Multiproxy analysis of tsunami deposits—The Tirúa example, central Chile

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ABSTRACT

The study of prehistoric tsunami deposits enables us to appraise the recurrence frequency and improve risk assessments of coastlines. We present a study of tsunami deposits (A.D. 500 to present) exposed at Tirúa, central Chile. Analyses of various proxies such as grain size, petrography, geochemistry, and diatoms were applied to the tsunami deposits intercalated in the uppermost 2 m of the Tirúa River floodplain (1.2–2 km inland) in order to distinguish these tsunami deposits from the surrounding river marl sediment layers. The tsunami sand layers are characterized by erosional bases and landward thinning and fining, and they consist of well-sorted, unimodal sand commonly >3Φ. The fluvial marl sediments, in contrast, exhibit polymodal grain-size distributions with a distinct fraction commonly <3Φ. However, some older tsunami sand layers (sand layers 4 and 5) exhibit similar grain-size characteristics as the surrounding river marl sediments, implying environmental changes. The diatom data also indicate environmental changes caused by neotectonic movement, at least for the lowermost sand layers of the profile (sand layers 4, 5, and 6). However, all six sand deposits are enriched in Ca, Si, Sr, Ti, and Fe and depleted in Al, K, and Rb, and they yield low loss on ignition (LOI) values, resulting from heavy mineral accumulations in these layers. Based on the bulk of these findings, in combination with an optically stimulated luminescence (OSL) age inversion in sand layer 5, all six sand layers in the floodplain succession at Tirúa are interpreted as tsunami deposits. They represent six different events connected to vertical neotectonic movements that occurred at the central Chilean margin during the last 1500 yr, extending the historical tsunami record in central Chile over 1000 yr.

INTRODUCTION

The study of deposits of recent tsunami events (e.g., the tsunami of the Mw 8.8 2010 Maule earthquake) is a major aspect of geological research aimed at understanding how and when tsunami inundations have impacted coastlines in the past. Studies of recent tsunami events focus on understanding how tsunamis impact coastlines, while studies of paleotsunami deposits, which are older than historical accounts, can provide important temporal information of tsunami impacts. The central Chilean coast between 27°S to 41°S is exposed to tsunamigenic earthquakes (Mw >7.5) at a high frequency (25–90 yr) due to the frequent and strong seismic activity of the Peru-Chile subduction system (Fig. 1; Lockridge, 1986; NOAA, 2015). This high frequency of tsunamigenic earthquakes is documented by historical and modern geologic and eyewitness records (Lomnitz, 2004; International Oceanographic Commission, 2014). Knowledge of the prehistorical tsunami record, however, is incomplete (e.g., Cisternas et al., 2005; Nentwig et al., 2015). The study of paleotsunami deposits represents the only means by which the tsunami history of any active continental margin can be extended beyond the historical record.

In central Chile, the historical tsunami record extends back to the mid-sixteenth century (Lomnitz, 2004; Lockridge, 1985). This chronology documents tsunamigenic events affecting central Chile in the years A.D. 1562, 1730, 1835, 1837, 1960, and 2010. The list does not include seismic events producing small-scale and moderate tsunamis (run up heights <4 m; Lockridge, 1985). Some additional tsunami events that occurred further north in northern Chile and Peru have also caused severe damage along the coast of central Chile. These events include the A.D. 1586 and 1746 Callao earthquakes (run up heights in both cases: ~24 m), the 1868 Iquique earthquake (run up height: ~15 m in Arica), and the 1877 Arica earthquake (run up height: ~24 m in Tocopilla; Lockridge, 1985). When reconstructing the geological tsunami record, it should also be considered that in addition to local small-scale tsunamis, both near-field and far-field tsunamis can be responsible for leaving behind tsunami sediments to the same extent as major central Chilean tsunamis.

The determination of an onshore tsunami record depends on the unambiguous recognition of tsunami deposits. A combination of several indicators for individual extreme marine inundation events is required, including shell fragments, sedimentological features like grading as evidence of deposition from suspended load, and erosive bases and rip-up clasts as evidence of erosion and entainment of the underlying soil, etc. (e.g., Dawson and Shi, 2000; Dawson and Stewart, 2007; Jaffe and Gelfenbaum, 2007; Peters and Jaffe, 2010).

However, there are no typical characteristics that differentiate a tsunami deposit from other sediments that might yield similar characteristics due to deposition from high-energy environments such as storm surges or extensive fluvial flooding (e.g., Dawson and Shi, 2000; Peters and Jaffe, 2010). To enhance the accuracy of the identification of tsunami deposits, it is productive to combine the sedimentological analysis with a set of complementary methods (e.g., Goff et al., 2010; Chagué-Goff et al., 2011; Ramirez-Herrera et al., 2012).
Figure 1. (A) Overview and (B) location of Tirúa in central Chile, showing coast-parallel rupture zones of the 2010 and 1960 earthquakes. (C) Satellite image (Google Earth, 2015) of Tirúa locality and surrounding environments. (D) Sampling locations on satellite image (Google Earth, 2015) on the river floodplain, which (E) was covered by a gray sand sheet after the tsunami impact in February 2010 (gray arrows mark the landward direction of bent grass) and which (F) contains different sand intercalations partially eroded by the Rio Tirúa, as indicated by horizontal scour lines within the river marsh sediments along the riverbank.
Such methods include (1) analysis of micropaleontological assemblages, e.g., diatoms and foraminifers, including sediment deoxyribonucleic acid (DNA) records, (2) geochemical analysis of different marine proxies in combination with sediment sources, (3) analysis of isotopes, elemental ratios like C/N, and biomarkers, (4) analysis of luminescence characteristics of tsunami deposits, (5) analysis of grain-size surface microtextures, and (6) study of magnetic susceptibility (e.g., Hemphill-Haley, 1996; Huntley and Clague, 1996; Bruzzi and Prone, 2000; Hawkes et al., 2007; Wassmer et al., 2010; Chagué-Goff et al., 2011, 2012; Szczuciński et al., 2016).

Not all tsunami deposits will be preserved in the geological record. Erosion, secondary changes in grain sizes, and bioturbation are some of the processes that can modify or delete geological traces of tsunamis within a few years after deposition (Szczuciński, 2012; Spiske et al., 2013; Bahhurg and Spiske, 2015). Therefore, it is important to study deposits of recent tsunami events in post-tsunami surveys but also to employ a set of complementary methods in the study of paleotsunami deposits.

One locality of particular interest for this kind of study is the vicinity of the village of Tirúa on the coast of central Chile. This locality has been the focus of several posttsunami surveys after the tsunami from the Mw 8.8 Maule earthquake in February 2010, due to locally extremely high run-up heights (25–30 m; e.g., Fritz et al., 2011; Bahbhub and Spiske, 2012; Ely et al., 2014). Field studies after the 2010 tsunami at Tirúa revealed the presence of a sedimentary record with more than one sand layer of potential tsunami origin (Ely et al., 2014; Nentwig et al., 2015). In order to ascertain the tsunami origin of the six sand layers dated by Nentwig et al. (2015), we obtained complementary data by analyzing the spatial distribution, granulometry, geochemical composition, and the abundance of marine diatom assemblages of these layers. This was combined with an analysis of quartz grain surface microtextures (Bellanova et al., 2016). Here, we present implications drawn from the different analyses that identify these as tsunami deposits and place them within the temporal framework established using the quartz-based single-aliquot regenerative dose (SAR) optically stimulated luminescence protocol by Nentwig et al. (2015). This geological tsunami record exceeds the local historical tsunami record by over 1000 yr and therefore allows for a more precise calculation of tsunami recurrence intervals for the past 1500 yr. Additionally, this tsunami record offers invaluable information about the characteristics of past onshore tsunami inundations, improving our current understanding of coastal tsunami impacts.

