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Back to the bases: Building a terrestrial water $\delta^{18}O$ baseline for archaeological studies in North Patagonia (Argentina)

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ABSTRACT

Archaeology has been using stable oxygen as an isotopic tracer linked with water consumption for decades, and it has been demonstrated to be a powerful tool to assess paleomobility in bioarchaeology. Central-eastern North Patagonia (Argentina) is an especially appropriate region to apply it since it presents a high density of hunter-gatherer burials, it was a nodal zone criss-crossed by an extensive network of important routes, and it is characterized by a high environmental fragmentation due to the scarcity of fresh water sources. The aim of this paper is to build an empirical stable oxygen isotope baseline of terrestrial surface waters to assess the potentiality of tracing past human movement. We analyzed 46 water samples from 13 locations with permanent sources (rivers, springs, streams), compared it with predictions of precipitation and evaluated it considering seasonal variation, altitude and distance from the coast. Our results show that different post-precipitation processes change the isotopic signal from the sources with respect to the local precipitation, and highlight the relevance of analyzing terrestrial water sources. According to their oxygen isotope values we defined five hydrologic zones: Colorado River, Negro River, Closed Basins and Plains, Eastern and Western Somuncurá Foothills. Their identification shows the potential to address past human movement using stable oxygen water baselines in central-eastern North Patagonia.

1. Introduction

The isotopic composition of oxygen in mammalian biogenic tissues is linked to that of the body water pool (Longinelli, 1984; Luz et al., 1984), which is strongly influenced by the intake of drinking water (Bryant and Froelich, 1995; Podlesak et al., 2008). Luz and Kolodny (1989) pointed out the potential of using stable oxygen isotopes ($\delta^{18}O$) to trace past human movement and, a few years later, Schwarcz et al. (1991) performed the first archaeological application. From that time onward, $\delta^{18}O$ values have been increasingly used to assess paleomobility in archaeology (e.g. White et al., 1998; Dupras and Schwarcz, 2001; Knudson et al., 2009; Buzon et al., 2011; Gil et al., 2011; Ugan et al., 2012; Pellegrini et al., 2016; Barberena et al., 2017; Gregoricka et al., 2017). Nevertheless, there is no agreement about which kind baseline is the most suitable to assess mobility patterns (e.g. raw data distributions or transformed data into drinking water values).

The aim of this paper is to build a stable oxygen isotope baseline of terrestrial surface waters to assess the potential of tracing past human movement in central-eastern North Patagonia (Fig. 1). We consider this area, only occupied by hunter-gatherers, to be especially appropriate to tackle paleomobility studies using this approach for three main reasons. First, it presents a high density of human burials (Mariano, 2011; Prates and Di Prado, 2013; Serna, 2018), and there are no archaeological evidence associated to systematic water management (e.g. wells and irrigation systems) (Prates, 2008; Prates and Mange, 2016; Mange, 2019); second, it was a nodal zone criss-crossed by an extensive network of important indigenous routes; and third, it is characterized by the low availability and heterogeneous distribution of fresh water sources (Deodat, 1958–1959; Casamiquela, 1985; Musters, [1869–1870]1997). Due to this environmental setting and archaeological background, hunter-gatherers should have relied on the most important fresh water sources available in the region (Colorado River, Negro River and Valcheta Stream) (Fig. 1).

