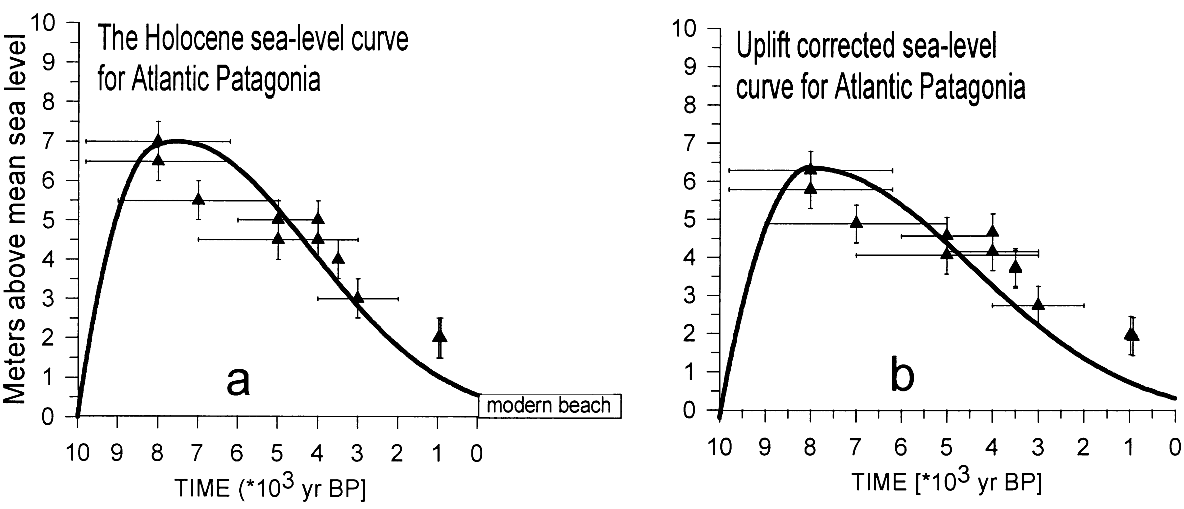

Fig. 10. Curve (a) shows the Holocene relative sea-level history for Patagonia based on Th/U data. Samples of equal age and different elevations were taken from different sites along more than 1000 km of the coast-line. The elevations of shorelines of the same age in the south are higher than their counterparts in the north, a consequence of the south–north tilting of the coast (see text and Fig. 11 (a). Curve (b) shows the Holocene sea-level history, where uplift is corrected for the inferred uplift rate of 0.09 mm/yr^{-1} of the Patagonia study area. The vertical and horizontal bars denote estimated errors in altitude and standard errors of the ages, respectively.

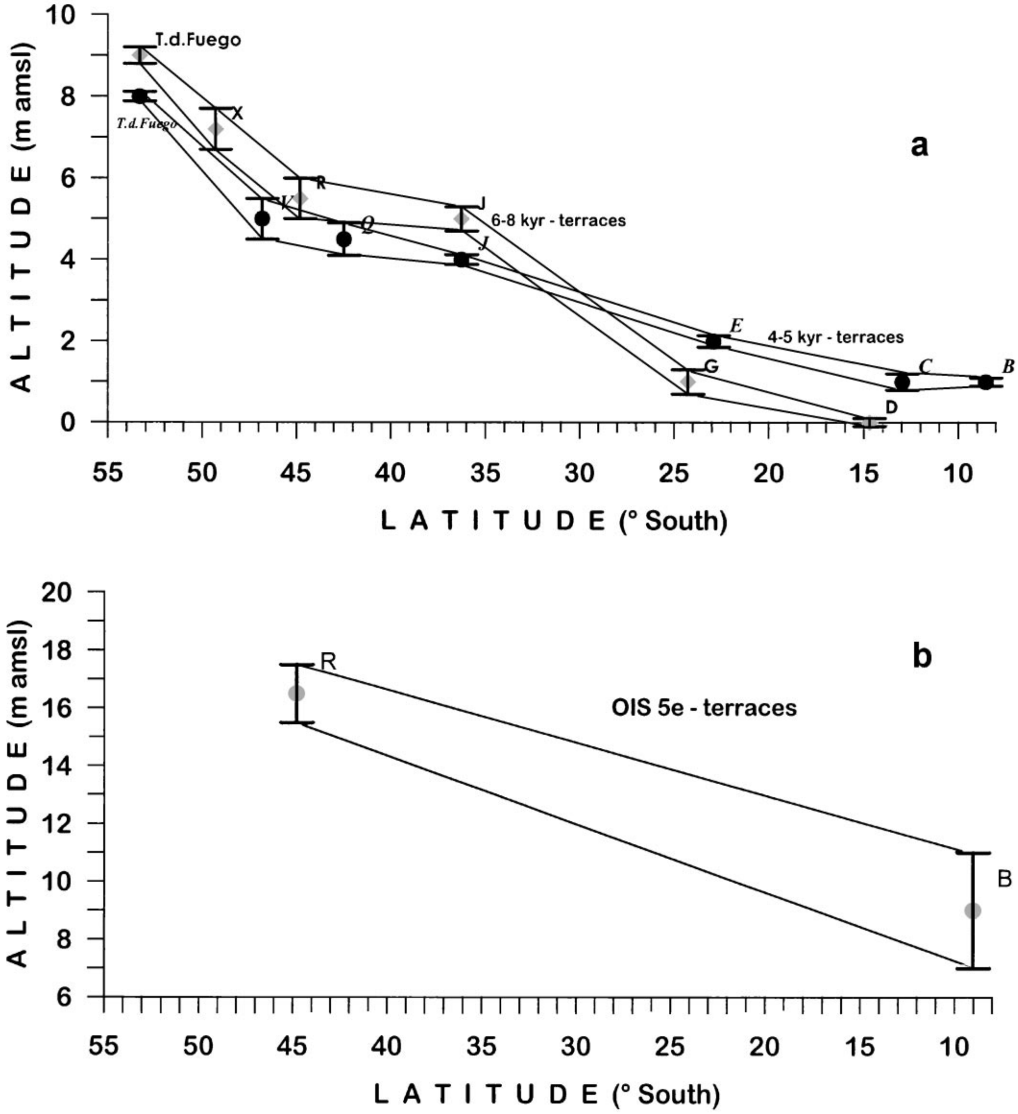

Fig. 11.(a) Emerged Holocene marine terraces along the eastern South American continent. Holocene sea-level changes in Brazil (B = Delibrias and Laborel, 1971; C, D and E = Martin et al., 1987; G = Suguio et al., 1980), in northern Argentina (J = Codignotto et al., 1992), in Patagonia study area (Q, R, V and X = this paper) and in Tierra del Fuego (Gordillo et al., 1992, 1993). Pleistocene sea-level changes in Brazil (B = Martin et al., 1982 and Barbosa et al., 1986), in Patagonia study area (R = this work). (b) Emerged Pleistocene marine terraces along the eastern South American continent. Pleistocene sea-level changes in Brazil (B = Martin et al., 1982 and Barbosa et al., 1986), in Patagonia study area (R = this work). For letters and their corresponding sites see Fig. 1b.

increase in terrace elevations along the coast for terraces of the same age implies a south-north tilting of the entire east coast of the South American continent. While Holocene terraces of age 6000–7500 BP are found at 0–1 m amsl in Brazil (Barbosa et al., 1986; Martin et al., 1987; Bezerra et al., 1999) the same features in northern Argentina are at 5 m amsl (Codignotto and Aguirre, 1993, see above). In Patagonia they are at 6–7 m amsl, and at 8–9 m amsl in the southernmost region of the continent (e.g. Gordillo et al., 1993). Schnack et al. (1987) independently argue that Holocene littoral zones of the same age are higher in the south of Atlantic Patagonia than in the north. This also agrees with the observation made by Radtke (1989) that Holocene terraces are higher in southern Patagonia than their counterparts to the north. In Fig. 11b we illustrate the south to north variation of the OIS 5e terraces. Although the data is sparce the tilt on this horizon appears to be close to that which characterizes the Holocene features, a fact which would appear to suggest that the entire coast was tilted tectonically by some late Holocene event. This suggestion clearly deserves a much more detailed analysis than we are able to provide herein.