## STUDY AREA

Tirúa is located 200 km south of the city of Concepción on the central Chilean coast (Figs. 1A and 1B; Table S1 in the Supplemental Material). Here, the Rio Tirúa cuts through the Cordillera de la Costa, which rises steeply from a coastal cliff to altitudes of ~700 m. The Cordillera de la Costa mainly consists of metamorphic belt and accretionary complex (Hervé, 1988; Willner, 2005). This unit was intruded by granitoid batholiths associated with volcanosedimentary successions that jointly represent the Mesozoic magmatic arc of this active margin (Hervé, 1988; Willner, 2005; Parada et al., 2007). Both assemblages are in turn overlain by Neogene marine sedimentary units (Encinas et al., 2006, 2012).

Tirúa lies north of the Tirúa River and its estuary (Fig. 1C). The main stream has its source ~30 km SE in the vicinity of Las Araucarias, at ~700 m altitude in the Coastal Cordillera. In the lower course of the river, at altitudes of less than 50 m, the Rio Tirúa has eroded a valley less than 1 km wide (Fig. 1C). Here, the narrow river has developed a floodplain that is used for agricultural purposes, but there are no further anthropogenic constructions or infrastructure (Figs. 1C, 1D, and 1E). Observations during field work indicate that the estuary of the Rio Tirúa has a semidiurnal microtidal range upstream up to couple of kilometers. These tidal movements and currents have led to sharp and steep riverbanks with average heights of 1–1.5 m and several linear-horizontal scours resulting from the differential erosion of sands and silts of the floodplain (Fig. 1F).

The Tirúa locality was surveyed in February 2013 and 2014 after recent traces of the February 2010 tsunami had been briefly described at this locality in a posttsunami survey by Fritz et al. (2011), Bahbuhr and Spiske (2012), and Ely et al. (2014). Maximum tsunami inundation distances of 1.5 km upstream along Rio Tirúa were documented by Bahbhub and Spiske (2012) by mapping a sheet of gray sand with thicknesses of 10–18 cm deposited by the tsunami on the river floodplain. Ely et al. (2014) described the same sand deposits but noted small boulder-sized building rubble. Both survey teams documented the landward flow direction of the tsunami on the floodplain using evidence from bent vegetation (Fig. 1E).

Tirúa, located between the Arauco-Longuimay segment in the north and the Valdivia-Liguine segment in the south, has been affected by megathrust earthquakes occurring along different segments of the Chilean subduction zone, including the 1960 Mw 9.5 Valdivia earthquake (Fig. 1A; Lomnitz, 1970; Hackney et al., 2007; Ely et al., 2014). Resulting neotectonic changes are documented by the uppermost four sand layers and indicate alternating uplift (sand layers 1 and 3) and subsidence (sand layers 2 and 4) events based on quantified changes in diatom assemblages in combination with estimations of the changes in relative mean sea level associated with the 2010 Maule earthquake at Tirúa (Ely et al., 2014).

We studied five distinct sand layers in addition to the gray sand layer of the 2010 tsunami event for the Tirúa River floodplain area located ~1.2 km upstream from the coastline and estuary (Fig. 1C).

## METHODS

### Field Analysis and Sampling

Samples were obtained from eight riverbank profiles on the river floodplain, 37 drill holes (percussion coring), and one trench for OSL sampling (Fig. 1D). Sampling along the riverbank was carried out in order to...
map the different stratigraphic units of the river floodplain. The riverbank profiles varied in depth (40–140 cm below ground level, depending on accessibility of the banks during low tide) and were positioned at the western and northern parts of the river floodplain (Fig. 1D). A grid of 27 drill holes was completed along the western bank side (Fig. 1D). The core grid consisted of four transects perpendicular to the river with percussion cores taken every 30 m, reaching 120 m inland (Fig. 1D). The southernmost transect (T1: TIR 1 to TIR 7) and all four riverbank profiles (TIR 1, 9, 16, and 22) were additionally sampled with separate liner cores (1 m PVC tubes; Fig. 1D). The core depth generally varied from 2 to 6 m, of which the uppermost 2 m turned out to reflect the crucial part of the profile in terms of potential tsunami deposits. For the coring, a two-stroke motorized Wacker vibrocorer with 6-cm-diameter probes was used.

Prior to sampling, all drill cores were described in detail. We documented color, macroscopically visible components, organic contents, and sedimentary features and structures, e.g., layer contacts, etc. Sampling of the outcrops and cores was mainly undertaken for grain-size analysis, whereas the sampling for geochemical and diatom analyses was conducted from opened liner cores in the laboratory. Digital high-resolution imaging was performed on the liner cores of the first transect with a GEOSCAN color line scan charge-couple device (CCD) camera (3 × 1024 pixel CCD arrays) incorporated in a multisensor core logger at the Center for Marine Environmental Sciences (MARUM), University of Bremen, Bremen, Germany.

Each stratigraphic unit was sampled at representative parts of each lithology found in the profile, whereas potential tsunami sand layers were sampled multiple times vertically if the layer thickness was sufficient to document changes in composition and grain size. Reference samples from different facies adjacent to the river floodplain were also sampled in order to obtain comparative data in terms of composition and grain-size distribution. These samples were taken from the marine intertidal (TIR R1), beach (TIR R2), dune crest (TIR R4), and river intertidal (TIR R3, proximal, TIR R5, intermediate distance, and TIR R6, distal along the river to the coastline; Fig. 1C; Table S1 [see footnote 1]) areas.

Laboratory Analysis

After organic compounds were removed and samples were dried, grain aggregates in the samples were gently disaggregated by sieving and low-pressure pestling, so that small representative amounts (~1 g) of each sample could be separated for grain-size measurements (distributions and statistics) with a settling tube according to Poole (1957), Cook (1969), and Gibbs et al. (1971). Grain-size distributions and statistics were computed using the software SeTuPs by Cheng and Weiss (2011) based on Stokes’ law, as discussed in Rubey (1933) and Ferguson and Church (2004).

Samples for taxonomic and quantitative diatom assemblage analysis were taken from all potential event layers and from the underlying and overlying soil (from the drawn liner cores). A proximal core and a distal core (proximal core: TIR 2, 0 m distance to the riverbank; distal core: TIR 6, 120 m to the riverbank) were sampled to check for landward changes in the diatom assemblages. In total, 132 samples with diatoms were processed and analyzed at the Faculty of Geosciences at the University of Szczecin, Poland. The samples were dried and weighed, and organic and carbonatic matter was removed. Next, they were thoroughly cleaned of additives used for removal and settled onto cover slips (three per sample) for sublimation. Then, the cover slips were mounted onto slides, from which quantitative analysis (counting 300 valves of each sample) of the diatoms could be performed under a light microscope at 1000x magnification according to Schrader and Gersonde (1978) and Witkowski et al. (2012). The identification and determination of salinity preferences of diatoms were based on: Metzeltin and Lange-Bertalot (1998), Rumrich et al. (2000), Witkowski et al. (2000), Sar et al. (2003), Metzeltin et al. (2006), Horton et al. (2011), Chuh-Hwan Koh et al. (2012), Sánchez et al. (2012), and Trobajo et al. (2012).