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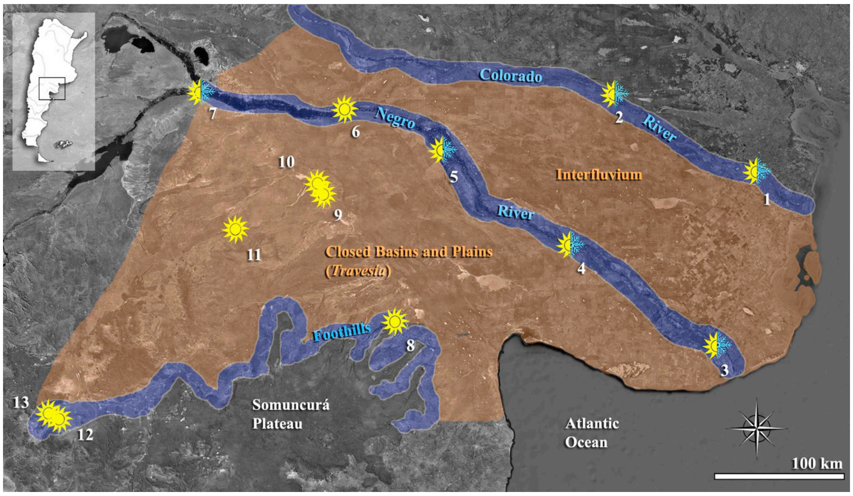

Fig. 1. Study region showing the dry lands (orange) and the wet strips (blue) with the location of sampled water sources: 1-CR lower, 2-CR middle, 3-NR lower, 4-NR mid-low, 5-NR middle, 6-NR mid-up, 7-RN upper, 8-SFE Valcheta, 9-BP Cuiña, 10-BP Trapalcó 2, 11-CF Santa Victoria, 12-SFW Santana, 13-SFW Chacras. Color of symbols indicate season of sampling: yellow = summer; yellow/blue = summer/winter. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

1.1. Climatological and hydrological background

North Patagonia is predominately influenced by the Pacific anticyclone, additionally affected by air masses from the Atlantic Ocean in the eastern sectors, and by west winds -westerlies- (Paruelo et al., 1998). The Andes intercepts humid air masses from the Pacific Ocean generating "orographic precipitation" on the windward side and, as the air masses pass over the mountains to the east, the amount of moisture decreases (Stern and Blisniuk, 2002). As a result, rainfall ranges from 3000 mm/year on the summit of the Andes to 400 mm/year in the foothills and less than 200 mm/year to the east (Schäbitz, 2003). Unlike Central and South Patagonia where precipitations are concentrated in winter, in North Patagonia the seasonal regime is highly variable depending on the area (Paruelo et al., 1998). According to paleoclimate reconstructions, extra-Andean North Patagonia seems to have been dominated by arid/semi-arid conditions for the last *ca.* 3000 years BP (Schäbitz, 2003). Paleoclimate proxies do not suggest large dry events, such as those occurring in Andean northwestern Patagonia (Villalba, 1994; Boninsegna, 1995). These conditions generate both a permanent hydric deficit (Román and Sisul, 1984) and a severe environmental fragmentation, characterized by two large "dry lands", with quite few non-permanent water sources interrupted by three "wet stripes": Colorado River, Negro River and Somuncurá Plateau Foothills' stream system, from north to south (Fig. 1) (see Prates and Mange, 2016). The Colorado and Negro rivers are separated from each other by an interfluvial space of ca. 40000 km^2 (120 km apart in average from each other), and the latter is separated from the Somuncurá Foothills by a larger area -more than 70000 km^2-called the Closed Basins and Plains or "*travesía*", one of most unfavorable landscapes of Patagonia (Fig. 1) (Casamiquela, 1985; Prates and Mange, 2016).

The Colorado River rises from the confluence of the Grande and Barrancas rivers -both with feeding basins in the arid Andes of southern Mendoza and northern Neuquén-, and flows southeast towards the Atlantic (Fig. 1) (Spalletti and Isla, 2003). The Grande River, Colorado's main tributary, is almost exclusively fed by melt water (Cazenave, 2011), and the Barrancas is a mix between snow and snow-rain fed basins (Subsecretaría de Recursos Hídricos, 2004). Downstream of its headwaters, the Colorado is fed by two minor tributaries (Butacó Creek and Curacó system) (Spalletti and Isla, 2003; Cazenave, 2011). Its regime is permanent (average annual flow 150 m^3/sec) and almost exclusively based on melt water, with some irregular influxes of rainfall from the winter trans-Andean pluvial regime (Arenas and Turazzini, 1990) and or the exceptional spring-summer precipitations generated by the South Atlantic anticyclone (Burgos, 1974).

