Non-steady long-term uplift rates and Pleistocene marine terrace development along the Andean margin of Chile (31°S) inferred from 10Be dating

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ARTICLE INFO

Article history:
Received 21 January 2008
Received in revised form 18 September 2008
Accepted 26 September 2008
Available online 22 November 2008
Editor: C.P. Jaupart

Keywords:
Beryllium-10 
marine terrace 
 Pleistocene 
 uplift rate 
 Central Andes 
 underplating

ABSTRACT

Pleistocene uplift of the Chilean coast is recorded by the formation of wave-cut platforms resulting from marine erosion during sea-level highstands. In the Altos de Talinay area (~31°S), we have identified a sequence of 5 wave-cut platforms. Using in situ produced 10Be exposure ages we show that these platforms were formed during interglacial periods at 6, 122, 232, 321 and 690 ka. These ages correspond to marine isotopic stages (MIS) or substages (MISS) 1, 5e, 7e, 9c and 17. Shoreline angle elevations used in conjunction with our chronology of wave-cut platform formation, illustrate that surface uplift rates vary from 103±69 mm/ka between 122 and 6 ka, to 1158±416 mm/ka between 321 and 232 ka. The absence of preserved platforms related to the MIS 11, 13 and 15 highstands likely reflects slow uplift rates during these times. We suggest that since 700 ka, the Altos de Talinay area was predominantly uplifted during 2 short periods following MIS 17 and MISS 9c. This episodic uplift of the Chilean coast in the Pleistocene may result from subduction related processes, such as pulses of tectonic accretion at the base of the forearc wedge.

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1. Introduction

A former assumption used in the analysis of tectonic deformation over geodetic (10 yr) and geologic (10⁶ yr) time scales is that the rate of deformation should remain constant over these time scales. In order to resolve this question and provide new constraints for integrating studies of geodetic, seismologic and geochronologic data, our study focuses on uplift rate calculations over various time ranges. This should lead to a unique estimation of long-term rates of deformation (Chéry and Vernant, 2006) while, at the same time, provide an estimation of variability over shorter time scales. This premise is currently being tested (cf. Friedrich et al., 2003; Allmendinger et al., 2005). However when calculating slip- or uplift-rates in order to define active tectonics, it was necessary in most of previous studies to assume that the short-term rate inferred from the displacement of a geologic marker (<100 ka), may be a robust signal of the average long-term rate over a time period of ~0.4–1 Ma (Bloom et al., 1974; Bull, 1985; Lajoie, 1986; Burbank and Anderson, 2001; Marquardt et al., 2004). In this paper, we study and date a sequence of five marine terraces to constrain the variability of the coastal Andean uplift rate. Considering that each marine terrace has a time resolution related to one sea-level highstand, our data integrate uplift rate over 100 kyr time steps, the periodicity of marine terrace formation. More rapid uplift rate variations that may have occurred during this time step will not be represented on the morphological scale and thus are not considered.

Marine terraces are an important class of geologic markers that are produced through complex interactions between sea-level fluctuations and uplift along an active margin (Bradley and Griggs, 1976; Merritts and Bull, 1988; Anderson et al., 1999; Mubs et al., 1990; Lajoie et al., 1991) that can in theory, provide important constraints on the rates of coastal uplift (Chappell, 1983; Lajoie, 1986; Burbank and Anderson, 2001). Along the Andean forearc, sequences of marine terraces record changes in sea-level together with the uplift history of the coastal area (DeVries, 1988; Hsu, 1992; Goy et al., 1992; Macháré and Ortlieb, 1992; Ortlieb et al., 1992, 1996; Ota et al., 1995; Zazo, 1999; Zazo et al., 1994; Cantalamessa and DiCelma, 2004; Marquardt et al., 2004; Pedoja, 2003; Pedoja et al., 2006; Quezada et al., 2007). However, at many locations (e.g., Papua New Guinea, USA, Peru, Chile,
Ecuador) where marine terraces have been studied, steady uplift is often assumed and chronostratigraphical correlation of the terraces to sea-level highstands is used to assign ages due to restricted access to datable material (Muhs et al., 1990; Hsu, 1992; Macharé and Ortlieb, 1992; Goy et al., 1992; Ota et al., 1995; Burbank and Anderson, 2001; Cantalamessa and Di Celma, 2004; Pedoja et al., 2006). In this study, we focus on the Altos de Talinay area of Chile (~30°–31°S). The general morphology of the area has been previously described (Paskoff, 1966, 1970; Chávez, 1967; Herm, 1969; Ota et al., 1995; Benado, 2000; Heinze, 2003). This area shows a well developed sequence of five wave-cut platforms (WCPs) directly abraded into the bedrock.

The study area is located above the Pacific shoreline, at the foot of the Chilean Coastal Cordillera along a ~100 km-long stretch of coast from Tongoy (30.25°S) to south of Bahía El Teniente (31°S; Coquimbo region, Chile; Fig. 1). Due to the subduction of the Nazca Plate beneath the South American plate at a rate of ~82 mm/a (DeMets et al., 1994), this segment of the Chilean coast is seismically active and the area experiences a recurrent major interplate event approximately every 100 yrs (1647, Mw=8.5; 1730, Mw=8.7; 1880, Mw=7.7; 1943, Mw=8.3; Beck et al., 1998; Heinze, 2003). Moreover, this region shows a high degree of seismic coupling between the subducting Nazca and the overriding South America plates (Lemoine et al., 2001; Pedoja et al., 2006).
Pardo et al., 2002; Gardi et al., 2006). This suggests that the upper plate morphology may display evidence of the ongoing subduction processes related to the position of the seismogenic zone (i.e. marine terraces or/and active tectonics; Audin et al., 2008).