Energy-dispersive X-ray fluorescence (EDXRF) was applied for quantitative geochemical analysis. Samples of the potential tsunami sand layers and the river marsh sediments from cores TIR 2 (proximal riverbank core) and TIR 6 (distal core) were processed to fine powders for the determination of the loss on ignition (LOI) and elemental concentrations (measured from fused beads produced from the sample powder and a binding additive) according to Brouwer (2010). Furthermore, a petrographic analysis of selected samples was performed using grain mounts with a representative amount of each sample (~25 µm thick) without performing a density separation beforehand. The Gazzi-Dickinson point-counting method was applied for a quantitative determination of framework components (Dickinson et al., 1983; Ingersoll et al., 1984). The content of total inorganic carbon (TIC) was measured to further pursue the source of potential Ca fluctuations within the profile. This measurement was performed via infrared-spectroscopy of CO2 (the gas was liberated with 25% HCl at 70 °C) with a CS-Mat 5600 (Ströhlein Instruments, Berlin, Germany). Calcium carbonate (Merck reagent grade CaCO3, 12.00 ± 0.36 wt%) was used as a standard reference material to monitor the accuracy and precision of these measurements.

Further details of the laboratory analyses of grain sizes, diatoms, and geochemical properties can be found in the Supplementary Material (see section on Laboratory Analysis [footnote 1]).

**RESULTS**

**Sedimentological and Grain-Size Analysis**

**Stratigraphy and Grain-Size Analysis of Profile TIR 1**

Profile TIR 1 is 1.40 m in depth and contains six different gray layers of loose sand interbedded in beige-brownish river marsh sediments composed of sand and silt (Fig. 2). Grass grows from the topmost gray sand layer (sand layer 1), which varies in thicknesses between 4 and 18.5 cm.
Figure 2. Riverbank profile (TIR 1) of Tirúa with six different sand intercalations within the river marsh sediment and the associated grain-size distributions of sand layers 1, 2, 3, and 6 (dark gray). The grain-size distributions of the river marsh samples below the mentioned sand layers are also displayed, showing the poly-modal distributions in comparison to the unimodal distributions of the sand intercalations. For orientation, a dashed line is shown in the plots at the grain size of 3φ. Details for the profile section within the dotted square are discussed in Figure 3.
Underneath the gray sand layer 1, there is a brownish-gray soil containing sand and silt (Fig. 2). The marsh sediments frequently contain roots of recent vegetation and root debris. An additional five sand layers are visible at depths below ground level (Fig. 2) of: 13–15 cm (sand layer 2), 25–36 cm (sand layer 3), 53–55 cm (sand layer 4), 72–85 cm (sand layer 5) and 122–124 cm (sand layer 6).

The sediment units are mostly structureless, except for sand layer 5 (72–85 cm), which contains bright brownish lenses of finer sediment several centimeters in length (Figs. 2 and 3). These features are interpreted here as small rip-up clasts (Figs. 2 and 3). Most sand layers exhibit erosional bases, but they are less pronounced in sand layers 2 and 4 (Fig. 2). These erosional bases are noticeable by the sudden change in grain size, consistency, and color (Fig. 2). They range in scale from millimeters to a few centimeters in depth, and some of these erosional bases can be traced laterally along the riverbank upstream (several hundreds of meters) as the scour lines in the riverbank imply (Figs. 1D, 1F, 2, and 3).

The grain-size distributions of samples along the riverbank of the top sand layer 1 are unimodal, with mean values ranging from 2.1e to 2.2e (Fig. 2; Fig. S1, Table S2 [footnote 1]). This contrasts with the river marsh samples beneath sand layer 1, which yield mostly polymodal grain-size distributions with mean values of 2.8–3.2e (Fig. 2; Fig. S1, Table S2). Another difference in the grain-size distribution between these two different sediments is the fraction of >3φ present in the river marsh sediments, which the sand layer 1 generally lacks (Fig. 2; Fig. S1, Table S2). This distinction between the sand layer and the river marsh sediment is also observed in sand layer 3 and to some extent in sand layers 2 and 6 (Fig. 2). Sand layers 2 and 6 have unimodal distributions and contain a small fraction of the sediment larger than 3φ (Fig. 2; Table S2). Sand layer 3 has a mean of 2.4φ, which is slightly coarser than those of sand layers 2 and 6, with a mean of 2.7e (Fig. 2; Table S2). However, the underlying and underlying river marsh sediments of sand layers 2 and 6 have means mostly >3φ, ranging from 3.0 φ to 3.4φ (Fig. 2; Table S2). Therefore, these two sand layers can also be differentiated from the surrounding river marsh sediments by their grain-size distributions. Samples from sand layers 4 and 5 and their surrounding river marsh sediments yielded unimodal grain-size distributions, with mean values ranging from 2.3φ to 2.6φ and no significant fraction exceeding 3.0φ (Fig. 3). Clear distinction between these two sand layers at depths of 50–85 cm in TIR 1 from the river marsh sedimentation is therefore not as apparent as in the cases of the other sand layers (Figs. 2 and 3). Slight changes of color, the scouring along of the riverbank, and some small rip-up clasts in sand layer 5 indicate the presence of two sandy intercalations (sand layers 4 and 5) in this section of the profile (Figs. 2 and 3). Further grain-size parameters were consulted for the distinction of the sand layers and river marsh sediments: (1) A minor distinction could be made by using sorting values. Generally, the sand layers are very well to well sorted (0.3–0.4), whereas the river marsh sediments range from 0.4 to 0.6 and are well to moderately well sorted (Table S2). (2) No distinction could be made by kurtosis, as all samples yield very leptokurtic grain-size distributions (Fig. S2, Table S2). The sand layers range from 3.2 to 5.1, with sand layer 4 reaching a kurtosis value as high as 11.1 (Table S2). (3) In terms of skewness values, it can be noted that the grain-size distributions of nearly all samples are positively to strongly positively skewed (Fig. S2, Table S2). Exceptions are sand layer 1, with a skewness value of –0.1, and the fluvial reference samples, which exhibit skewness values ranging from 0.1 to 0.3 (Table S2). The other sand layers yield skewness values from 0.4 (sand layers 2 and 3) to 1.7 (sand layer 4). In comparison, the river marsh sediments in the first 50 cm below ground level yield skewness values from 0.4 to 0.8, those in the middle part of the profile (50–85 cm) yield skewness values from 1.1 to 1.3, and those in the bottom part of the profile (85–140 cm) yield skewness values from 1.4 to 1.9. The increasing skewness with increasing depth indicates a slight increase in the removal of the fine fraction of the river marsh sediments with increasing profile depth (Table S2).

Reference Samples

The grain-size distributions of reference samples from the adjacent intertidal, beach, and dune environments show mostly unimodal distributions (Fig. 3; Table S2 [footnote 1]). The intertidal and beach samples have means of 2.1e (Fig. 3; Table S2). The sample of the coastal dune yields a slightly larger mean of 2.3φ (Fig. 3; Table S2). The three fluvial reference samples are also unimodal, with means of 2.0φ for the proximal (to the coastline), 2.1φ for the intermediate, and 2.7φ for the more distal sampled river sediments (Fig. 3; Table S2). All river samples contain a fraction of grains >3φ, unlike the beach, marine intertidal, and dune samples (Fig. 3). In addition, a minor difference in sorting between the nonfluvial and fluvial reference samples was observed: All reference samples are very well sorted (0.2–0.3), with exception of the three fluvial samples, which range from 0.5 to 0.8 (Table S2).