The Negro River, the most important watercourse in Patagonia (Soldano, 1947), flows in a meandering way eastward from the confluence of the Limay and Neuquén rivers, crossing northern Patagonia before draining into the Atlantic (Fig. 1) (González Díaz and Malagnino, 1984). The Limay River is its main tributary with an annual average contribution of ~700 m^3/sec -max. of ~1700 m^3/sec in June and a min. of ~300m^3/sec in April-, derived from rainwater. The Neuquén River annual flow is subject to the rainfall and snowmelt, and registers an average of ~300m^3/sec -with a max. of ~600 m^3/sec in November and a min. of ~100 m^3/sec in April- (Soldano, 1947). The Negro River is remarkable for its large flow, with an average of ca. 1000 m^3/sec (Quiroga, 1992: 23), and floods exceeding 4000 m^3/sec (Cuevas Acevedo, 1981: 70).

The third and southernmost wet stripe of the study area is the line connecting fresh water springs along the Somuncurá Foothills (Fig. 1). This Tertiary basaltic plateau (~1500-1000 m.a.s.l.) constitutes a large natural reservoir of water (Aguilera, 2005). Rain and melting snow quickly infiltrate the basaltic mantle, giving rise to springs at the foothills. These springs drain northward as small temporary watercourses, with the exception of the Valcheta Stream, which is the only permanent one (Fig. 1) (Fontana, 2001; Prates and Mange, 2016).

1.2. Stable oxygen isotopes as a tracer in archaeology

The usefulness of stable oxygen isotopes (δ^{18}O) to assess mammalian movement resides in the fact that δ^{18}O values of biogenic tissues are mainly correlated with those from the sources of imbibed water, which in turn vary geographically (Bryant and Froelich, 1995; Daux et al., 2008; Podlesak et al., 2008). Methodologically, the use of δ^{18}O values from archaeological samples as a tracer can be performed either by using raw oxygen carbonate/phosphate data ($\delta^{18}O_{c/p}$) or by transforming it into drinking water values ($\delta^{18}O_{dw}$). The "raw data approach" usually assumes a $\delta^{18}O_{c/p}$ "local signal", built on a certain number of individuals with normally distributed isotopic values, as a reference to identify non-local/migrants or outliers (Lightfoot and O'Connell, 2016; cf. Pellegrini et al., 2016). On this basis, the research problem is conceived as a "local *vs.* non-local/migrant" dichotomy. Since it is assumed that "locals" lived and consumed "local" resources within the same geographic space over their lifetime, so they are actually a reliable local representation, this approach becomes mainly (or only) applicable in the context of low-mobility populations with large cemeteries (Scharlotta, 2016). In contrast, the "drinking water approach" does not need a local signal built on several presumably local individuals, because it uses the raw data to calculate the drinking water values ($\delta^{18}O_{dw}$) and compare each one against a water baseline ($\delta^{18}O_w$) (e.g. Chenery et al., 2010; Buzon et al., 2011). In these cases, the use of the "local *vs.* non-local/migrant" dichotomy would be inappropriate (Scharlotta, 2016), as the concept of "local" is not easily applicable to

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groups with different ranges of residential mobility. Hence, this approach seems to be more suitable for non-sedentary/mobile groups (e.g. hunter-gatherers), because it allows to link humans with specific water sources in a more straightforward way. As for other biogeochemical tracers, the success of the drinking water approach mainly relies on its baseline.