In the Coquimbo region, the Coastal Cordillera progressively passes to the Precordillera and finally the high Cordillera of the Andes. In contrast to the morphology north of La Serena (~29.9°S) or south of 33°S, the morphology of the margin along this segment of the Andean Cordillera does not show a Central Depression nor an active volcanic arc. This particular morphological sequence is developed above a segment of flat subduction – between 30°S and 33°S – which is bordered to the north and south by normal dipping subduction zones (Gutscher et al., 2000; Yañez et al., 2001; Fig. 1). Yañez et al. (2001) suggest that the flat slab segment of the Nazca plate in this area is...
related to the aseismic Juan Fernandez ridge (33°S), which has been continuously subducting beneath South America at the same piercing point since 10 Ma. Thus, the Juan Fernandez ridge did not induce the Pleistocene WCP formation observed in the Tongoy region (Fig. 1).

A number of different methodologies have previously been employed to establish the chronology of terrace formation in this area. These include amino-acid racemization, Electron Spin Resonance, 14C dating and U-series dating of shells (Herm, 1969; Ota et al., 1995). However, due to the poor shell content of the deposits, accurate age determinations are still lacking. In contrast, in situ produced cosmogenic radionuclides (CRNs) have been demonstrated to be a robust tool for the absolute dating of marine terraces (Perg et al., 2001; Kim and Sutherland, 2004; Quezada et al., 2007; Alvarez-Marrón et al., 2008). Abrasion surfaces dating constitutes an original approach with respect to built marine terraces (Ortlieb, 1987). We thus avoid any source of errors due to inheritance or transport because of the availability of in situ exposed bedrock. In this study, we present new chronologic data from the sequence of marine terraces in this area to provide absolute coastal uplift rates on the Andean margin over the last ~1 Ma. Our approach is to quantify uplift rates along the Chilean coastal active margin using mapped geomorphic markers, cosmogenic 10Be surface exposure ages of the WCP sequence and known sea level changes.

2. Geologic setting

Within the study area, the Coastal Cordillera is a zone of distinct relief reaching elevations of 742 masl, locally called the Altos de Talinay. Here, wide and well-defined WCPs occur along the western edge of the Coastal Cordillera. The eastern edge of the Coastal Cordillera is delimited to the north of the study area by the Tongoy Bay in the north. The Río Limarí drainage basin is located in the southern part of the study area together with an intervening lowland referred to here as the Tongoy Paleobay Depression (Fig. 2). The contact between the Coastal Cordillera and Tongoy Paleobay Depression is a steeply east dipping normal fault, the Puerto Aldea Fault (PAF). This NNW–SSE fault shows a left-lateral component and offsets Plio–Pleistocene alluvial and fluvial strata of the Limari Formation by ~5 m. A number of micro-basins, filled with Plio–Pleistocene sediments or Quaternary alluvial deposits, are juxtaposed along the PAF (Heinze, 2003). The regional bedrock is mainly Devonian–Carboniferous metasediments and Triassic–Jurassic igneous complexes (Bohn Horst, 1967; Heinze, 2003; Fig. 2). As a result of erosive wave-action, the WCPs were developed indifferently across this heterogeneous substratum.

In contrast, the Tongoy Paleobay Depression has undergone a different coastal dynamic and thus, is filled with the Miocene–Pliocene marine and continental sedimentary series of the Coquimbo.
3. Morphotectonic analysis

3.1. Geomorphologic model of WCP formation

A WCP is formed by wave and wind erosional processes during a sea-level highstand. The extent of WCP development depends on how long the base level is static. When base level is static over a long period, increased coastal cliff recession results in a wider WCP (Trenhaile, 2000; de Lange and Moon, 2005). The complete morphology of a WCP contains a scarp (a previous coastal cliff), a planar wave-cut surface and a shoreline angle or inner edge (Fig. 3A). The shoreline angle is a geomorphologic feature that marks the change in slope between the scarp of the previous WCP and the planar wave-cut surface. It reflects the position of the former coastline and its elevation characterizes each WCP (Lajoie, 1986). We measured shoreline angle elevations using a GPS profile. In the field, outcrops of the basement bedrock exist in numerous places along the wave-cut platform, free of debris. But in most cases where the shoreline angle is covered by debris flows, we considered the shoreline angle elevation as the intersection between the regional slope of the planar wave-cut surface and the mean slope of the above scarp (cf. Fig. 4). According to our model, when base level decreases as a result of sea-level fall and/or coastal uplift, the WCP is exposed and preserved while a new coastal cliff is eroded. However, if a subsequent sea-level highstand is high enough and persists over a long time period, coastal cliff recession may be so great that it erodes a previously formed WCP, or a succession of WCPs (Anderson et al., 1999). Marine terraces can also be completely eroded during the next transgression if the uplift rate is low and/or the time interval between highstands is small (Rosenbloom and Anderson, 1994). As a result, a sequence of preserved WCPs will not likely record all of the sea level highstands in the marine record.