It is likely that the 7000 yr BP highstand terraces in Brazil, and probably the 7000–8000 yr BP highstand terraces in the north of Argentina as well, can be explained by the redistribution of water in the ocean basins, and as a consequence of hydro-isostatic adjustment, following ice-sheet melting. These processes clearly dominate the adjustment process in the "far-field" of ice-sheets (Clark et al., 1978; Peltier et al., 1978; Peltier, 1982, 1989; Mitrovica and Peltier, 1991; Peltier, 1998a). However, these processes alone may be insufficient to account for the observed elevations of the Patagonian terraces relative to those to the north. The fact that raised beaches found in the north are relatively low compared with those in the Patagonian study area (Fig. 11) implies that relative isostatic movement has taken place between these regions. If we accept the estimate of Flint (1971) that Tierra del Fuego was covered with approximately 1200 m ice, which was centered on the high Cordillera and which did not extend north of approximately 41°S (according to Flint and Fidalgo, 1969), then we would expect an additional rise in relative sea-level due to the influence of proglacial forebulge collapse in our study area. As the observations demonstrate (Fig. 12b and c), Holocene marine terraces in the study area are not submerged, but rather are uplifted to elevations of approximately 7 m amsl. Because sea-level rise driven by Patagonian proglacial forebulge collapse is not significant in the study area, this suggests that the original estimate of Patagonian ice-sheet thickness by Flint (1971) is exaggerated. Indeed, according to the recent Patagonian ice-sheet reconstruction of Hulton et al. (1994), which agrees with the ICE-4G model of Peltier (1994) in suggesting a maximum thickness of approximately 400 m, direct calculation demonstrates that removal of this ice-sheet has no significant effect upon the r.s.l. prediction for sites along the coast of Argentina. Since the local effect of the Fuego-Patagonian ice-sheet is insignificant, we therefore suggest that, in addition to hydro-isostasy, other effects may also have contributed to determining the elevations of these beach horizons.

Codignotto et al. (1992) have suggested that the Holocene uplift history along the Argentinian coast was connected to two different geomorphologic "provinces", one characterized by an uplift rate of 0.12 mm/yr which applies to sedimentary basin regions and other of 1.63 mm/yr associated with interbasin terrain. Schnack et al. (1987) concluded, by comparing the Holocene littoral zones of Patagonia with those to the north in the region of the Pampa (see Fig. 1a and b), that the Patagonian coast must has risen (or sea-level have fallen) at a rate of, at least, 1.0 mm/yr since 6000 BP. The results of Schnack et al. (1987) are reasonable for the Patagonian beach risers if we consider that Holocene littoral zones along the coast of Isla Grande, Tierra del Fuego, yielded an uplift rate of 2.90 mm/yr in the last 1000 yr (Gordillo et al., 1993). However, by employing our calculated uplift rate of 0.09 mm/yr for the Holocene beach risers we obtain the sea-level curve for Patagonia illustrated in Fig. 10b. The single uplift-corrected curve determined for Patagonia is characterized by a highstand of + 6 m amsl for ca. 8000 yr BP. This highstand is influenced by many factors and cannot be taken to represent the variation of eustatic sea-level. A significant component of this feature is controlled by the influence of hydro-isostatic adjustment following deglaciation, an influence that we investigate in detail in the following.

6. ICE-4G (VM2) model predictions of Holocene relative sea-level history in Patagonia

The recently constructed ICE-4G model of glacial isostatic adjustment and postglacial relative sea-level change (Peltier, 1994, 1996, 1998b) reconciles a very large quantity of Holocene data, both from glaciated and non-glaciated regions. In this model the timing of deglaciation is constrained by dated terminal moraines from the glaciated regions and by ice thickness as inferred from both the amplitudes of postglacial rebound, and the net rise of sea-level since the last glacial maximum [primarily constrained by the Barbados coral-based relative sea-level history (Fairbanks, 1989; Bard et al., 1990)]. In our nomenclature "VM2" (Viscosity Model 2) is a model inferred from the formal inversion of subset of glacial isostatic adjustment information (Peltier, 1996; Peltier and Jiang, 1996, 1997; Peltier, 1998c).

Fig. 12 displays a sample of the observed RSL histories and theoretical predictions based upon the ICE-4G (VM2) model of Glacial Isostatic Adjustment (GIA) along the eastern coast of South America. To investigate the predictive power of the ICE-4G (VM2) model for "far-field" sites in a larger geographic region than that of the study area of Argentinian Patagonia itself, we have collected published ^{14}C data from Venezuela and Brazil as well as from eight coastal sites from northern Argentina. Since the northern sites were reported in radiocarbon ages and because ^{14}C ages are not interchangeable with sidereal age, we converted them into sidereal time using the Calib 3.0 software of Stuiver and Reimer (1993). For the theoretical predictions the ICE-4G deglaciation model was employed together with two viscosity models, namely VM1 and VM2. Viscosity model VM1 (for which predictions are shown as dotted lines in Fig. 12) is characterized by a lithospheric thickness of 120.7 km, an upper mantle and transition zone viscosity of $v_{UM} = 1 \times 10^{21}$ Pa s and a lower mantle viscosity of $v_{LM} = 2 \times 10^{21}$ Pa s. Results for VM2 have also been computed assuming lithospheric thicknesses of both 120.7 km (solid lines on Fig. 12) and 80 km (dashed lines on Fig. 12). The VM2 model of mantle viscosity, with lithospheric thickness of 120.6 km, which was inferred from a formal inversion procedure that employed model VM1 as a first approximation (Peltier, 1996) reconciles a previously