There are different strategies to build a $\delta^{18}O_w$ baseline to associate isotopic bioarchaeological signals with geographic regions. One way is to use precipitation ($\delta^{18}O_{pp}$), since its isotopic composition is geographically structured and determined by some environmental variables or "Dansgaard's effects" (e.g. distance from the coast, altitude, latitude), which are due to temperature-dependent continuous fractionation during the phase changes in the hydrologic cycle (Yurtsever and Gat, 1981; McGuire and McDonnell, 2007). These effects can be understood as a measure of the average degree of rainout from a vapor mass, on its way from its origin to the site of precipitation (Rozanski et al., 1993). So, a progressive ^{18}O-depletion of rainwater is observed as the cloud moves far inland, as the result of the isotopic effects during consecutive rainouts. Further precipitation will be enriched compared to the remaining water vapor, but it will be ^{18}O-depleted in relation to the previous precipitation from the same vapor mass (Clark and Fritz, 1997). The analysis relies on different strategies: a) isotopic measurements of precipitation, usually from the IAEA/WMO Global Networks of Isotopes in Precipitation or GNIP stations (Knudson et al., 2012; Knipper et al., 2014; Guede et al., 2017); b) interpolation maps of modern precipitation such as the one built by Bowen and Revenaugh (2003) (Mitchell and Millard, 2009; Scheeres et al., 2014; Redfern et al., 2016); c) predicted values through temperature dependent equations (Hamre and Daux, 2016); or d) a combination of them (Keenleyside et al., 2011; Hamre and Daux, 2016). Nevertheless, post-precipitation processes (e.g. mixing or evaporation) and or water transport are capable of deviating the isotopic values of the terrestrial water sources in relation to the local rainfall in a given region (Kendall and Coplen, 2001; Darling, 2004). Therefore, knowing the isotopic values from the actual water sources -rivers, streams, springs, etc.- to build a terrestrial surface water ($\delta^{18}O_{tsw}$) baseline might be a more accurate strategy (Leach et al., 2009; Chenery et al., 2010; Buzon et al., 2011; Webb et al., 2011; Ugan et al., 2012; Groves et al., 2013; Toyne et al., 2014; among others).

2. Material and methods

2.1. Water samples and isotopic determination

The water sampling strategy was designed to survey the main natural sources that would have been available in the past (i.e. modern human-made wells were not sampled). We collected 46 surface samples from 13 sampling points with permanent sources (Fig. 1), and from different sources (rivers, streams and natural springs). In the case of the large rivers, we sampled them in summer and winter to estimate seasonal variation (Table 1).

In order to observe possible biases due to specific climatic anomalies or events (e.g. extraordinary storms), we compared the average monthly temperature and amount of precipitation during the sampling months with data from 2000 to 2014 using information provided by the *Departamento Provincial de Aguas de Río Negro* (2015). Sampling was done following Mazor (2003), using 15 ml polypropylene test tubes, filling them completely, and sealing them with a plastic film to avoid the exchange of gases. The samples were kept upside down in the dark at room temperature (~24 °C).

Samples were prepared and analyzed at the Department of Geological Sciences and the Department of Archaeology at University of Cape Town (South Africa). The preparation methods for hydrogen (δ^2H) and oxygen ($\delta^{18}O$) isotopic analysis were the Zinc Reduction (Coleman et al., 1982) and CO_2 Equilibration (Socki et al., 1992), respectively. We used two in-house standards, the CTMP2010 (Cape Town Millipore Water 2010) and RMW (Rocky Mountain Water), which were calibrated against V-SMOW (Clark and Fritz, 1997). Isotopic ratios were measured on a Thermo Finnigan MAT 252 IRMS and expressed in delta (δ) notation relative to V-SMOW in parts per mil (‰). The precision of analysis was calculated as 1‰ for the δ^2H, and 0.1‰ for $\delta^{18}O$.