3.2. WCPs of the Altos de Talinay area

In the Altos de Talinay area, five staircased WCPs up to 400 masl are preserved in a less than 10 km stretch of coast. We refer to these terraces as Talinay I–V from the oldest and highest (Talinay I; T1) to the youngest and lowest (Talinay V; T5; Ota et al., 1995; Figs. 2–4). Each platform is an erosional wave-cut surface carved out igneous rocks, which are now covered with scattered siliciclastic sediments less than 20 cm thick. The WCPs dip gently seaward and are discontinuous along the coast. Where the discontinuous terrace segments are identified along the coast, there are slight variations in the elevation of their shoreline angles and in their platform widths. We mapped and sampled this sequence of platforms in two locations within the study area, one in the northern end and one in the southern end (Fig. 4).

T1 is the widest WCP within the study area and extends from Punta Lengua de Vaca to south of 30°6’S. T1 is continuous from south of Punta Limari Sur to south of Bahía El Teniente. T1 extends over a minimum of a 1.3 km to a maximum of a 7.5 km-wide area with the widest part in the southern portion of the study area. While the shoreline angle is not always preserved, we measured its elevation where it is best preserved. On the A–B GPS profile (Fig. 4), the T1 shoreline angle elevation is well constrained. We calculated the uncertainty on this measurement according to the C–D GPS profile. The T1 shoreline angle elevation is ~425 ± 15 masl.

T2 is the best-preserved WCP of the sequence. It is continuous over the ~100 km-long coastal area studied and it tapers out to the south near Bahía El Teniente (Figs. 2 and 4). T2 extends over a minimum of a 500 m to a maximum of a 2.5 km-wide area and has a very shallow slope (less than 1°). Contrary to T1, T2 is wider in the northern portion of the study area. The WCP shoreline angle elevation varies from 150 masl to 190 masl. A steep and dissected scarp, the remnant coastal cliff, separates T2 from T1. Some sand deposits overlie the scoured bedrock particularly close to the shoreline angle near the scarp of T2.

T3 is the least well developed WCP. Along the coastal area, this platform is very discontinuous and narrow with a maximum width of 400 m. T3 is more developed in the southern part of the study area and extends from Punta Limari Norte to north of Bahía El Teniente. Only a small remnant of T3 is preserved in the north, at the extreme northern end near Punta Lengua de Vaca. Its shoreline angle elevation is ~55 ± 5 masl. T3 morphology often shows stacks of the previous surface (T1) which shows that T3 is not completely abraded. Additionally, the eroded bedrock surface is partially covered by a few centimeters of sand deposits.

T4 is quite continuous along the ~100 km coastal area. It is narrow with a maximum width of ~1 km near Mina Talca. This WCP contains many remnants of the previous surface and is covered by a thin layer of sand with sub-angular clasts (~10–50 cm in diameter). The elevation of the shoreline angle varies between 22 and 28 masl.

T5 is discontinuous along the coastal area from Punta Lengua de Vaca to south of Bahía El Teniente. It is very narrow and does not exceed 150 m in width. In the topographically lowest portions of this WCP, marine action (waves and spindrift) still abrades this surface. The abraded bedrock is covered by several centimeters of beach deposits overlain by angular to sub-angular clasts (~10–40 cm in diameter). The elevation of the shoreline angle varies between 5 and 7 masl.

3.3. Fault systems affecting the WCPs

Along the western edge of the Altos de Talinay, numerous faults striking sub-parallel to the coast offset these planar surfaces. In general, the structures are NNW–SEE to N–S trending, subvertical, normal faults (e.g., Quebrada del Teniente Fault, Los Huiros Fault, Los Loros Fault, El Fraile Fault) and, generally, the block is uplifted (Fig. 4). Dip parallel striae suggest that the faults are predominantly dip slip, with no evidence of a strike-slip component. They affect mainly T1 (e.g., Quebrada Palo Cortado, Quebrada del Teniente Fault, Los Huiros Fault) and to a lesser degree T2 (e.g., Quebrada Palo Cortado Fault, Los Loros Fault, Los Huiros Fault, El Fraile Fault). There is no robust evidence of fault displacements on T1, T2 and T3 along the coast.

Active faulting influences the morphology of the WCPs by affecting shoreline angle elevations, surface gradients, and platform widths. For example, shoreline angle elevations of the five WCPs are lower in the south than in the north. T1 is not only cut by a greater number of faults than the other four surfaces but also with larger vertical offsets. Some faults cut obliquely through the set of WCPs and thus, we can determine the recurrent activity of those faults. For example, based on offset shoreline angle elevations, the Quebrada Palo Cortado Fault displaces T1.
by 100 m and TII by 20 m (Ota et al., 1995; Fig. 4). No vertical offset along this fault is observed on TIV and TV. As a second example, the Quebrada del Teniente Fault cuts TII and TIII with a varying amount of vertical offset along the same terrace, decreasing along strike from south – where TII is offset by 26 m at Bahía El Teniente – to the north where TIII is offset by 1 m. There is no vertical displacement on TIB, TIV and TV.