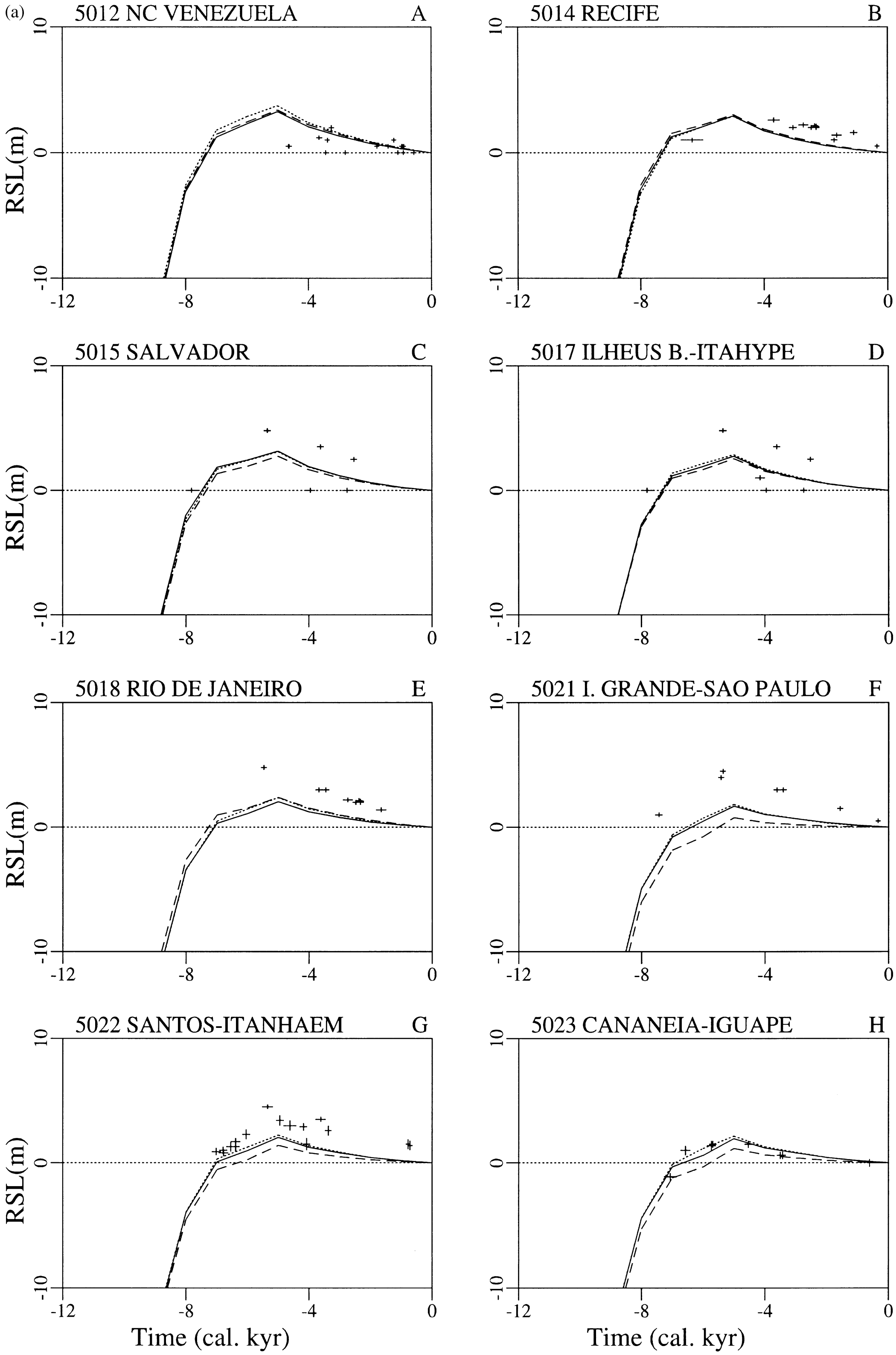

Fig. 12. (a) Comparison between theoretically predicted and observed (error bars) of relative sea- level history for eight "far-field" sites along the northern part of the South American Continent for Venezuela and for Brazil. (For Venezuela Valastro et al. (1980) and in Brazil for Recife Fairbridge (1967), Delibrias and Laborel (1971); for Salvador and Ilheus and Itahype Martin et al. (1987) and Delibrias and Laborel (1971); for Rio de Janeiro Fairbridge (1967) and Martin et al. (1987); for Isla Grande-Sao Paulo, Delibrias and Laborel (1971) and Suguio et al. (1980); for Santos-Itanhaem Suguio et al. (1980) and Martin et al. (1987) and for Cananheia-Iguape Suguio et al. (1980). The solid lines are predictions from the ICE-4G (VM2) model with lithospheric thickness of 120.7 km whereas the dashed lines denote VM2 with lithospheric thickness of 80 km. Dotted lines are predictions based upon the use of viscosity model VM1 with lithospheric thickness of 120.7 km (this model has an upper mantle viscosity of $v_{UM} = 1 \times 10^{21}$ Pa s and a lower mantle viscosity of $v_{LM} = 2 \times 10^{21}$ Pa s).(b) Same as in Fig. 12a, but for sites in the Atlantic Patagonia, south of 42°S latitude (this work and Codignotto et al., 1992). (c) Same as in Fig. 12a, but for the northern coast of Argentina (33° S–42° S latitude, Codignotto et al., 1992).