2.2. Analytical procedures

In order to explore the similarity between the empirical and predicted data, we compared the δ^2H and $\delta^{18}O$ values from the terrestrial water samples against predictions of modern precipitation for the same coordinates annually and seasonally calculated using the Online Isotopes in Precipitation Calculator -OIPC (Bowen, 2018)- (Table A.1). To compare our data with the predictions, we used simple linear regressions (least squares method) and added the Global Meteoric Water Line (GMWL) as a global reference for descriptive purposes (Craig, 1961). The $\delta^{18}O$ distributions of rivers were observed to assess seasonal variation, tested with Shapiro-Wilk (W) for normality in small data sets and Student's t-test to evaluate differences between them (Zar, 1999). Then, we analyzed the relationship between our $\delta^{18}O$ data with altitude and distance from the coast with Pearson's correlation (r). All standard deviations -SD-are express as 1σ, and the statistical procedures were carried out with $\alpha = 0.05$ with SPSS 20.

3. Results

The detailed isotopic results for the terrestrial surface water are presented in Table 1. The overall range for the δ^2H spans from −99.6 to −62.7‰ with a mean standard deviation of 0.99‰, and the $\delta^{18}O$ varies from −12.48 to −7.65‰ with a mean standard deviation of 0.16‰.

3.1. Terrestrial surface waters vs. precipitation predictions

The overall results show that the water samples have lower isotope values than OIPC predictions. The comparison of the slopes shows that the annually predicted precipitation $\delta^{18}O$ values (7.78*$\delta^{18}O$) and water sample $\delta^{18}O$ values (7.46*$\delta^{18}O$) fit relatively well with the GMWL (8*$\delta^{18}O$), with the slope of the water sample moderately lower than the GMWL line. In contrast, the slope for seasonally predicted precipitation is quite lower than the others (4.78*$\delta^{18}O$), and the regression line is above the rest (Fig. 2).

3.2. Seasonal and river basin variation

In Fig. 3 the Colorado lower basin shows the most negative $\delta^{18}O$ values in winter ($\delta^{18}O_{summer} = -9.77 \pm 0.1$‰ and $\delta^{18}O_{winter} = -10.08 \pm 0.05$‰), and the middle basin inverts this trend exhibiting ^{18}O-depleted water in summer ($\delta^{18}O_{summer} = -11.06 \pm 0.07$‰ and $\delta^{18}O_{winter} = -10.45 \pm 0.0007$‰). The maximum $\delta^{18}O$ seasonal difference registered for the Colorado River is ~0.6‰. The results for the Negro River show that the three sectors are depleted in winter with a maximum $\delta^{18}O$ difference of ~0.6‰ (upper: $\delta^{18}O_{summer} = -8.35 \pm 0.08$‰ and $\delta^{18}O_{winter} = -8.96 \pm 0.14$‰; middle: $\delta^{18}O_{summer} = -8.34 \pm 0.04$‰ and $\delta^{18}O_{winter} = -8.77 \pm 0.12$‰; lower: $\delta^{18}O_{summer} = -8.06 \pm 0.4$‰ and $\delta^{18}O_{winter} = -8.65 \pm 0.36$‰) (Fig. 3).

Due to the small seasonal isotopic differences in both rivers (< 1‰, see section 4.1.), summer and winter values were averaged (Fig. 4). The plot shows that the Negro is more enriched than the Colorado ($\delta^{18}O = -8.53 \pm 0.36$‰ ranged from −8.36 to −8.86‰, Shapiro-Wilk test: W=0.939, p = 0.169; and $\delta^{18}O = -10.25 \pm 0.49$‰ ranged from −9.67 to −11.10‰, Shapiro-Wilk test: W=0.91, p = 0.283,respectively), and the difference is statistically significant ($t = -11.28, p < 0.01$).

Table 1
Environmental and isotopic data of the terrestrial surface water samples.