4. Methodology: ¹⁰Be exposure age models and sampling strategy

The use of cosmogenic radionuclides (CRN) to date alluvial and marine terraces is established as a robust tool in many previous studies (Perg et al., 2001; Brocard et al., 2003; Siame et al., 2004; Kim and Sutherland, 2004; Vassalo et al., 2007; Regard et al., 2006). In situ produced CRN – here we use ¹⁰Be – record the duration of surface exposure to cosmic rays (cf. Gosse and Phillips, 2001; Siame et al., 2006) and can thus be used to provide an absolute exposure age of a surface. We apply the following strategy to sample WCPs in the Altos de Talinay area and estimate the abandonment age of the marine terrace. In our model, the surface CRN concentration is zero during WCP formation as a result of the abrasion process and last steps of terrace leveling. CRN accumulation begins after terrace abandonment due to either marine regression or a period of coastal uplift (or a combination of both). We sampled each WCP away from the shoreline angle to limit burying problems (such as alluvial or colluvial deposits) and away from the outer scarp due to scarp diffusion. Considering our hypothesis the obtained age of one site per WCP should apply to the whole WCP function of the direct correlation to the sea level curve. We assume that erosion during the formation of the WCP is vertical and similar along the whole WCP surface. Additionally, we assume that the period of abandonment span is short compared to the eustatic cycle period.

We collected samples from the top 5 cm of abraded quartz-rich plutonic bedrock from each WCP (Fig. 4) except for TIV and TV, where we collected thin pieces from the tops of quartz-rich igneous clasts (~10–50 cm in diameter). At each site we collected 2–3 samples within an area of ~6 m in diameter. We have analyzed 3 of the collected samples from each TIB, TII, TIII, TIV and TV site and 2 from TIV site.

As the Atacama Desert climate has been arid since ~15 Ma and hyper arid since at least 3 Ma, this region experiences extremely low erosion rates (Alpers and Brimhall, 1988; Hartley, 2003; Kober et al., 2007). Erosion rates reported here are between 0.1e-3 and 1e-3 mm/an (Dunai et al., 2005; Kober et al., 2007). In central Chile, paleobotanical evidence also supports aridification starting at ~15 Ma (Le Roux et al., 2006 and references therein). Additionally, we sampled planar sites on the wave-cut surface without any relief and surface incision and away from gullies. Thus, we present model ages calculated with a zero erosion model, as these are likely a close approximation to the WCP formation age. The parameters for the model ages and the analytical errors are summarized in Table 1. The internal uncertainty reflects only the analytical error and the external uncertainty includes errors associated with the calculated production rates. Geologic processes, such as differential erosion, inheritance, or covering by a debris mantle that has since been eroded away, would all induce further scattering of the data. In fact, these postformation geologic processes introduce the largest source of error in the calculation. At present, we cannot uniquely identify which processes have been active on the surface. At this site, the sparse river network (two main rivers) is older than the terrace formation and constantly incises the sequence of WCPs, developing canyons. Thus, it is unlikely that external or further depositional processes provide significant surface covering on the WCPs, except for colluvial deposits near the scarp. To address any other erosion processes, we present the mean and standard deviation for each WCP (Table 1).

5. Results: ages of WCPs

In our field area, we have sampled exposed bedrock from TIB, TII, and TIII, as there is no significant sediment cover on the abraded WCP.

### Table 1

<table>
<thead>
<tr>
<th>WCP</th>
<th>Latitude (°S)</th>
<th>Longitude (°W)</th>
<th>Distance from shoreline (m)</th>
<th>Elevation (masl)</th>
<th>¹⁰Be production rate (atoms/g/yr)</th>
<th>Shielding factor</th>
<th>Sample factor</th>
<th>Production rate correction</th>
<th>Error in atoms</th>
<th>¹⁰Be concentration (atoms/g)</th>
<th>Age (ka)</th>
<th>Standard deviation (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TIB</td>
<td>30.90</td>
<td>71.66</td>
<td>5</td>
<td>241</td>
<td>21.3</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>20008</td>
<td>20008</td>
</tr>
<tr>
<td>TII</td>
<td>30.64</td>
<td>71.69</td>
<td>5</td>
<td>102</td>
<td>20.1</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>19864</td>
<td>19864</td>
</tr>
<tr>
<td>TIII</td>
<td>30.87</td>
<td>71.68</td>
<td>5</td>
<td>52</td>
<td>38.9</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>19738</td>
<td>19738</td>
</tr>
<tr>
<td>TIV</td>
<td>30.48</td>
<td>71.69</td>
<td>5</td>
<td>18</td>
<td>30.4</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>1.000</td>
<td>19612</td>
<td>19612</td>
</tr>
<tr>
<td>TV</td>
<td>30.80</td>
<td>71.67</td>
<td>5</td>
<td>4</td>
<td>11.9</td>
<td>1.000</td>
<td>1.000</td>
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<td>1.000</td>
<td>1.000</td>
<td>19488</td>
<td>19488</td>
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</table>

All the samples are surface samples. Data are calculated using CRONUS online cosmogenic-nuclide calculators and the production rates are made at the Center for Accelerator Mass Spectrometry at the Lawrence Livermore National Laboratory and normalized to ICN10Be standards prepared by K. Nishiizumi (07KNSTD3110) using a ¹⁰Be half-life of 1.5 Ma. Thus, data do not need to be renormalized before using the online calculation. The average ages of TIB, TII, TIII, TIV and TV are represented with their respective error bars (1σ).
Fig. 5. The average WCP ages based on $^{10}$Be surface exposure dating plotted with the eustatic curve over the last 800 ka (modified from Siddall et al. (2006)). $^{10}$Be ages are represented with their respective error bars ($1\sigma$). Odd numbers above the eustatic curve are marine isotopic stages (MIS) corresponding to interglacial periods and letters correspond to marine isotopic substages.
bedrock. In contrast to studies where sediment covering WCPs has been sampled (e.g. Perg et al., 2001), we are independent of any assumptions regarding the transport history. For these surfaces, we calculate average exposure ages of 679±8 ka for TI, 317±1 ka for TII, and 225±12 ka for TIII (Table 1). In contrast, for TIV and TV we collected pieces of clasts deposited on the top of the WCP. We have assumed that the clasts from the same terrace surface have the same exposure history as the terrace, however they are likely to differ in their prior exposure history and therefore contain different inherited nuclide concentrations. The mean age is 149±20 ka for TIV and 25±14 ka for TV, but for TIV and TV, exposure ages scatter more than for TI, TII and TIII, due to the type of sample. We have assumed that the youngest clast sample age has a lower inheritance and thus, most closely reflects the WCP formation age. Consequently, we take the youngest age of TIV and TV and we sum the internal and external uncertainties of the youngest age as error bar. This model yields ages of 122±14 ka for TIV and 11±2 ka for TV (Table 1 and Figs. 3B and 4).