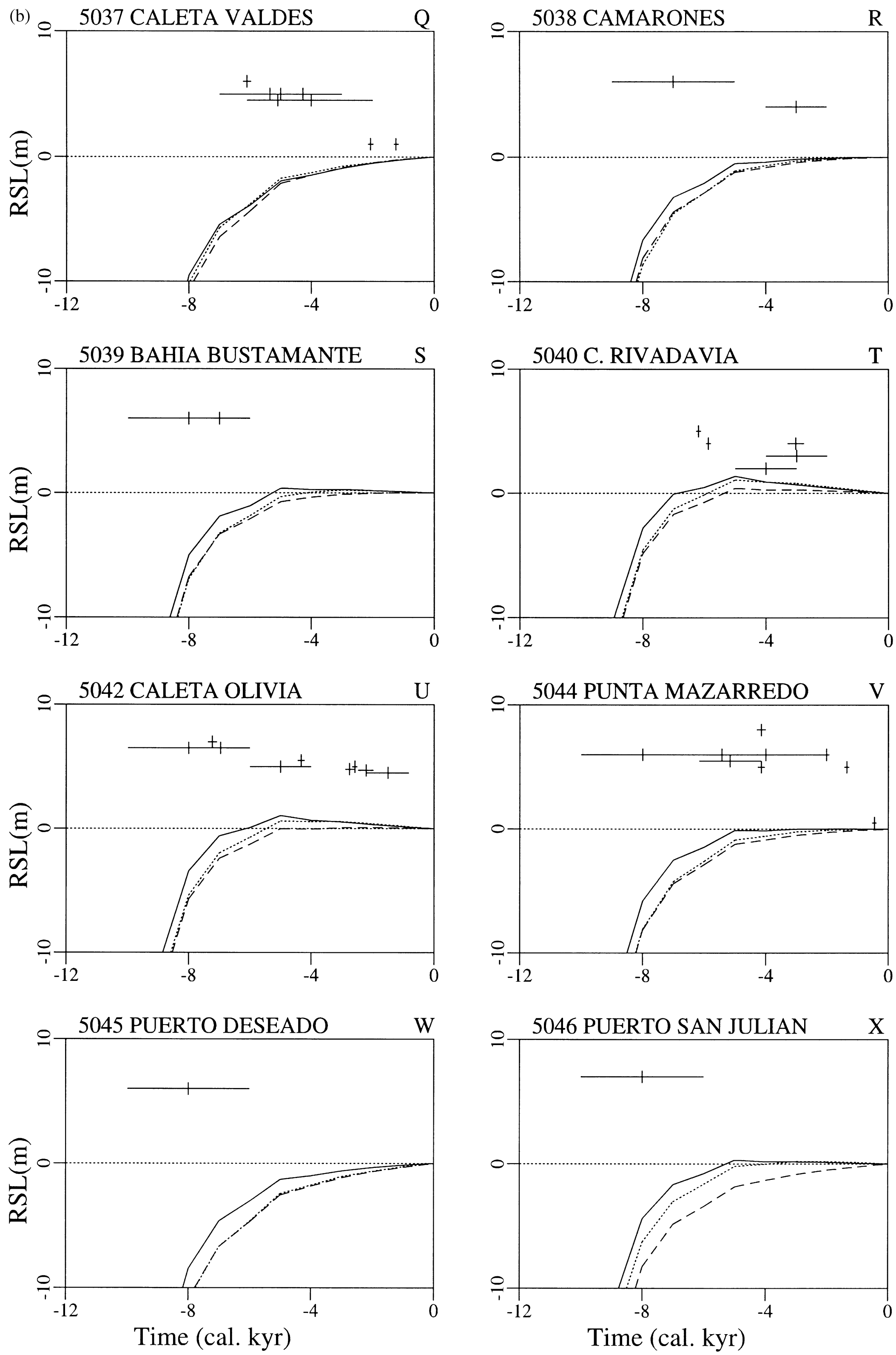

Fig. 12. (continued)

identified misfit of the VM1 based model to the data from the US east coast, a region which straddles the region of proglacial forebulge collapse that lies beyond the Laurentide ice-sheet. The individual sites for which model-data comparisons shown on Fig. 12 are also denoted by a number which identifies them in the global

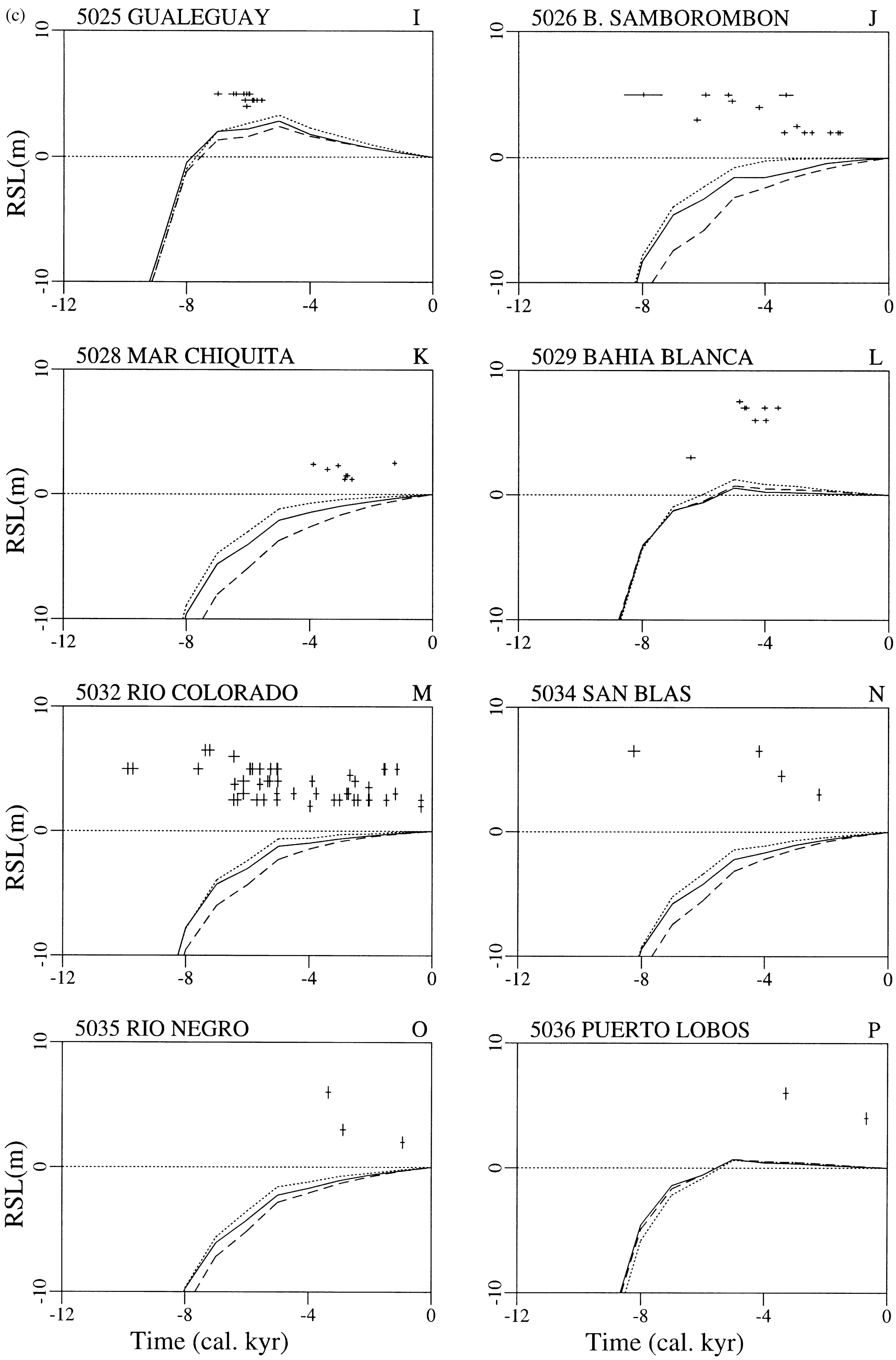

Fig. 12. (continued)

database of relative sea-level history that is being assembled at the University of Toronto to replace the crude reconnaissance collection of Tushingham and Peltier (1992). Each of the plates of Fig. 12 is also labelled with a letter from *A* through *X*, the locations of which are marked on the tectonic map of Fig. 1b. In Fig. 12a we