Lateral Variation in the Floodplain Stratigraphy

Sand layer 1 in all riverbank profiles (TIR 1, TIR 9, TIR 16, and TIR 22) is relatively thin (3.5–8 cm) in comparison to the thicknesses documented in cores taken 30 m landward from the riverbank (TIR 4, TIR 12, TIR 18, and TIR 24; Fig. 4). The thickness of sand layer 1 increases up to 12–18.5 cm at 0–30 m distances landward from the riverbank, whereas the thickness continuously decreases from 30 m up to 120 m from the riverbank (TIR 6, TIR 11, TIR 21, and TIR 27; 0–6 cm; Figs. 4 and 5). In core TIR 11, this sand layer is not present or distinguishable from the surrounding river marsh sediments by any features. The same applies to cores taken at distances of 150 m (TIR 7) and 200 m (TIR 32) from the riverbank. As sand layer 1 generally tapers out between 120 and 150 m inland, its mean grain-size value decreases in the landward direction (Fig. 4). A narrow mean range of 2.1e–2.2e was documented for sand layer 1 in all four riverbank profiles (Fig. 4). However, the mean values of sand layer 1 increase to a range of 2.4e–2.9e in the cores taken at 120 m landward distance, indicating a clear landward-finining trend of this layer (Fig. 4). It is noteworthy that the southernmost transect (T1) shows the smallest mean in the core at 120 m (2.4φ), whereas the most northern transect of this river floodplain has the largest mean of 2.9φ (Fig. 4).
Figure 3. (A) Section of the reference riverbank profile (TIR 1) between depths of 70 and 90 cm (~20 cm). This section contains sand layers 4 and 5, which are not clearly distinguishable from the surrounding river marsh sediments in terms of grain-size distributions. Additionally, this section contains gray silt lenses (rip-up clasts) resembling the soil underneath this section. (B) SW beach side of Tirúa where the reference samples TIR R1 and TIR R2 were taken. (C) Sampling site of TIR R4 from the dune crest. (D) Rio Tirúa, east of the village of Tirúa. In total, three different river sediment samples from different parts of the river were taken for reference (TIR R3, TIR R5, and TIR R6).
The horizontal correlation of the other sand layers is more difficult due to the uneven microtopography of the floodplain and agricultural furrows from the thirteenth century (Ely et al., 2014). Some sand layers are thick at 30 m distance from the riverbank; others are not present at all at this distance but reoccur at a distance of 90 m, beyond which most of the sand layers have tapered out (Fig. 5). Along a cross section across the river floodplain, sand layers 2 and 3 can be traced to a distance of 60 m, with minor thickness variations before they taper out (Fig. 6). Sand layers 4 and 6 taper out at a distance of less than 30 m (Fig. 6). Subunits of sand layer 5 are present in all cores along both sections with decreasing grain sizes in the landward direction (2.7ϕ in TIR 6 and 2.4ϕ in TIR 19; Figs. 5 and 6). The only core in both sections that does not contain sand layer 5 is TIR 1 (in 120 m distance to the riverbank of the northeastern most transect of the core grid; Fig. 6). This specific core yields mean values of 2.9ϕ at the top, which gradually increase to 3.6ϕ at a depth of 80 cm (Table S2 [footnote 1]). From 80 cm to 190 cm, the mean values gradually decrease to 2.1ϕ, with one exception (Table S2). This exception can be found at depths from 175 to 177 cm, where an abrupt change of mean value to 2.5ϕ was documented (Table S2).

Implications of the Sedimentological and Grain-Size Analysis

Some sand layers exhibit different grain-size distributions in comparison to those of the surrounding river marsh sediments. In the case of sand layers 1, 2, 3, and 6, the distinction criteria are larger mean values, unimodal grain-size distributions, and the lack of a fine fraction (>3ϕ; Fig. 2). In comparison, the river marsh sediments yield less homogeneous grain-size distributions with a fraction of finer material >3ϕ (Fig. 2). These criteria are significantly distinctive to claim that the sand layers originated from different depositional processes than the river marsh sediments. The river marsh sediments result from fluviolacustrine transported sediments of dominantly sand- and silt-sized grains deriving from the Coastal Cordillera (Figs. 2 and 3). Since the Coastal Cordillera can be considered as primary source of the sand intercalations, the lack of the fraction >3ϕ implies that the material of the sand layers must have undergone a sorting process that did not affect the river marsh sediments (Friedman, 1962). This sorting process most likely happened in two consecutive steps: (1) The first sorting process of the sediment occurred during the transport of the sediments to the coast by waves, tides, coastal currents, and wind. This assumption is supported by the similarity of the grain-size distributions among all sand layers of the river floodplain and the reference samples of the intertidal area, beach, and dunes (Figs. 2 and 3). (2) In the second part of the sorting process, the sediment of the different coastal environments was eroded by high-energy waves that then inundated the onshore area, including the river floodplain. The inundation of the floodplain by high-energy waves must have occurred with larger basal shear stress, resulting in larger flow velocities, as indicated by the erosional bases of the sand intercalations. Erosion of the floodplain reached 1.2 km landward of the coastline. This explains the presence of slightly coarser well-sorted sand.
Figure 5. Cross section of the cores (first 2 m) along transect 1 (from riverbank to 120 m landward distance) on the river floodplain, correlating the sand layers (sl) intercalated in the river marsh sediments. The gray shading between the cores represents the correlation of these sand intercalations, whereas the purple line represents the boundary of unoxidized river sediments for orientation in increasing depths. It needs to be noted that due to the lack of geochronological data for the landward cores, not all sand layers can be correlated unambiguously. The dotted lines and light-gray shading represent additional potential correlations.
Stratigraphical, diagonal cross section of the Tirúa floodplain

Figure 6. Diagonal cross section of the cores (first 2 m) of the core grid (from riverbank to 120 m landward distance) on the river floodplain. The gray shading between the cores represents the correlation of the sand layer (sl) intercalations, whereas the purple line represents the boundary of unoxidized river sediments for orientation in increasing depths. It needs to be noted that due to the lack of geochronological data for the landward cores, not all sand layers can be correlated unambiguously. The dotted lines represent additional potential correlations.
intercalations of varying thickness, which taper out and decrease in grain size with increasing landward direction. Therefore, episodic river flooding due to increased discharge, e.g., during snowmelt, is not considered causative for the deposition of these sand intercalations. The presumably originally present finer fraction of the sand layers was either removed by sorting processes and/or removed and redeposited beyond the study area with the eroded finer material of the river marsh sediments by the decelerating high-energy flow inland.

Limitations of the grain-size analytical distinction criteria for potential tsunami deposits in the Tīrūa case become apparent in sand layers 4 and 5. The distinction from the river marsh sediments only by grain size is not clear in sand layers 4 and 5, as the river sediments at depths of 50–85 cm in the profile yield similar distributional patterns (Figs. 2 and 3). Two potential processes might be the reason for the similar appearance of those two sand layers and the river marsh sediments in that section of the profile:

(1) Bioturbation could have resulted in mixing of these two different sediment types and thereby obliterated distinctive former sedimentological signatures. However, in that case, the bioturbation would mix but not eliminate the dominant silt fraction of the river marsh sediments, presuming these had the same grain-size distributions as the river marsh sediments outlying the profile section at 50–85 cm. In profile TIR 1, sand layer 4 yields a thickness of 2 cm, and sand layer 5 is 13 cm thick, which means the abundant fraction of fine grains (>3 μm) was removed from an ~35-cm-thick section of river marsh sediments by bioturbation, which does not seem very likely. Also, essential bioturbation producers that could cause sufficient sediment mixing in this profile section would most likely have left behind elongated, vertical structures with clear rims in the range of several centimeters to decimeters (Bromley, 1996). There are structures present in that section indicating slight bioturbation, but not of a vertical nature and not big enough in size to create sufficient mixing of the different sediments. Therefore, it is more likely that the river marsh sediments in this specific section did not experience any significant alteration of grain-size distribution after deposition.