Map	Location	Source	Coordinate	Season*	Alt. (m.a.s.l.)	Dist. (km)	δ^2H	Mean δ^2H	SD δ^2H	$\delta^{18}O$	Mean $\delta^{18}O$	SD $\delta^{18}O$
1	CR lower	river	39°31′S; 62°39′W	summer	24	54	−82.99	−83.73	0.74	−9.67	−9.77	0.1
							−83.73			−9.76		
							−84.47			−9.86		
				winter	24	54	−81.99	−81.76	2.2	−10.05	−10.08	−0.05
							−83.82			−10.14		
							−79.45			−10.05		
2	CR middle	river	38°58′S; 64°06′W	summer	100	190	−83.73	−92.5	1.19	−11.1	−11.06	0.07
							−93.35			−10.98		
							−91.66			−11.1		
				winter	100	190	−86.09	−85.92	0.15	−10.45	−10.45	0
							−85.81			−12.22*		
							−85.87			−10.45		
3	NR lower	river	40°48′S; 62°59′W	summer	5	31	−64.43	−64.21	0.24	−8.44	−8.06	0.4
							−64.26			−8.08		
							−63.95			−7.65		
				winter	5	31	−67.8	−67.06	0.95	−8.39	−8.65	0.36
							−65.98			−9.06		
							−67.39			−8.51		
4	NR mid-low	river	40°6′S; 64°27′W	summer	62	178	−67.6	−67.91	0.51	−8.45	−8.39	0.08
							−67.62			−8.29		
							−68.5			−8.42		
				winter	62	178	−68.81	−69.09	0.64	−12.37**		
							−69.82			−11.34**		
							−68.63			−10.41**		
5	NR middle	river	39°25′S; 65°43′W	summer	136	310	−65.56	−64.66	1.68	−8.31	−8.34	0.04
							−65.71			−8.38		
							−62.72			−8.32		
				winter	136	310	−66.7	−67.94	1.08	−8.79	−8.77	0.12
							−68.73			−8.89		
							−68.39			−8.64		
6	NR mid-up	river	39°6′S; 66°38′W	summer	180	396	−68.83	−68.61	0.31	−8.51	−8.86	0.5
							−68.39			−9.21		
7	NR upper	river	38°58′S; 68°3′W	summer	260	506	−67.62	−66.39	1.47	−8.44	−8.35	0.08
							−66.79			−8.3		
							−64.77			−8.3		
				winter	260	506	−69	−67.31	1.73	−9.06	−8.96	0.14
							−67.39			−9.02		
							−65.53			−8.8		
8	SFE Valcheta	stream	40°40′S; 66°10′W	summer	183	95	−75.44			−9.03		
9	BP Cuiña	spring	39°39′S; 66°54′W	summer	64	203	−77.9	−79	1.73	−8.99	−8.98	0.02
							−80.34			−8.96		
10	BP Trapalcó 2	spring	39°45′S; 66°51′W	summer	68	192	−73.92	−74.1	0.21	−7.95	−7.8	0.22
							−74.22			−7.65		
11	CF Santa Victoria	spring	40°0′S; 67°42′W	summer	523	242	−72.21			−8.89		
12	SFW Santana	spring	41°20′S; 69°31′W	summer	916	380	−93.45			−11.83		
13	SFW Chacras	spring	41°21′S; 69°32′W	summer	1044	382	−99.6			−12.48		
$n = 46$												

NR: Negro River; CR: Colorado River; BP: Closed Basins and Plains; CF: Cuy Foothills; SFE: Somuncurá Foothills East; SFW: Somuncurá Foothills West. Alt.: Altitude. Dist.: distance from the coast. *Summer: January; Winter: July. **Samples removed from posterior analysis for being anomalous and or suspected of contamination (Chris Harris, Personal communication 2016).