6. Discussion

6.1. Marine isotopic stage correlations

Comparing our model ages to the eustatic curve of Siddall et al. (2006 and references therein; Fig. 5), it appears that TI age is well correlated with the marine isotopic stage (MIS) 17 (690 ka; Table 2). Similarly, TII was likely last eroded during marine isotopic substage (MIS) 9c (321±6 ka; Table 2).}

**Table 2** Summary table of data for each marine terrace

<table>
<thead>
<tr>
<th>WCPs</th>
<th>Shoreline angle elevation (masl)</th>
<th>Age (ka)</th>
<th>MIS</th>
<th>Age MIS from</th>
<th>Sea level elevation (masl)</th>
<th>Sea level elevation from</th>
<th>Uplift rates (mm/ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TI</td>
<td>425±15</td>
<td>679±8</td>
<td>17</td>
<td>690</td>
<td>0 to −20±10 (−15±15)</td>
<td>Murray-Wallace, 2002;</td>
<td>738±151</td>
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<td></td>
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<td>Pirazzoli et al., 1991;</td>
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<td>Cutler et al., 2003;</td>
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<td></td>
<td></td>
<td>Stirling et al., 2001</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>9c</td>
<td>318±3 to 324±3 (321±6)</td>
<td>−3 to 8 (2.5±5.5)</td>
<td>Schellmann and Radtke, 2004</td>
<td>1158±416</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>7e</td>
<td>230 to 235 (232±2.5)</td>
<td>−5 to −15 (10±5)</td>
<td>Antonioli et al., 2004;</td>
<td>389±149</td>
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<td></td>
<td>Li et al., 1989;</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Schellmann and Radtke, 2004</td>
<td>103±69</td>
</tr>
<tr>
<td>TIV</td>
<td>25±3</td>
<td>123±14</td>
<td>5e</td>
<td>116±1 to 128±1 (122±7)</td>
<td>0 to +6 (3±3)</td>
<td>Fleming et al., 1998;</td>
<td>1667±434</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>Lambeck et al., 2002</td>
<td></td>
</tr>
<tr>
<td>TV</td>
<td>6±1</td>
<td>11±2</td>
<td>1</td>
<td>6±1</td>
<td>−3 to −5 (−4±1)</td>
<td>Fleming et al., 1998;</td>
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<td>Lambeck et al., 2002</td>
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Shoreline angle elevation and 1⁰Be age of marine terraces, correlations to MIS, age of MISs in literature, elevation of the former sea level during MIS and uplift rates calculated from Eq. 1. See text 5.1 and 5.2 for full explanation of MIS correlation and calculation of uplift rates. Uplift rates are calculated over a time interval between two WCPs. Data of MIS ages and respective sea level elevation are from Siddall et al. (2006) and references therein, Fleming et al. (1998) and Lambeck et al. (2002), as indicated in the sixth and eighth column.

**Fig. 6.** WCP shoreline angle elevation vs. last sea-level highstand ages. The complete curve corresponds to uplift rates for the last 700 ka (T1) and thus represents the timing of the uplift of the Andean coastal forearc in the study area. The top curve represents the cumulative history undergone by each marine terrace, which is derived from combining each measured segment (or time interval) of the curve. Ages and magnitudes of sea-level highstands used in the calculation of uplift rates are based on Siddall et al. (2006) and references therein, Fleming et al. (1998) and Lambeck et al. (2002). Solid lines correspond to the uplift rates calculated using equation 1 (modified from Lajoie (1986)) based on the ages for MIS 1 (6±1 ka and −4±1 masl), MISS 5e (122±7 ka and +3±3 masl), MISS 7e (232±2.5 ka and −10±5 masl), MISS 9c (321±6 ka and +2.5±5.5 masl) MIS 17 (690 ka and −15±15 masl). See text 5.2 for further explanations.
6 ka), and abandoned as a result of the sea regression during MISS 9b (Table 2 and Fig. 5). TIII is a bit less straightforward and could have last been eroded during MISS 7e (232.5±2.5 ka) or 7c (216±4 ka). However, as we present zero erosion 10Be model ages, it is important to note that the effect of any erosion would underestimate the age of the sample. Therefore, we suggest that TIII likely was formed during MISS 7e and was subsequently abandoned at the onset of the MISS 7d sea regression. TIV was last eroded during MISS 5e (122±7 ka), a period which corresponds to the most common marine terrace formation age worldwide, and was likely abandoned at the onset of MISS 5d (Table 2 and Fig. 5). For TIV, correlation to an MIS is less clear. Indeed, no sea-level highstands clearly correspond to 11 ka. The closest highstand occurred at 6±1 ka and corresponds to MIS 1 (Fleming et al., 1998; Lambeck et al., 2002; Table 2 and Fig. 5). If our samples have inherited radionuclide concentrations, then the real age of the marine terrace should be younger than 11±2 ka. Thus, it is at least plausible to assign TIV to MIS 1, however, further data need to be collected to confirm this assumption.