(2) The similarity of the two different sediment types is rather an indication of a change in the sedimentological environment by, e.g., change in river discharge and river flow velocities, position of the riverbed, and/or distance to shoreline, which are all very likely to have been triggered by neotectonic subsidence or uplift of that area (allocyclic) or river avulsion flooding (auto-cyclic). This assumption is supported by the normal grading trend in core TIR 11 to a depth of 80 cm, which was replaced by an inverse grading trend of the marsh sediments throughout the lower part of this core. The change in sedimentation occurs in a gradual manner and starts at similar depths, as sand layers 4 and 5 can be found in more proximal cores. The intercalated sand layer at depths of 175–177 cm with a mean value of 2.5 μm, surrounded by coarser sediments (mean value of 2.5 μm), could therefore be an artifact of sand layer 6 (Fig. 6).

Diatom Analysis

In total, 180 taxa were identified, of which 83 taxa were identified to species level, representing 57 genera in the liner core TIR 2 (positioned next to TIR 1 at the riverbank). The remaining taxa were identified to genus level. The most common identified diatoms of this core were: Denticula sp. and Pinnularia sp., Planolithium frequentissimum, Planolithium minutissimum (Kraske) Lange-Bertalot, Falcacera meridionalis Metzeltin, Lange-Bertalot, and GarcíaRodríguez, Cosmoneis pusilla (W. Smith), D.G. Mann, and A.J. Stickle, and Eunotia subarcuataides Alles, Nörpel, and Lange-Bertalot. The samples of core TIR 2 showed variations in the number of valves (1–188 million/g). The sand layers yielded lower amounts of valves in comparison to the river marsh sediments, since these layers contained a higher proportion of fragmented valves. Marine and brackish-marine species such as Amphora cf. exililitata Giffen, Amphora cf. helenensis Giffen, Cocconeis costata Gregory, Dimeregramma minor var. minor, Diploneis cf. smithii, Paralia sulcata, Plagiogramma sp., Planolithium delicatum (Kützing) Round and Bukthiyarova, Pseudofallaxia cf. tenera, and Seminavis robusta Danielidis and Mann were commonly present in the sand layers. Also marine plankton (e.g., Skeletonema sp.) were present in the sand layers, with abundances up to 9%, while the majority of diatom species in the sand layers (77%) were benthic.

The diatom species present in the sand layers and river marsh sediments were subdivided in accordance to their salinity preferences into freshwater, brackish, and marine assemblages for a quantitative analysis. Two different kinds of changing patterns in the different assemblages responding to different salinity tolerances were observed along the 2 m profile (Fig. 7). The first assemblage change occurred in the uppermost part of the profile (0–50 cm), involving the first three sand layers. It appears that the ratio of the three assemblage groups changes abruptly in the sand layers in comparison to the surrounding river marsh sediments (Fig. 7). A sudden increase in the amount of marine and/or brackish species (>30%) along with a simultaneous decrease in the amount of freshwater diatoms occurs within these layers (<20%; Fig. 7). The abundance patterns of the assemblages within the first three sand layers slightly vary from each other (Fig. 7). However, these abundance patterns are very clearly distinguishable from diatom assemblages of the surrounding river marsh sediments (Fig. 7). In the remaining depths of the profile, changes of the diatom assemblages in accordance with their salinity preferences also can be observed within the remaining sand layers, e.g., an increase in the amount of marine diatoms in sand layers 4, 5, and 6 by over 20% (Fig. 7). However, considering the general ratio of these three different diatom halobous groups in the river marsh and in the three lowermost sand layers, it can be observed that these changes do not occur abruptly (Fig. 7). It appears that with increasing depths from 100 cm to 160 cm, the amount of marine (20% to 40%) and brackish (40–60%) diatoms present in the sediments increases gradually (Fig. 7). A slight retrograde development from this trend occurs with the marine diatoms exceeding a depth of 160 cm, whereas the abundance of brackish species continues to increase (>80%; Fig. 7). Simultaneously, the abundance of freshwater diatoms gradually decreases at depths of 110 to 200 cm from 60% to <20% (Fig. 7).
Figure 7. Abundances of the three different diatom salinity assemblages (freshwater, brackish water, and marine diatoms) plotted along the core (2 m) of the reference profile at the riverbank of Tirúa. The gray bars represent the sand layer (sl) intercalations, whereas the dotted bars cannot be clearly assigned to marine inundations of short duration. The deeper the profile, the more gradual is the marine influence at the river floodplain (indicated by the dotted arrows).
Implications of the Diatom Analysis

The distinct abundance patterns of mixed diatom habitat assemblages generally have been attributed to the influence of different sedimentological environments on the river floodplain of Tirúa (Dawson et al., 1996). The dominant presence of marine and brackish assemblages in the sand layers indicates a redeposition of marine sediment after marine high-energy transport on land (Kokociński et al., 2009). This indication is supported by the low number of valves and increased number of fragmented valves in these layers (Horton et al., 2011). The majority of marine diatoms in the sand layers are benthic forms, implying the sediment was redeposited over a short distance from shallow marine water to the river floodplain.

The first 50 cm section of the profile seems to reflect three short-term marine influences by an increase in marine and brackish diatom species and a simultaneous decrease of freshwater diatom species within the first three sand layers. However, with increasing depth, the changes in diatom assemblages become more gradual, indicating that the first three sand layers resulted from short-duration marine inundation events. In the upper part of the profile (0–100 cm), the sand layers can be distinguished from the river marsh sediments by the abundances of the different diatom assemblages, whereas in the lower part of the profile (100–200 cm), the different sediments are not distinguishable from each other by the diatom salinity assemblages (Fig. 7). This downward development of a more pronounced marine influence implied by a similar increase of marine and brackish diatom species in both types of sediments possibly reflects changes in the sedimentological environment induced by neotectonic processes rather than avulsion flooding (Fig. 7).

Geochemical and Mineralogical Analysis

General Trends

The geochemical and mineralogical compositions of marsh and event deposits show distinct patterns and systematic variations: The sand layers in core TIR 2 from the riverbank show depletion in K, P, and Rb (Fig. 8; Table 1). There is also a significant depletion of Al within sand layers 1, 2, and 4 (Fig. 8; Table 1). In addition, the LOI of the sand layers is smaller in comparison to the river marsh sediments (Fig. 8; Table 1). It can also be noted that Si, Ca, and Sr show increased concentrations in comparison to the river marsh sediments (Fig. 8; Table 1). Sand layers 1 and 2 are enriched in the elements Ti, Fe, Mn, and Mg compared to the marsh sediments (Fig. 8; Table 1). It seems that the concentrations of these latter elements approximate the concentrations of Ti, Fe, Mn, and Mg of the marsh sediments with increasing depth (from 50 cm under ground level; Fig. 8). However, small peaks of these elements occur at depths of 70 cm, 90 cm, 109 cm, and 180 cm (Fig. 8). These depths correlate with the lowermost three sand layers of the profile (Fig. 8).