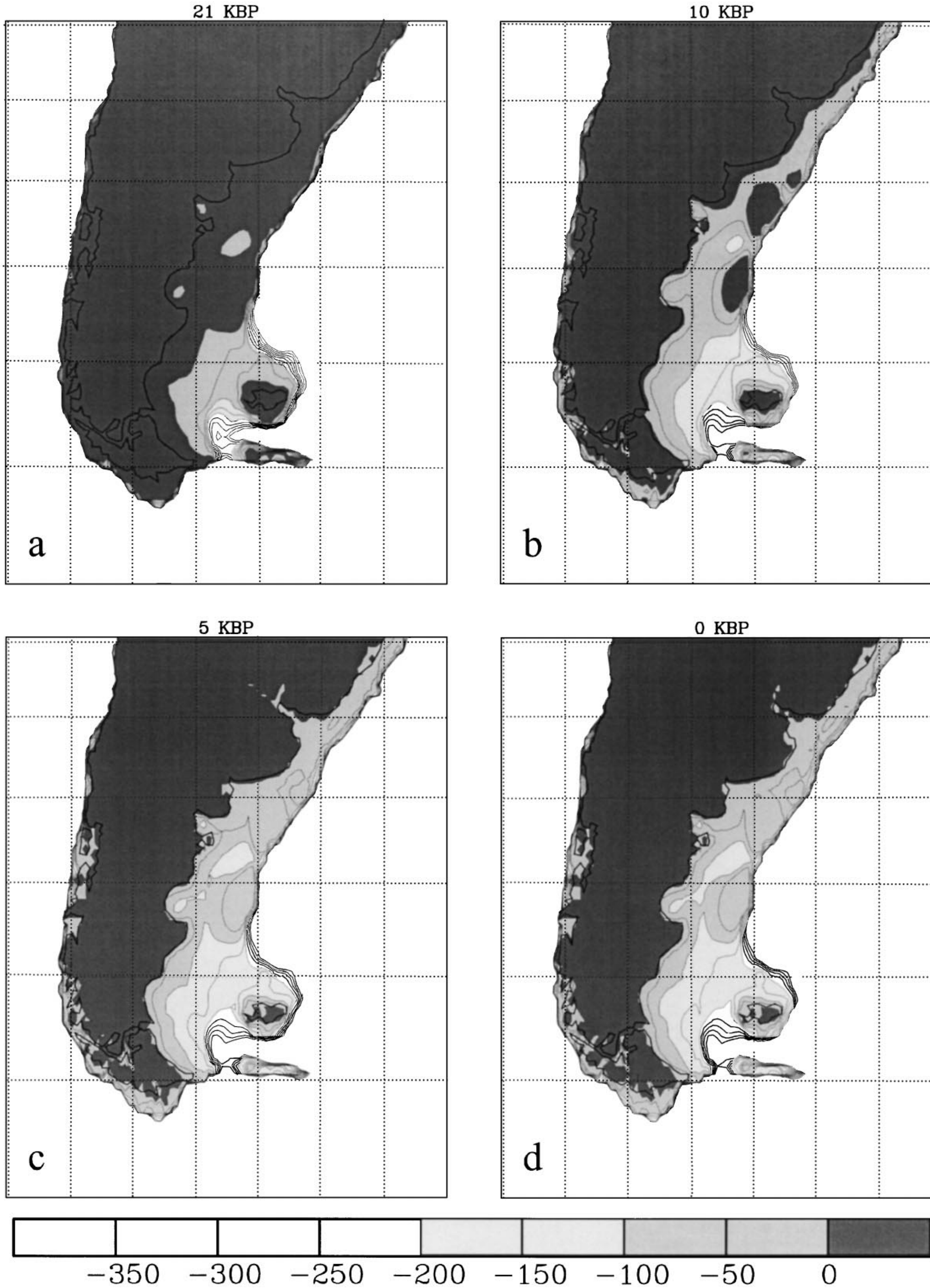

Fig. 13. The development of the paleotopography for the South American Continental Shelf as a function of time (see text).

display both observed and predicted Holocene r.s.l. histories for eight sites in northern South America [for Venezuela, Valastro et al. (1980); for Recife, Fairbridge (1976), Delibrias and Laborel (1971); for Salvador and Ilheus and Itahype, Martin et al. (1987) and Delibrias and Laborel (1971); for Rio de Janeiro, Fairbridge (1976) and