In support of our age assignments, we note that MISs 1, 5, 7, 9 and 17 are all interglacial periods corresponding to maximum sea-level highstands (Fig. 5). 10Be ages have a good correlation with sea-level highstands, which confirm our geomorphologic model of WCPs formation. However, we do not observe any dated WCPs between MIS 17 (690 ka) and MISS 9c. Our interpretation of this sequence is that any WCPs that were formed during these sea-level highstands were eroded before MISS 9c (Fig. 5). The lack of scatter of the ages on the T1, TIII, and TIV surfaces gives us confidence that the strategy of sampling from in situ-abraded bedrock is the best approach to date WCPs. Furthermore, as the standard deviation of the mean model ages from a given surface is small, we infer that 1) by sampling in situ-abraded bedrock, we have minimized the effects of potential inherited nuclide concentrations and 2) geologic processes acting on the surfaces following abandonment are minimal. This latter interpretation is supported by the aridity along the Central Andean Pacific coast which aids the preservation of geomorphic features in this area. The ages from TIV and TIV are more scattered than those of the three other WCPs, plausibly because we sampled boulders (in contrast to bedrock for the other WCPs) which could contain different inheritances.

6.2. Calculation of uplift rates

We calculate the mean uplift rate over a time interval $t_i - t_j$ ($\Delta t$), using the sea level corrected elevation of the WCP (subtracting the
elevation of the sea-level highstand to the present-day elevation of the WCP) and applying the following equation (modified from Lajoie (1986)) to each level of WCP (Table 2).

\[
\text{UpliftRate}_{ij} = \left( \frac{(ShA_i - MH_j) - (ShA_j - MH_j)}{(Age_i - Age_j)} \right)
\]

(1)

Where \(ShA\) is the present-day elevation of the shoreline angle of the older WCP (time \(t_i\)), \(MH\) is the elevation of the sea-level highstand at \(t_i\) compared to the present sea-level, \(ShA\) is the present-day elevation of the shoreline angle of the younger WCP (time \(t_j\)), \(MH\) is the elevation of the sea-level highstand at \(t_j\) compared to the present sea-level, \(Age\) is the age of the WCP from MIS correlations \((t_i)\) and \(Age\) is the age of the WCP from MIS correlations \((t_j)\). Our use of the MIS ages is a direct product of the fact that the geomorphic model of WCP formation considers that the marine terraces were formed during sea-level highstands (Lajoie, 1986; Anderson et al., 1999). We use the marine record (cf. Siddall et al., 2006) to correlate our \(^{10}\text{Be}\) model ages to these highstands. If we follow the classical assumption of constant uplift, the mean uplift rate would be 638 ± 43 mm/ka since MIS 17 (690 ka), using the shoreline angle elevation of the older WCP \((t_j)\) and age \(t_j\) from MIS correlation. Over a time interval between two WCPs, we obtain uplift rates (Table 2 and Fig. 6) of 738 ± 151 mm/ka over the time interval \(\Delta t\) of 690 \((t_i)\) = 321 ± 6 \((t_j)\), 1158 ± 416 mm/ka over the time interval 321 ± 6 \((t_j)\) = 233 ± 3 \((t_j)\), 389 ± 149 mm/ka for the time interval between 233 ± 3 \((t_j)\) = 122 ± 7 \((t_i)\), 103 ± 69 mm/ka for the time interval 122 ± 7 \((t_i)\) = 6 ± 1 \((t_i)\), and 1667 ± 434 mm/ka for 6 ± 1 \((t_i)\) to the present-day \((t_i)\). The errors on the uplift rates include the errors involved in the age and elevation of highstands and shoreline angle elevation of WCPs. We note, the uplift rate of 1667 ± 434 mm/ka since 6 ka has an error that could be as much as factor of 2, if the 11 ka age is correct. These calculated uplift rates, compared to the mean uplift rate since 690 ka (more than 2 times lower), demonstrate the variability of uplift during time (over 700 ka). This result leads to the conclusion that a mean uplift rate should not be used to validate long-term uplift estimations.

The good correlation of the \(^{10}\text{Be}\) ages with the sea level curve demonstrates the validity of the zero erosion hypothesis. Finally, as \(^{10}\text{Be}\) ages show a good correlation with sea-level highstands, the hypothesis of the initial zero CRN concentration and rapid regression of the sea level in comparison to the exposure time is consistent with our geomorphologic model of WCP formation.