3.3. Altitude and distance from the coast effects

Fig. 5 shows similar trends in altitude and distance effects: higher altitudes and greater distances to the coast, produce a more ^{18}O-depleted values (Fig. 5). Nevertheless, the trend in the Negro River is weak and non-significant (altitude: $r = -0.37$, $p = 0.081$; distance: $r = -0.38$, $p = 0.073$). The rest of the water samples correspond to streams or springs and, as pointed above, the $\delta^{18}O$ values become more negative with increasing altitude and or distance from the coast (SFW: $\delta^{18}O_{Chacras} = -12.48‰$ and $\delta^{18}O_{Santana} = -11.83‰$; SFE: $\delta^{18}O_{Valcheta} = -9.03‰$; CF: $\delta^{18}O_{Santa\ Victoria} = -8.89‰$; BP: $\delta^{18}O_{Cuiña} = -8.98 \pm 0.02‰$ and $\delta^{18}O_{Trapalcó\ 2} = -7.80 \pm 0.22‰$) (Fig. 5).

4. Discussion

4.1. Terrestrial surface water in central-eastern North Patagonia

We compared the isotopic data from terrestrial surface water (TSW) sources with the GMWL and predictions of precipitation (PP) (Fig. 2). Given the arid to semi-arid conditions, the slightly lower slope showed by TSW compared to the GMWL was within the expectations due to evaporative effects. Also, the results show that the TSW isotopic values do not match those of PP. Both seasonally and annually calculated values are higher when compared with the water samples. Several hydrogeochemical (e.g. Fritz et al., 1987; Ingraham and Taylor, 1991; Dutton et al., 2005) and archaeological studies (e.g. Daux et al., 2008; Knudson, 2009; Scherer et al., 2015) have shown that $\delta^{18}O$ from water samples may differ from rainfall and that the surface water oxygen values not always follow the expected patterns based on environmental variables. In our case, this discrepancy could be due to the lack of local empirical references (i.e. GNIP stations), which increases the interpolation uncertainty and makes the model predictions weaker (see Bowen and Revenaugh, 2003). However, post-precipitation processes might be playing an important role in inducing this difference as well (Kendall and Coplen, 2001). Incorporation of rainfall into terrestrial systems is neither complete nor indiscriminate, as some of the rain will

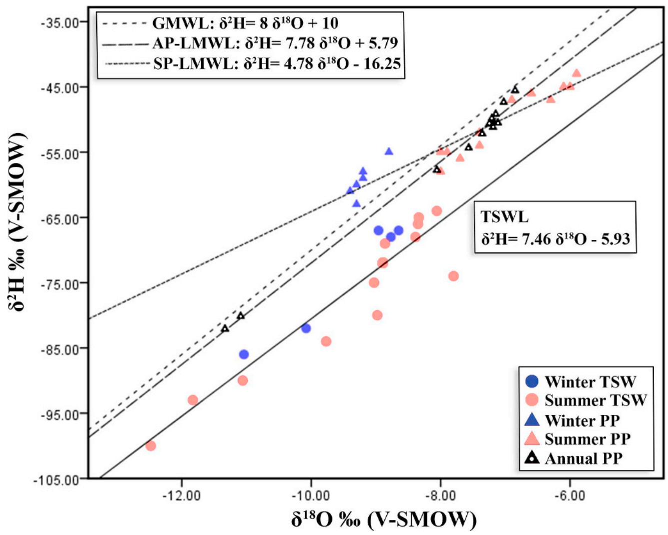

Fig. 2. δ^2H and δ^{18}O from the analyzed water samples and from the predicted precipitations. TSW: terrestrial surface water; PP: precipitation prediction; GMWL: global meteoric water line; AP-LMWL: annually predicted local meteoric water line; SP-LMWL: seasonally predicted local meteoric water line; TSWL: terrestrial surface water line.

return to the atmosphere by evapotranspiration, leaving the rest to infiltrate into the ground or become surface runoff that will feed directly water bodies. In arid environments, isotopic shifts between that incident rain and the runoff are especially significant (Gat, 1998).