Some element abundances vary within sand layer 1, the most recent sand layer. The lower part of this sand layer shows a slight decrease in Si (from 53% to 51%) and a slight increase in Ti (from 1.9% to 2.1%) and Fe (from 15.2% to 16.5%; Fig. 8; Table 1). In other sand layers that could be sampled more than once, these trends do not occur (Fig. 8). Similar depletion and enrichment patterns occur in core TIR 6, which was taken from 120 m distance from the riverbank, as well as core TIR 2. These patterns in the more distal cores also occur at depths of 5 cm, 20–25 cm, 100 cm, and 105 cm (Fig. S3 [footnote 1]).

Nearly all samples contain quartz (8%–18%), plagioclase (10%–16%) and muscovite (0%–17%) and rock fragments of magmatic (15%–51%) and metamorphic (0%–5%) origin as framework components (Fig. S4 [footnote 1]). All samples also contain abundant heavy minerals (clinopyroxene, orthopyroxene, olivine and minor amounts of biotite, epidote, hornblende and limonite, rutile, fluorite, and opaque minerals; Fig. S4). The bulk of heavy minerals amount to 13.0%–40.3% in the reference samples, 25.6%–43.6% within the sand intercalations, and 10.0%–36.6% within the river marsh samples (Table 1; Fig. S4). Within the river marsh samples, the most abundant grains present are of a finer fraction (18.3%–54.3%) too small for a mineralogical determination with the polarization microscope (Fig. S4).

Element Ratios

The more different tsunami deposits were analyzed geochemically over time, the more it became possible to link different element distributions of these deposits to different sedimentological and geochemical facies (e.g., Chagué-Goff and Goff, 1999; Dominey-Howes et al., 2006; Nichol et al., 2007). In order to understand the origin of these specific elemental enrichment and depletion patterns, the source dependency of the potential tsunami deposits has to be taken into consideration (e.g., Dawson et al., 1996; Font et al., 2013).

Comparing the abundances of the most abundant element Si (SiO₂) with Al, an element commonly used for correcting for grain-size effects (Al₂O₃; e.g., Ackermann, 1980; Van Daele et al., 2014), all sand intercalations plot with high SiO₂ (48%–62%) and low Al₂O₃ (11%–16%) concentrations (Fig. 9). The river marsh sediments, with no regard to the distance to the riverbank, yield more scattered concentrations, typically exceeding Al₂O₃ concentrations >13% (Fig. 9). More distal marsh sediments at 200 m from the riverbank yield similar Si and Al concentrations as the sand intercalation samples from the proximal (0 m) and intermediate distances (120 m; Fig. 9). The beach sediment reference sample exceeds the Al₂O₃ concentration of all sand intercalations, with a value of 16.7% (Fig. 9).

Enrichments of the elements Ca (plotted as CaO) and Sr in sand intercalations in marsh successions are commonly considered proxies of marine inundation events, reflecting an increased abundance of shell fragments in a deposit (e.g., Chagué-Goff et al., 2012). However, increased abundances may also be due to an increased presence of silicate minerals rich in Ca and Sr in a deposit.
Sr and CaO show a positive linear correlation in the sand layers of the proximal (0 m) and distal (120 m) cores (Fig. 9). The samples of the sand layers represent the upper end member of this trend, with increased values in CaO (>4%–7%) and Sr (>230–360 ppm; Fig. 9). The majority of the river marsh sediments have lower concentrations of Ca (<4%) and Sr (<230 ppm).

Some river marsh samples yield higher concentrations in both elements within all three cores (Fig. 9), even though they do not contain any shell fragments. The beach sample has the highest concentrations of Ca (~7%) and Sr (410 ppm). The TIC measurements of the samples of the riverbank profile show that there is no carbonate within the sand intercalations or the river marsh samples.

We therefore conclude that the increased concentrations of Ca and Sr are due to the abundance of pyroxenes and anorthite-rich plagioclase in the sediments. Beach deposits of central Chile are commonly rich in heavy minerals, with elevated abundances of pyroxenes, olivine, and hornblende (e.g., Fiske, 2015). Sources of these minerals are widely distributed in the late Paleozoic metamorphic and igneous rocks constituting a large part of the Coastal Cordillera of central Chile, and in the volcanic Andean Cordillera (Hervé et al., 2007; Parada et al., 2007).

Ti is considered a proxy either for organic-rich fine-grained terrestrial sediments that have not originated from tsunamis or for an enrichment of heavy minerals, and rutile, titanite, and ilmenite in particular, in a deposit (Chagué-Goff et al., 2012; Ramírez-Herrera et al., 2012). The sand layers yield increased Ti concentrations (Fig. 9). Normalized to the LOI and plotted against the depth of the samples, the TiO₂/LOI ratios of the sand intercalations and the beach sand are elevated, ranging from 0.2 to 1.6, whereas the river marsh sediments yield lower ratios (<0.3; Fig. 9). Similar to the increased CaO and Sr concentrations, the distribution pattern of the TiO₂/LOI ratios here also suggests an independence of the Ti concentrations from the LOI. The LOI values are low and appear to be associated with minor amounts of terrestrial organic matter, as carbonate is absent from the deposit.

The high Ti concentrations and low LOI values within the sand intercalations cause the high ratios within the potential tsunami layers (Fig. 9).
The primary composition of terrestrial sediments. The enrichment of elements within dune systems or redeposition by high-energy tsunami flow will alter secondary processes (e.g., sorting of grain sizes by waves or eolian transport). Despite the similar primary provenance of all sediments at Tirúa, elements are often related to signatures indicative of terrestrial input (e.g., Nentwig et al., 2012). These elements are enriched or depleted in comparison to surrounding sediments, and they reflect abrupt changes in uniform sedimentation patterns, in this case, episodic marine flooding of the Tirúa estuary and river floodplain (e.g., Chagué-Goff et al., 2012). The studied sand layers show depletion in Al, K, and P, in combination with decreased LOI values, in comparison to the river marsh sediments (Figs. 8 and 9). The heavy mineral accumulations are correlated with sand layers (sl) and river marsh (rm) sample concentrations exceeding the maximum measured concentration in the river marsh are coloured dark red and concentrations exceeding the mean concentration of the river marsh are coloured light red. Concentrations in the sand layers falling below the minimal measured concentration in the river marsh are coloured dark blue and concentrations falling below the mean concentration of the river marsh are coloured light blue.

### Implications of the Geochemical and Mineralogical Analysis

Certain elements and components within high-energy coastal sediments are enriched or depleted in comparison to surrounding sediments, and they represent abrupt changes in uniform sedimentation patterns, in this case, episodic marine flooding of the Tirúa estuary and river floodplain (e.g., Chagué-Goff et al., 2012). The studied sand layers show depletion in Al, K, and P, in combination with decreased LOI values, in comparison to the river marsh sediments (Fig. 8). These depletions are consistent with geochemical analyses of other tsunami deposits (e.g., Chagué-Goff et al., 2012). Therefore, the geochemical signature of enrichment and depletion of the different elements, reflecting higher amounts of heavy minerals, can be used in the Tirúa case to help distinguish the sand layers from the river marsh sediments. It would also explain the covariance of the enriched elements, e.g., CaO and Sr (Figs. 8 and 9).