6.3. Comparison of chronologic data

The chronology reported here differs somewhat from that reported by Ota et al. (1995) who assigned ages of ≈330 ka, ≈330 ka, 300 ka and 125 ka to \(t_0\), \(t_1\), \(t_2\) and \(t_3\) respectively. This discrepancy likely reflects the differences in our dating methods. Ota et al. (1995) proposed ages for the WCPs by geographically correlating the preserved marine terraces in the Altos de Talinay area with those in the Coquimbo Bay area, 100 km to the north (29.9°S), where obtained ages are based on mollusk shell aminostratigraphy and electron spin resonance. However, the morphostratigraphic correlation between the two sets of terraces is not straightforward considering 1) the distance between the sites, 2) the site locations with respect to the regional tectonic setting, and 3) the numerous faults that locally displace the WCPs. Marine terraces lie at much lower elevations in the Coquimbo Bay compared with those along the Altos de Talinay. Furthermore, the terraces in the Coquimbo Bay are described to have undergone subsequent marine reoccupation (Radtke, 1987; Leonard and Wehmiller, 1992) whereas we do not observe any reoccupation in our new data set along the Altos de Talinay area. Shells in marine terraces are notorious for evolving in open systems and thus may be subject to post deposition contamination that prevents their use as a chronologic indicator (Ortlieb et al., 1992; Pedoja et al., 2006). Thus, this method does not provide a robust means of absolute dating, in contrast with the CRN method. Finally, the strong agreement between our ages and the sea-level highstands supports both the recent models of WCP formation (e.g. Lajoie, 1986; Anderson et al., 1999) as well as our \(^{10}\text{Be}\) chronology.

6.4. Temporal variability of uplift

Based on \(^{21}\text{Ne}\) ages of marine terraces in the Caldera–Bahía Inglesa area (27°S), 350 km to the north of our study area, Quezada et al. (2007) report a gap between 780 ka (MIS 19) and 425 ka (MIS 11). Quezada et al. (2007) proposed that the absence of marine terraces during some periods of sea-level highstands could be an indicator of (1) low uplift rates during that time interval, (2) weak interstadials during this interval and/or (3) stronger erosion of the littoral topography during the MIS 11 sea-level highstand (425 ka). We also find an absence of marine terraces over a similar time interval in the Altos de Talinay area, but the ages of the preserved marine terraces (MIS 17 and MIS 9) are not the same as those in the Caldera–Bahía Inglesa area.

In our study area, MIS 11 is not recorded in the WCP succession although it was a major sea-level highstand. This suggests that local tectonic factors (in this case low uplift rates during this time interval) may temporarily have been the dominant control on the coastal morphology. The different chronologies of marine terrace formation in the Altos de Talinay and Caldera–Bahía Inglesa areas show that the preservation of these terraces is not simply controlled by the magnitude of interstadials. Quezada et al. (2007) proposed that stronger erosion during MIS 11 could explain the absence of marine terraces during MIS 13 to MIS 17 in the Caldera–Bahía Inglesa area. Applying the same model to the Altos de Talinay area requires stronger erosion processes during MIS 9. Why marine erosion would have been particularly efficient during MIS 11 in Caldera, and MIS 9 in the Altos de Talinay, is not clear. Thus, we interpret that the absence of marine terraces corresponding to an interval of major sea-level highstands is mainly related to local tectonic processes and indicates low and irregular uplift rates. As we noted earlier, two factors control the morphological evolution of the coastal forearc: 1) the rate of coastal uplift and 2) the magnitude of sea-level highstands. If the uplift rate is large, the WCP was formed during the previous highstand and is displaced high enough above the coastline and can be preserved even given a subsequent highstand. In contrast, if the uplift rate is small, the old WCP, being close to sea-level, is easily eroded during the following highstand (Fig. 7). In fact, when a WCP is preserved in the landscape, it means that the coastal uplift has been faster than marine erosion and that tectonic uplift was the predominant factor that controlled the morphology of the coastal forearc. Whereas, when a WCP is missing in the landscape, the amount of erosion during the subsequent sea-level highstand was the predominant factor, suggesting a moderate or minor uplift rate for that interval (Fig. 7).

In the present study area, there have been at least three distinct periods of morphology formation since 700 ka. In the first (following MIS 17) and third (following MISS 9c) periods, the coastal morphology has been dominated by tectonic uplift. In contrast, during the second time period (during ~250 kyr between MIS15 and MISS9), marine erosion has been predominant and terraces that formed during MIS 15, 13 and 11 were eroded by the MISS 9c highstand. This suggests that the uplift rate was lower between MIS 17 and 9, which resulted in the erosion of previous marine terraces during this sea-level highstand (Fig. 7). Nevertheless, no direct evidence proves that MISS 9c eroded all of the marine terraces formed during MIS 15, 13 and 11. Alternatively, another end member model may be that each terrace formed during MIS 15, 13 and 11 was eroded away during the subsequent sea-level highstand: the marine terrace formed during MIS 15 could have been eroded during MIS 13, the marine terrace formed during MIS 13 could have been eroded during MIS 11 and finally the marine terrace formed during MIS 15 could have been eroded during MISS 9c (Figs. 5 and 7).

Fig. 6 shows the uplift rate vs. time in the Altos de Talinay area since 700 ka. The solid line corresponds to the average uplift rates
calculated between each preserved WCP; gray boxes represent the corresponding error bars. The absence of some marine terraces suggests that the uplift rate has been null or low between MIS 15 and 9. The dashed line on Fig. 8 corresponds to a model of uplift compatible with these data and explains the absence of marine terraces related to MIS 15, 13 and 11. As MIS 15, 13 and 11 terraces are missing from the landscape, the uplift rate calculated between MIS 17 and 9 is a mean rate. However, the uplift rate during this period is likely more variable than that represented by the overall mean rate.

Finally, we have identified at least two periods of rapid uplift since 700 ka, one between MIS 17 and 15 and one between MIS 9 and 7. The other possible period of rapid uplift, i.e. that since 6 ka, needs further confirmation before any firm conclusions can be drawn (Fig. 8).