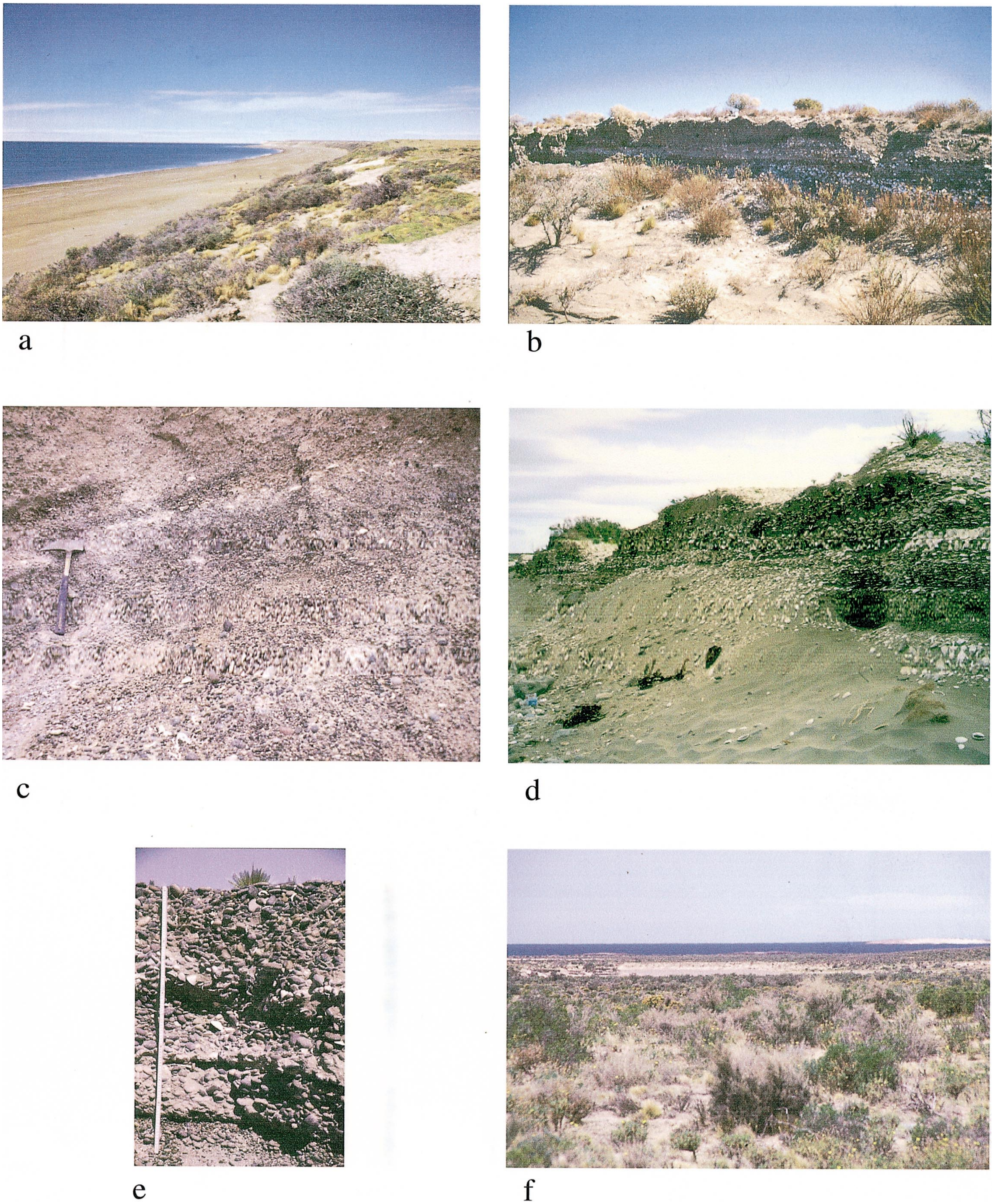

Fig. 14. (a) A view of a Pleistocene wave cut platform at 16–17 m elevation at the Camarones location. The hole excavated on the top of the terrace had to be deepened to 130 cm. (b) view of the Comodoro Rivadávia emerged Holocene beach deposits at 3 m elevation and at 5 m amsl (c) Lateral view at the higher Holocene shoreline (5 m) at the Comodoro Rivadávia. *Mercenaria* bearing horizon, consisting of silty sand and gravel, is 25 cm thick. (d) view at a Holocene "storm-beach" deposits at the Comodoro Rivadávia location. This non-representative feature is 6 m amsl. (e) Lateral view at Holocene beach-line at 7 m at Caleta Olivia (for the morphology of site see Fig. 5c). Molluscan fauna horizons were over 100 cm thick, which were subsequently washed out. Whole sediment is, as result of the dearth of matrix materials, loose and implies a non-representative location for sampling. (f) Seaward view of Pleistocene terrace at 16–17 m amsl at Caleta Olivia. The top of this prominent paleo shoreline was exposed in a hole of 20 m × 30 m × 2–3 m (i.e. breadth × length × depth).

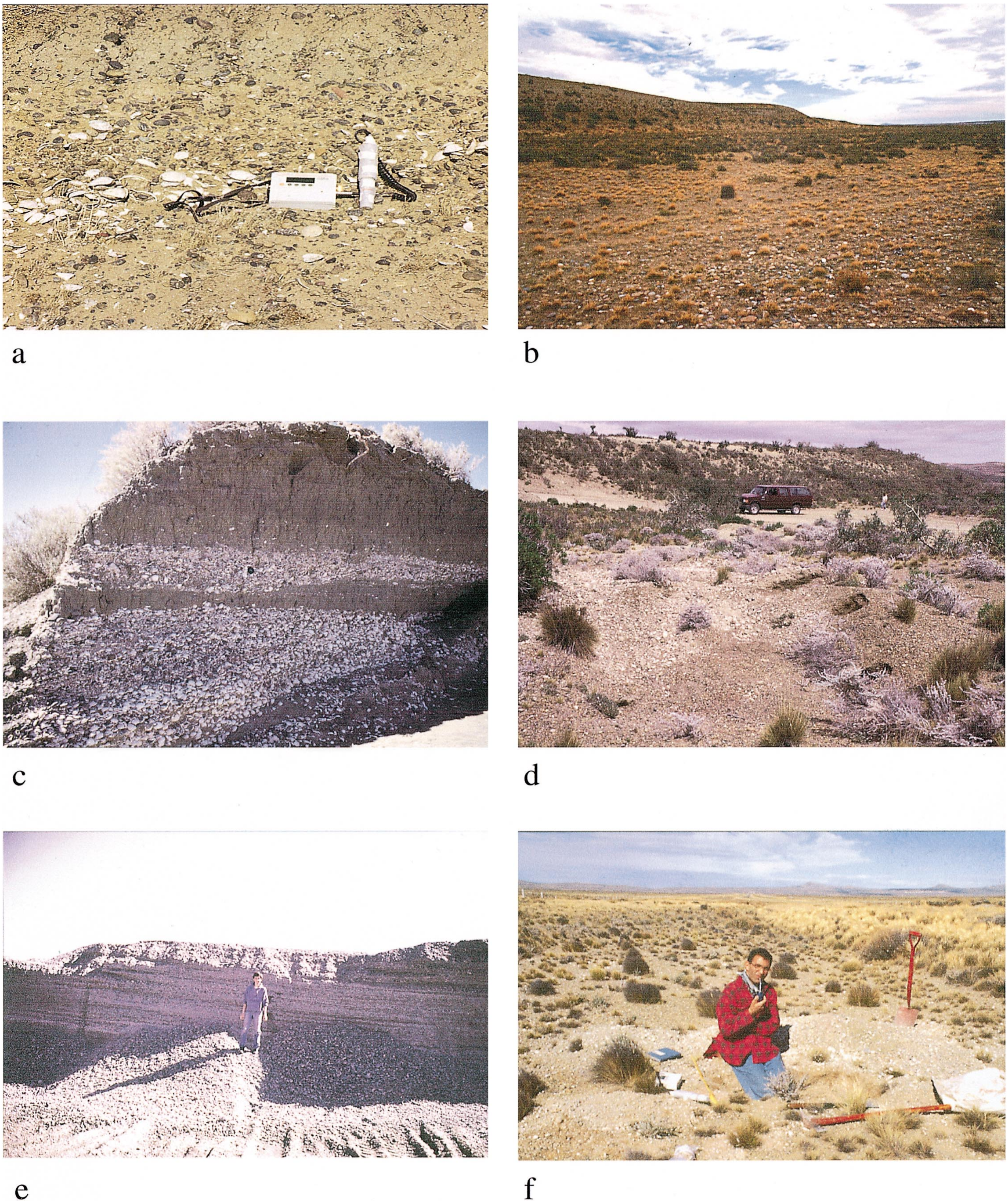

Fig. 15. (a) Lateral view at the Pleistocene beach-line presented in Fig. 14f at the Caleta Olivia. The laminated sandy basis contains, in addition to shells in living position, silty sand and gravel of dark color. This site was sampled after 50–60 cm sediment was removed from surface. (b) A view of the Quaternary wave cut platforms at Punta Mazarredo location. These representative terraces were at 16–17 m and at 6 m amsl. (c) View at a Holocene "storm-beach" deposits at the San Julian location. This non-representative feature is 7–8 m amsl. The black object in the picture is the lens cover of the camera. All shells found here were polished by energetic waves. (d) Quaternary wave cut platform at 16–17 amsl and at 7–8 m amsl at the San Julian location. To get fossil molluscs a pit was dug to 130 cm depth on the top the Pleistocene unit. (e) View of Holocene beach deposits at the San Julian site (f) Pleistocene paleo beach-deposit at approximately 32–34 m amsl at the San Julian location. To get representative fossil molluscs a pit was dug to a depth of approximately 130 cm on the top the Pleistocene unit.

Martin et al. (1987); for Isla Grande-Sao Paulo, Delibrias and Laborel (1971) and Suguio et al. (1980); for Santos-Itanhaem, Suguio et al. (1980) and Martin et al. (1987) and for Cananheia-Iguape, Suguio et al. (1980)]. Taking the scatter of the data into account, the predictions of the ICE-4G (VM2) model accord reasonably well with the observations. In Fig. 12b we present similar intercomparisons for the eight locations in our own study area. To maximize the number of data points being interpreted, we have supplemented our own data for Valdés, C. Olivia and P. Mazarredo with the additional data presented in Codignotto et al. (1992).