The river marsh sediments also exhibit a spectrum of heavy minerals similar to those of the sand layers (Fig. S4). However, due to the larger amounts of heavy minerals (e.g., Ca-rich pyroxenes, olivine, Ti-oxides, etc.) in these sand layers (Figs. 8 and 9; Fig. S3). The river marsh sediments also exhibit a spectrum of heavy minerals similar to those of the sand layers (Fig. S4). However, due to the larger amounts of heavy minerals (e.g., pyroxene and olivine) and rock fragments with heavy minerals are less abundant in the marsh sediments (Fig. S4). Therefore, the geochemical signatures of enrichment and depletion of the different elements, reflecting higher amounts of heavy minerals, can be used in the Tirúa case to help distinguish the sand layers from the river marsh sediments. It would also explain the covariance of the enriched elements, e.g., CaO and Sr (Figs. 8 and 9).

At depths of 70–110 cm in the proximal core, the elemental pattern occurs four times (Fig. 8). The heavy mineral accumulations are correlated with sand layers 4 (at 70 cm depth) and three units of sand layer 5 (at depths of 90–

### TABLE 1. CONCENTRATIONS OF SELECTED ELEMENTS AND HEAVY MINERALS PRESENT IN THE SAND LAYERS AND RIVER MARSH SAMPLES

<table>
<thead>
<tr>
<th>Deposits</th>
<th>LOI (%)</th>
<th>Al (%)</th>
<th>K (%)</th>
<th>P (%)</th>
<th>Rb (ppm)</th>
<th>Si (%)</th>
<th>Ca (%)</th>
<th>Sr (ppm)</th>
<th>Ti (%)</th>
<th>Fe (%)</th>
<th>Mg (%)</th>
<th>Heavy minerals (%)</th>
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<tr>
<td>sl 1 (top)</td>
<td>2.5</td>
<td>11.1</td>
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<td>0.2</td>
<td>40</td>
<td>52.9</td>
<td>5.1</td>
<td>246</td>
<td>1.9</td>
<td>15.2</td>
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<td>0.2</td>
<td>37</td>
<td>50.6</td>
<td>5.2</td>
<td>250</td>
<td>2.1</td>
<td>16.5</td>
<td>7.5</td>
<td>77.3</td>
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<td>1.2</td>
<td>0.2</td>
<td>51</td>
<td>53.4</td>
<td>4.7</td>
<td>256</td>
<td>1.9</td>
<td>13.5</td>
<td>4.4</td>
<td>64.0</td>
</tr>
<tr>
<td>sl 2 (bottom)</td>
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<td>1.2</td>
<td>0.2</td>
<td>56</td>
<td>48.0</td>
<td>4.4</td>
<td>235</td>
<td>1.9</td>
<td>15.0</td>
<td>4.9</td>
<td>64.0</td>
</tr>
<tr>
<td>sl 3 (top)</td>
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<td>14.4</td>
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<td>0.2</td>
<td>41</td>
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<td>8.1</td>
<td>318</td>
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<tr>
<td>sl 3 (bottom)</td>
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<td>5.5</td>
<td>306</td>
<td>1.6</td>
<td>11.6</td>
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<tr>
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<td>61.5</td>
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<td>296</td>
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<tr>
<td>sl 6 (bottom)</td>
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<td>14.3</td>
<td>1.3</td>
<td>0.3</td>
<td>49</td>
<td>61.2</td>
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<td>1.3</td>
<td>8.6</td>
<td>2.8</td>
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**Note:** Concentrations of selected elements and heavy minerals (pyroxene, olivine, biotite, limonite, epidote, rutile, hornblende, fluorite, and igneous and metamorphic rock fragments present in high-energy mineral fraction after density separation) present in the sand layers (sl) and river marsh (rm) samples. Fifteen river marsh samples were analysed with energy-dispersive X-ray fluorescence (EDXRF) and 9 river marsh samples were used for the mineralogical analysis. The element and heavy mineral concentrations yield depletion (blue) and enrichment (red) patterns of the sand layers in reference to the concentrations of the river marsh sediments (grey). Element and heavy mineral concentrations in the sand layers exceeding the minimum measured concentration in the river marsh are coloured dark red and concentrations exceeding the mean concentration of the river marsh are coloured light red. Concentrations in the sand layers falling below the minimal measured concentration in the river marsh are coloured dark blue and concentrations falling below the mean concentration of the river marsh are coloured light blue.
The presence of three heavy mineral–enriched units within this sand layer could indicate multiple incidents of deposition during flooding of a single tsunami. Additionally, the concentrations of Ti and Fe are slightly increased within the basal part of sand layer 1 (Fig. 8). Therefore, the youngest sand layer indicates suspension load as the dominant transport load, allowing denser minerals to be deposited first in sand layer 1, even if this cannot be deduced from the grain-size data. Two of the three heavy mineral subunits are observed in core TIR 6 (at 120 m landward distance) and are similar to the heavy mineral accumulation of sand layer 1 and presumably of sand layer 2 (Fig. S3). This implies that the other sand layers have tapered out at distances less than 120 m from the riverbank.

Comparison of sand layers and reference samples shows that the increased Al concentration of the beach sample in contrast to the sand layer and river marsh samples exemplifies secondary depositional processes (Fig. 9). The intertidal and beach samples contain more rock fragments and less single-grained heavy minerals than all the other samples (Fig. S4). The bulk of these rock fragments consists of volcanic (andesitic origin) and plutonic (granitoid) components (Fig. S4). The slightly increased Al concentration...
could therefore relate to a higher amount of anorthite-rich plagioclase present in these rock fragments. The beach sample also yields the lowest variety of heavy minerals (Fig. S4).

The primary source of all sediments at Tirúa is most likely the Coastal Cordillera, which consists of micaceous metapelites, metabasites, especially green schists, gneiss, and metamorphosed ultramafites in the area dissected by the Río Tirúa (Mapa Geológico de Chile, 2003; Hervé et al., 2007). These source rocks generally yield the heavy mineral associations present in the sand layers. The dominant presence of volcanic rock fragments next to the mafic heavy minerals suggests also that the Andean Cordillera, with its active inner-arc volcanism, contributes to the provenance of central Chilean coastal sedimentation (Parada et al., 2007).

**DISCUSSION**

**Storm versus Tsunami in Central Chile**

The sand layers described here appear to have been deposited by marine inundation events. Sand layer 1 clearly originated from the tsunami of the 2010 Maule earthquake in central Chile, as the landward fining and thinning of this layer, its erosional base, and also the landward-bent vegetation shortly after the tsunami correspond with eyewitness observations during the posttsunami survey conducted in March 2010. Sand layer 5, the thickest sand layer, also shows a landward decrease in grain size along the length of the river floodplain and also contains rip-up clasts. These sedimentological characteristics tend to point to a tsunami rather than a storm surge origin (Morton et al., 2007). Additionally, historical records of extreme marine inundation events show that no cyclone or inundation by strong storm surges entering rivers to such great distance inland have been documented in central Chile. This part of Chile is positioned in moderate climatic latitudes with cold water upwelling contributing to the rare occurrences of coastal storms (Das Gupta et al., 2009). The central Chilean coastline has been very prone to seismicogenic triggered tsunamis of different scales. The historical record of documented extreme flood events describes numerous tsunamis having struck the coastline over the past 500 yr (Lockridge, 1985; Lomnitz, 2004). The OSL age ranges obtained by Nentwig et al. (2015) for this Tirúa profile (TIR 1) coincide with at least three tsunami events in the historical record. The oldest three tsunami deposits exceed 500 yr in age.