6.5. Causes of uplift rate variability

The difference in the records from the Altos de Talinay (an absence of MIS 17 and MIS 9 surfaces) and that at Caldera–Bahía Inglesa (missing MIS 19 and MIS 11 surfaces), suggests to us that uplift rate varies not only temporally, but also spatially along-strike. The causes of these tectonic pulses at a specific location should also be related to local coastal tectonic processes.

The WCPs were developed indifferently on a heterogeneous substratum. Therefore, lithologic variations in bedrock do not play a major role in the morphology of these terraces. The observation that T1 is more strongly deformed by normal faults than T2 together with the observation of no fault displacement affecting Tm, TIV and TV (Fig. 4) suggests that fault activity is confined to time periods when the uplift rate is rapid (Fig. 8). Along a fault cutting different WCPs, vertical offsets decrease toward the youngest WCPs, with no fault displacement on Tm, TIV and TV. Thus, local faulting in the Altos de Talinay area sometimes occurred after the formation of T1 to sometime after the formation of Tm, and there is no evidence that these faults have been active since the formation of Tm. The deformation related to faulting is also spatially variable in the Altos de Talinay area. Here, varying amounts of displacement are accommodated by the multiple faults along the coast illustrating that the area has had a complex Pleistocene tectonic history (Ota et al., 1995; Heinze, 2003).

One plausible explanation for the variability in the coastal morphology and surface uplift rates that we observe is that tectonic deformation in the forearc is strongly influenced by slab subduction processes. Indeed, the roughness of the subducted plate (e.g. irregularities, seamounts) can lead to subduction erosion and underplating of material (e.g. Hartley et al., 2000; Cleft and Hartley, 2007; Encinas et al., 2008). These processes in turn are associated with offshore subsidence and onshore uplift of the active margin (Lallemand et al., 1994; Delouis et al., 1998). Cleft and Hartley (2007) proposed that crustal thickening by tectonic underplating of subducted materials under the coastal regions occurs at the same time as tectonic erosion close to the trench (see also Adam and Reuther, 2000; Laursen et al., 2002; von Huene and Ranero, 2003) and thus material accreted eastward of the continental coast can explain the shift in the stress regime, from compression at the base of the forearc wedge, to extension close to the surface (Delouis et al., 1998; Adam and Reuther, 2000). In fact, deep accretion below the coastal part of the forearc could explain the uplift contemporaneous with the activity of normal faults that we observe in the Altos de Talinay area. In this context, we suggest that the uplift rate variability and tectonic deformation are related to pulses of accretion of subducted materials at the base of the forearc wedge. This results in rapid and episodic uplift, contemporaneous with the formation or activation of normal faults. Periods of lower uplift rates might alternate with periods of higher uplift rates over time scales <1 Myr due to topographic irregularities on the subducting plate or material in the trench. Thus, the changes in the Pleistocene uplift rates and activity of normal faults in this coastal part of the Chilean forearc may be a general phenomenon along oceanic–continental subduction zones. This implies that the estimation of long-term rates from geomorphic markers on intermediate time scales may be biased by these rate changes at the 106 yr timescale.

Nevertheless, accretion of subducted materials at the base of the forearc wedge is not the only possible mechanism responsible for strong variation in uplift rates of marine terraces in the Altos de Talinay area. With respect to the presence of marine terraces above different tectonic settings along the Andean coast, another plausible explanation for uplift rate variability could be related to the variation of the down-dip position of the locked zone along the plate interface with respect to the continental forearc. Further data under process will be required to confirm this hypothesis and determine which mechanism controls the coastal uplift kinematics.

7. Conclusion

During the last decade, the widespread application of CRN dating methods has allowed us to decipher the geomorphic evolution of coastal areas. But still, as observed in alluvial environments, the depositional processes on marine terraces remain difficult to access. In contrast, in this study, we show that the sampling strategy from situ-bedrock is the best approach to date WCPs using 10Be surface exposure dating. The systematic sampling of successive WCPs demonstrates that it is not always suitable to correlate geomorphic markers based on chronostratigraphic observations. In the Altos de Talinay area, we show that the abandonment ages of WCPs correlate with sea-level highstands. We assign the five WCPs preserved in the Altos de Talinay area to the MIS/5s 1, 5e, 7c, 9c and 17. These new data do not support models of WCP formation that suggest the present-day elevation of WCPs above the modern sea-level has resulted from long-term, relatively slow, steady-state tectonic uplift of the continental margin (Cantalamessa and DiCulmo, 2004). Rather, we observe that both major Pleistocene eustatic changes and coastal uplift have had a strong control on the formation of geomorphic markers within the coastal forearc. We show as well, that recent uplift rates in the Chilenian coast have been highly variable during the last 700 ka at time scales intermediate between those obtained using geodetic methods (~10 yr) and those based on geological data (105 yr). The variability in the uplift rate we observe is likely due to localized processes linked to subduction, such as subduction erosion and underplating of subducted materials under the coastal zone. This highlights the importance of identifying several datable geomorphic markers over the timescale of interest when attempting to assess regional or local uplift rates.

Acknowledgements

This research project is led thanks to NSF EAR grant 0345895, the Institut de Recherche pour le Développement (IRD) and IGPP-LLNL. We thank S. Carretier for helpful discussions. We also thank O. Oncken, C. Garzione and an anonymous reviewer for constructive and critical comments of this manuscript.

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