By inspection of Fig. 12b, it is clear that there are significant misfits between theory and observation, except at the Comodoro Rivadávia site where the misfits are relatively small. Since Patagonia and Brazil/Venezuela are "far-field" coastal locations we might reasonably assume that both regions must have undergone similar glacial isostatic adjustment (see Peltier (1998b) for example), whereas Figs. 12a and b indicate extreme dissimilarity. Because these two regions are distant from one another it is clearly important to know where along the coast the transition occurs between the region in which r.s.l. data are reasonably well fit by the ICE-4G (VM2) model and the region of misfit. We have therefore also collected data from the intervening zone to enable further comparisons between theory and observation (Fig. 12c). This zone, for which we employ the observations of Codignotto et al. (1992), extends from Buenos Aires in the north to our northernmost site of Caleta Valdés. Inspection demonstrates that large misfits between theory and observation are characteristic of this entire intervening segment. The boundary that gradually separates the regions of "fit" and "misfit" is located near the Rio de la Plata (approximately site J on Figs. 12c and 1b). North of this hinge the data are reasonably well predicted by the theory but south of the hinge the disagreement is significant.

In attempting to explain this marked difference of behavior between the northernmost and southernmost regions of the South American continent, it is important to recognize several competing geodynamic interactions. The influence of local glacial isostatic adjustment, as discussed above, has no significant effect on the Patagonian sea-level observations. Here, we consider two likely mechanisms: hydro-isostatic adjustment and plate tectonic deformation.

The process of hydro-isostatic adjustment is the process through which the water load added to the ocean basins during deglaciation induces a flow of material from beneath the ocean basins to beneath the continents, thus causing a "tilting" of the crust associated with an off-shore rise of sea-level coupled with an onshore fall (i.e. a fall of geoid height with respect to the surface of the solid earth). In analyzing the influence of hydro-isostasy it is important to recognize the extraordinary character of the continental shelf off the coast of South America. As shown by the position of the shelf break on Fig. 1b (dashed line), along the northern part of the east coast of South America the shelf is relatively narrow whereas to the south (beginning approximately at site J) it is broad. In fact the region offshore of Argentinian Patagonia is one of the most extensive shelf regions of the world (500–550 km wide at San Blas, 550–560 km wide offshore of Comodoro Rivadávia and approximately 800 km wide offshore of Santa Cruz). In contrast to the north at Ilheus, in Brazil, the shelf obtains a maximum width of approximately 20 km and attains a maximum width at Cananéia and Santos, of approximately 170 km. Here we suggest that the extraordinary width of the shelf offshore of Argentinian Patagonia may have had an influence on the postglacial relative sea-level history in this region which has not been adequately captured in the theoretical model of the GIA process that we are employing.

A somewhat more likely explanation of the observed misfit of the data to the GIA model, however, concerns the tectonic environment in southern Argentina. As one moves southward along the east coast of South America, one approaches ever the more closely the Chile-trench, a subduction zone in which convergence rates decrease from almost 11 cm/yr in mid-continent to 3.3 cm/yr in Patagonia (DeMets et al., 1990 Fig. 1b). In this setting, the influence of subduction-related tectonic deformation might become more apparent with increasing proximity to the Chile-trench.

Fig. 15 illustrates the evolving land-sea distribution of the Patagonia study area for the times 21,000, 10,000, 5000 kyr BP and present. To reconstruct the regional paleotopography, we made use of the ETOPO 5 topographic dataset of the National Center for Atmospheric Research together with the topographically self-consistent form of the theoretical model of the global process of glacial isostatic adjustment (Peltier, 1994). Fig. 13b shows that the Patagonian shelf was first inundated by the sea during deglaciation approximately 10,000 yr BP. From these analyses the time variation of the position of the coast in the Patagonia study area makes it unsurprising that a careful account needs be taken of this influence for the accurate prediction of r.s.l. history. It is outside the scope of this paper to analyze this issue in further detail, but it may well be that large variations of tidal range occurred along the coast in response to the large changes in offshore bathymetry and that this could also be the explanation of some part of the intriguing misfits that our analyses have revealed.

7. Summary and conclusions

Because fossil mollusc shells are the only datable material contained within the Quaternary marine terrace sequences along the Patagonian coast of Argentina, we have elected to analyze them using Th/U, ESR and radiocarbon dating, in spite of the reservations previously expressed concerning the use of U-series ages derived from fossil molluscs shells (Kaufman et al., 1971). The results were consistent for common sites with those dated using the AAR and ESR methods (Rutter et al., 1989, 1990), and also correlate well with ages obtained using the ESR technique (Radtke, 1989; Schellmann and Radtke, 1997; Schellmann, 1998a, b). The estimated ages fall into four relatively well-defined groups. The first of these, with ages $>250{,}000$ yr BP, is associated with the terraces at 33–34 m elevation. The second category includes samples ranging from 160,000 to 240,000 yr BP and were associated with terraces at 14–15 m elevation. This group of shells is interpreted as being related to a sea-level highstand older than OIS 5e. A third category, ranging in age from 104,000 to 143,000 yr BP, are interpreted as being associated with the culmination of the Last Interglacial (OIS 5e). These samples were found in paleoshorelines which are now at 16–17 m amsl and were found in many locations. In terms of stratigraphy, sedimentology and morphology, this feature was the best-defined terrace structure identifiable within the study area.

The fourth group of fossil mussel shells which range in age from 3000 to 8000 yr BP represent Holocene sea-levels higher than present. In this category, the 7000 to 8000 yr old molluscs are related to the peak postglacial Holocene transgression, whereas younger features relate to subsequent regressions of sea-level.

From these results we have reconstructed the sea-level changes over the past 300 kyr along the relatively tectonically stable Atlantic coast of Patagonia. Assuming that individual terraces developed as rising sea-levels overtook the land, each mollusc-bearing terrace unit represents the culmination of each successive transgression. From the ages of these units and their elevations, we inferred a long-term uplift rate of 0.09 m per 1000 yr. On the basis of this low rate of Pleistocene uplift, we demonstrate that OIS 7 and 5e are expected to coincide, and we estimate that relative sea-level was at an elevation of $+6 \pm 2$ m for OIS 5e and of -4 ± 2 for OIS 7. Despite concerns regarding the ages for OIS 9, which should be considered as lower limits, we tentatively infer a sea-level elevation of approximately $+6 \pm 2$ m for this highstand. An important conclusion from these analyses (in good agreement with previous inferences based upon corals) is that fossil mollusc shells, if found "in situ", in reliable terrace horizons and carefully analyzed, may in fact yield useful age estimates.

On the basis of our work we were unable to confirm the interpretation of Codignotto et al. (1992) concerning the existence of two distinct uplift rates within the study area, 0.12 mm/yr which applies to sedimentary basin regions and 1.63 mm/yr associated with interbasin terrain. Our results for the shorelines revealed a higher rate of relative sea-level fall during the Holocene than the average Pleistocene rate, and that the magnitude of this effective uplift rate decreases northward along the coast. Our results demonstrate that relative sea-level predictions based upon the ICE-4G (VM2) deglaciation model, which acceptably reconcile the observations from the northern part of the east coast of the South American Continent, significantly misfit the observations for the southern part of South America, including the Patagonia study area. The exceptional width of the Argentinian shelf linked to hydro-isostasy/tidal range is a possible contributing cause of the misfits to the observations in this region. In circumstances in which the continental shelf is exceptionally broad, the effective location of the coast (insofar as the influence of hydro-isostasy is concerned) may be, perhaps significantly, outside its actual modern location. It seems equally plausible, however, that a Holocene uplift of the region has indeed occurred in association with subduction-related tectonic deformation close to the Chilean trench, and/or that the Holocene horizons throughout Argentinian Patagonia are anomalously raised by the high energy of this coastal environment. Future work will determine the relative importance of these various influences.