**Effectiveness of Different Methods at Tirúa**

The studies at Tirúa show the importance of applying a combination of different methods to distinguish tsunami deposits from other deposits, since not every method reveals necessary distinctive characteristics (e.g., Goff et al., 2010; Chagué-Goff et al., 2011; Ramirez-Herrera et al., 2012). This difficult matter of methodological differentiation of tsunami deposits from nontsunami deposits was pointed out before by Bellanova et al. (2016) during the microtextural analysis of quartz grains from the Tirúa floodplain and its intercalated sand layers. In this study, large numbers of separated quartz grains from samples of the four topmost sand layers, the surrounding river marsh sediments, and reference samples from the beach, dunes, and river were analyzed for distinctive microtextural families, such as angularity, fresh surfaces, percussion marks, adhering particles, and dissolution features, under a scanning electron microscope (SEM). This analysis showed similar microtextures on the grain surfaces in all different samples. The combination of the fluvial influence in most of these adjacent sedimentary environments, the potential short residence time of grains in these environments, the high amounts of heavy minerals in the sand layers, and the mixing of the different sediments during tsunami inundations did not allow any differentiation of tsunami deposits by this method.

Sedimentological and grain-size analyses can be used to distinguish the majority of tsunami sands from the river marsh sands of the Tirúa River floodplain. However, the studies show limitations, since the intercalated tsunami deposits cannot be distinguished easily from the adjacent river marsh deposits. The coarsening of the river marsh sediments in lower part of the first meter of the riverbank profiles (50–85 cm in TIR 1 and 70–90 cm in core TIR 2) and the gradual inverse grading trend in core TIR 11 (onset in similar depths) indicate a fluvial hydrodynamic environment with increased flow velocities in the Rio Tirúa and greater transport capacity of larger grains to the river floodplain. The diatom analysis provided useful indication of short-term marine inundations for the upper part of the profile, helping to identify three uppermost tsunami sand layers. At greater depths, the three lowermost tsunami deposits could not be identified due to postdepositional changes. These environmental changes are also associated with vertical neotectonic movements, which are frequently experienced co- and postseismically along various parts of the central Chilean coastline, including at Tirúa (Ely et al., 2014).

Ely et al. (2014) were able to reconstruct neotectonic movements resulting from four high-magnitude earthquakes along the Tirúa profile using increased amounts of marine diatom species in the four uppermost sand layers of the Tirúa profile. Additionally, Ely et al. (2014) reconstructed changes in the relative sea level with tide gauge measurements and a tide prediction model, indicating that the area of Tirúa experienced coseismic uplift of 0.48 ± 0.1 m in 2010. Other studies support these results based on the analysis of dried coralline algae (>0.5 m uplift by Farías et al., 2010) and exposed intertidal mussels (0.91 ± 0.17 m uplift by Melnick et al., 2012). Coseismic subsidence was documented during the earthquake in 1960 (–0.2 ± 0.4 m by Plafker and Savage, 1970; 0.3 ± 0.72 m by Garrett et al., 2013).

In contrast with the grain-size analysis and diatom analysis, the geochemical analyses show consistent differences in the major- and minor-element compositions throughout the whole profile. The geochemical data show heavy mineral accumulations in all tsunami deposits and enabled the identification of sand layer 4 and three different subunits of sand layer 5 as tsunami deposits. This is of special importance because sand layer 5 is thickest layer in
the Tirúa profile and appears to have been created by at least three different tsunami waves. The luminescence dating gives insight regarding the origin of this very specific sand layer. An age inversion was found within this layer by Nentwig et al. (2015). This age inversion was most likely caused by a two-step sediment entainment: (1) The young and well-bleached sediments of beach and dune surfaces were entrained and deposited on the river floodplain by the first tsunami wave, and (2) the older and badly or nonbleached sediments of the intertidal and lower parts of the beach and dunes were entrained by successive waves of the tsunami (Nentwig et al., 2015). This kind of luminescence age inversion has been observed in historical tsunami deposits on the Scilly Isles, UK, related to the 1755 Lisbon tsunami, dated by Banerjee et al. (2001), and in recent tsunami deposits related to the Indian Ocean tsunami of 2004 in Thailand, dated by Sanderson and Murphy (2010).

We conclude here that certain methods commonly applied to distinguish tsunami deposits from other sediments (e.g., grain-size analysis and diatom analysis in the case of Tirúa) are particularly useful for younger profiles and/or localities that have not experienced too much change in their arrangement of sedimentary and ecological facies. However, in environments with regular shifts of facies by neotectonic or climatic processes, such as at Tirúa, these methods cannot be used alone to identify older tsunami deposits buried at depth.

Reconstruction of the Geological Tsunami History of Tirúa

The tsunami chronology for the Tirúa riverbank profile is constrained by the OSL ages from Nentwig et al. (2015). The age ranges of the tsunami sand layers for this profile are: Sand layer 1 was deposited in 2010, sand layer 2 relates to the 1960 Chile tsunami with an age range of A.D. 1940 ± 10 yr, and sand layer 3 yielded an age of A.D. 1730 ± 100 and therefore could have resulted from the tsunami related to either the 1730 Valparaíso earthquake or, as Ely et al. (2014) suggested, the 1751 Concepción earthquake (Lomnitz, 1970). There is no evidence so far to determine which of those two different tsunami events was responsible for the deposition of sand layer 3. Further down the profile, the OSL ages of the tsunami deposits exceed the historical tsunami record for this area (Nentwig et al., 2015). Therefore, the age data for this study were complemented with independent age control from dating results of other tsunami deposit studies involving infrared stimulated luminescence ages and radiocarbon ages in central Chile, such as Cisternas et al. (2005) and Messens (2014). From these studies, it is suggested that sand layer 4 of the Tirúa profile equates to a tsunami event between A.D. 1300 and 1500, sand layer 5 equates to a tsunami event along the Chilean coast between A.D. 850 and 1250, and the oldest tsunami layer (sand layer 6) of this profile was presumably deposited between A.D. 400 and 600.

The geological tsunami recurrence interval of the Tirúa River floodplain (200–300 yr) differs from historical records, which give recurrence intervals <100 yr (Lomnitz, 1970; Lockridge, 1985). This difference in the geological and historical tsunami records is not surprising, since not all tsunamis documented by witnesses leave behind traces that are incorporated or preserved in the geological record (Bahlburg and Spiske, 2015; Szczuciński, 2012; Spiske et al., 2013).

CONCLUSIONS

The 2-m-thick floodplain profile at Tirúa contains six layers of sandy tsunami deposits representing six different tsunami events. Distinguishing these deposits from the floodplain sediments by grain-size analysis alone was not possible, since sand layers 4 and 5 were deposited within floodplain sediments with grain-size distributions similar to the six tsunami sand layers. In this study, diatom analysis could only be applied to the upper three tsunami sand layers, since the lower profile has been subject to gradual environmental changes caused by vertical neotectonic movement during and after earthquakes. The geochemical and mineralogical analyses revealed characteristic accumulations of heavy minerals within the six tsunami sand layers. Multiple heavy mineral layers and OSL age inversion in sand layer 5 could be related to multiple waves during that prehistoric tsunami event. This emphasizes the importance of combining methods that have different degrees of sensitivity to the short-term environmental changes associated with neotectonic movements as experienced along the highly seismic coast of central Chile. This interdisciplinary study of recent and paleotsunami deposits at Tirúa demonstrates the necessity of combining different proxies when reconstructing the tsunami record of a locality with a complex tsunami history associated with vertical neotectonic influence. The tsunami history of central Chile known from historical records is, however, not fully preserved in the Tirúa profile. Most probably, this is likely due to the low preservation potential of older tsunami deposits. However, this study extends the historical tsunami record by deciphering the geological tsunami record over the past 1000 yr at Tirúa.

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