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Appendix A

A.1. Geochronological issues and the ages of older units

In Patagonia, where fossil coral is scarce, mollusc shells provide the only datable material on the basis of which one may establish a chronology of paleoenvironmental events. However, Kaufman et al. (1971) reviewed about 400 analyses and reported discouraging results, to the effect that the $^{230}Th/^{234}U$ ages of molluscs were found to be unreliable and furthermore that the shell ages derived from using U-series with ages greater than 250,000 yr seemed even less promising (Kaufman et al., 1996). Most of the molluscs investigated in our paper confirm the observations of Kaufman et al. (1971) concerning post-mortem uranium uptake but do not necessarily confirm their conclusions that ages are unreliable. The range of uranium concentrations for the Holocene samples ($\leqslant$ 8000 years) is 0.03 ± 0.01 to 0.40 ± 0.03 ppm. The uranium activity ratio for this group of samples ranges from 1.07 ± 0.08 to 1.40 ± 0.04 (Table 1). For samples from the Last Interglacial, the uranium content in mussel shells ranges from 0.55 ± 0.04 to 1.88 ± 0.06 ppm with four additional shells ranging in uranium from 2.03 ± 0.04 to 2.92 ± 0.24 ppm that were collected from different locations.

The uranium activity ratio for the group of the Last Interglacial shells ranges in general from 1.18 ± 0.06 to 1.40 ± 0.05 with two additional samples of 1.56 ± 0.03 and 1.60 ± 0.13 activity ratios. For specimens belonging to Oxygen Isotope stage 9, uranium concentration is considerably higher and ranges from 1.01 ± 0.04 to 4.48 ± 0.06 ppm and in one case 5.46 ± 0.06 ppm. The activities of uranium in these shells were found to be 1.22 ± 0.06 to 1.50 ± 0.04. A significant number of the uranium activity ratios measured in the shells are higher than oceanic (1.150 ± 0.002, cf. Chen et al., 1986).

In calculating the U-series ages of these critical mussel shells, they are assumed to have remained ideal, closed systems throughout their deposition, that is, that there has been no post-depositional migration of U-isotopes nor is there in situ production of long-lived daughter product of ^{230}Th. As Table 1 demonstrates, we observe deviations in uranium concentrations as well as in U-activity ratios ($^{234}U/^{238}U$) that indicate an open system but we also have commensurably and comparably higher ^{230}Th in these same samples. As thorium itself is well known to be a particularly reactive element (that is because it adsorbs easily on the matrix material through which the ground water passes, and because of the absence of significant amounts of ^{232}Th in our samples, which implies that the shells are free of inherited ^{230}Th from detrital materials) we believe that the ^{230}Th measurements are reliable. Further support for the general validity of ^{230}Th measurements are $^{230}Th/^{232}Th$ activity ratios that are $\geqslant 25$ (Rutter and Catto, 1995) indicating that contamination of the shells by extraneous ^{230}Th cannot be responsible for the ^{230}Th content. However, we cannot rule out the possibility of contamination by proximate shells or sediments that may be responsible for elevating the apparent age above the true ages. It is nevertheless true that the ^{230}Th observed in the fossil molluscs could have resulted solely from radioactive decay of the parent uranium.

Our working assumption for the purpose of interpreting these observations is therefore that the deduced Th/U ages may be close to the real age of the specimens. In spite of the fact that the pathway through which and the time during which uranium diffuses into the shells is not well known, changes in environmental habitat of molluscs from marine to vadose/phreatic might lead, because of post-mortem chemical alteration in and within the shells, to accumulation of uranium after a few thousand years (Kaufman et al., 1996). Since we have no knowledge of the exact timing of U-uptake and of the activity ratio of the incorporated U-isotopes, which are generally much higher than the activity ratio of sea water, we assume that the molluscs have absorbed uranium post-mortem for perhaps 10 kyr after which the system again became closed. In this case the ^{230}Th measured in the samples would have resulted from radioactive decay of parent uranium and Th/U ratios will deliver apparent ages younger than the “true” age of the fossils. Particularly for our pre-OIS 5e terraces, this assumption is open to criticism. However, the chronology of these terraces has also been, frequently, confirmed by application of ESR and AAR dating methods (Radtke, 1989; Rutter et al., 1989, 1990; Schellmann and Radtke, 1997; Schellmann, 1998a, b). Furthermore, the evidence is convincing that stratigraphically higher levels and shorelines in our region of relative tectonic stability, may be identified as older sea-level peaks. Since the terrace found at 33–34 m amsl is morphologically the highest feature which still yields Th/U datable mollusc shells, we infer that these paleo beach-deposits were formed during Oxygen Isotope stage 9. If the terrace belonged to a transgression older than stage 9 (i.e. stage 11), then we would expect that these species would have already decomposed, because of the impact of weathering processes given the maximum thickness of approximately 2 mm of the fossil shells.

A.2. ESR-method

ESR dating is based on the ability of the carbonate lattice to store radioactive energy (Zeller et al., 1967). Constant exposure of the material to ionizing radiation resulting from the decay of uranium, thorium and potassium in the surrounding sediment produces free electrons in the carbonate, of which a certain number are trapped at impurity sites in the lattice (Barabas et al., 1992a, b). As the number of trapped electrons increases with time and as, during the mineralization of the carbonate, no radicals were present (i.e., zero ESR signal), the intensity of the ESR signal measured today is correlated to the age of the sample, as:

$$D_{\mathrm{E}} = \int_{0}^{\mathrm{Age}} \dot{D}(t)\mathrm{d}t,$$

with D_E the total dose received by the sample (in Gy) and $\dot{D}$ the dose rate of the surrounding sediment (in Gy/yr).

The equivalent dose D_E is determined by the additive-dose-method (Grün, 1989): the sample is successively artificially irradiated in the laboratory with a ^{60}Co–source and the ESR intensity is plotted against the applied dose. Extrapolation to zero ESR signal using an exponential saturation function yields the D_E as the negative intercept with the dose-axis (Barabas et al., 1988; Grün, 1990; Barabas et al., 1992a, b):

$$I(D) = I_{\max}\left[1 - \exp\left(\frac{-(D_E + D)^{\gamma}}{D_0}\right)\right]$$

where I is the ESR signal, $I_{\max}$, the saturation signal, D, the irradiation dose, D_0, the saturation dose and γ has a value of 0.65 (see Woda, 1996)

The dose rate ($\dot{D}$) is the sum of external and internal components. The external dose rate was determined from the concentrations of ^{232}Th,^{238}U and ^{40}K in the embedding sediments, assuming radioactive equilibrium and a negligible water content (Aitken, 1985). Because the outer part of the valves was removed (approx. 0.2 mm) the external contribution of alpha irradiation was neglected, whereas the external and internal contributions of the beta-irradiation were estimated at 20% and 80%, respectively. The internal dose rate was determined assuming incorporation of uranium shortly after sedimentation of the molluscs and an alpha efficiency of 0.05. For example, the internal dose rate of a 100 kyr old sample with 1 ppm uranium amounts to about 10% of the total dose rate.

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