Oxygen is a complex isotopic tracer, not just for its fractionation inside of the human body, but because the kind of baselines that we use impacts directly on the archaeological inferences (Knudson, 2009; Scherer et al., 2015; see also discussion in Barberena et al., 2017). The assumption that $\delta^{18}O_{pp}$ is an isotopic mirror of $\delta^{18}O_{tsw}$ has a significant archaeological implication, because it could lead to an overestimation of "non-locals" in provenance studies (e.g. Budd et al., 2004; see also Brettell et al., 2012). Our results agree with those presented by Knudson (2009: 185), who remarked that "attempting to match oxygen isotope signatures in precipitation to oxygen isotope signatures in human remains, without a more complex study of drinking water sources, is inadvisable" (see also Scherer et al., 2015). It is unlikely that the daily water intake of hunter-gatherers from central-eastern North Patagonia would has been based on direct drinking of rainwater. Apart from some rainy tropical areas, people do not base their water consumption directly from rain, but from sources that could be or not be fed by local precipitation. So, the main question may be how local meteoric water is incorporated into the terrestrial system and, more important, how

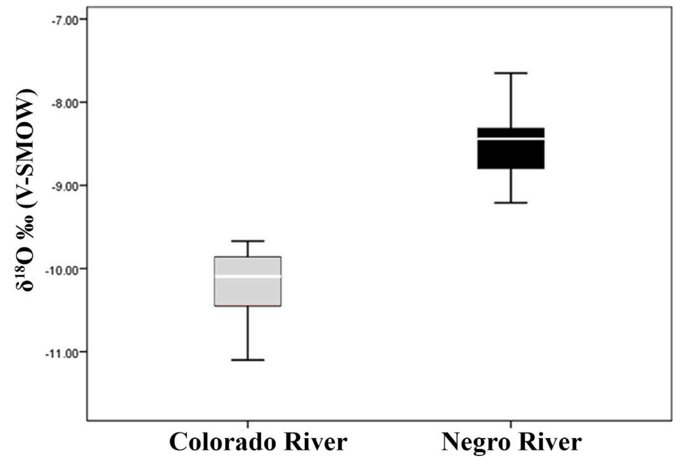

Fig. 4. Colorado and Negro rivers δ^{18}O distribution (seasonal data pooled together).

much it is contributing to the human available drinking water sources.

Regarding the seasonal variation in the δ^{18}O values from the lower Colorado and Negro rivers, we observe that both have more positive values on summer (Fig. 3). This is probably due to isotopic shifts over the precipitation that feeds them, since large rivers with abundant flow are usually poorly affected by evaporation (Darling, 2004). Although the difference is small, the middle basin of the Colorado showed an inverse pattern with more negative δ^{18}O values during the summer. Given the Andean influence over this river, these shifts are expected as inputs of ^{18}O-depleted melt water product of the snow melt during warm periods (Bottomley et al., 1986; Mook, 2002). Since there are no tributaries in between the sampling points, the ^{18}O-depleted summer signal from the middle basin -not registered in the lower basin-could be due to a stronger input of groundwater generated by low elevation rainfall near the delta of the Colorado River (Schäbitz, 2003). Our results also show that the δ^{18}O seasonal variation in both rivers is small (< 1‰, Fig. 3) (see Mook, 1970; Rank et al., 1998; Rosa et al., 2016), and could be induced by the recharge of long-term rainwater stored in shallow aquifers. In some cases, the groundwater can be the dominant contributor to the flow of a watercourse and weaken or suppress the season variation (Kendall and Coplen, 2001). This is because groundwater systems usually tend to represent an average of the rainfall (Clark and Fritz, 1997; although see Kendall and Coplen, 2001: 1385). Regarding the isotopic difference between river basins, the Negro is ^{18}O-enriched in relation to the Colorado (Fig. 4). One possible explanation for this could be that the Limay River, which provides ~58% of the Negro, is born from the Nahuel Huapi lake and is fed by a system of lakes that dams the water (Soldano, 1947) potentially generating ^{18}O-

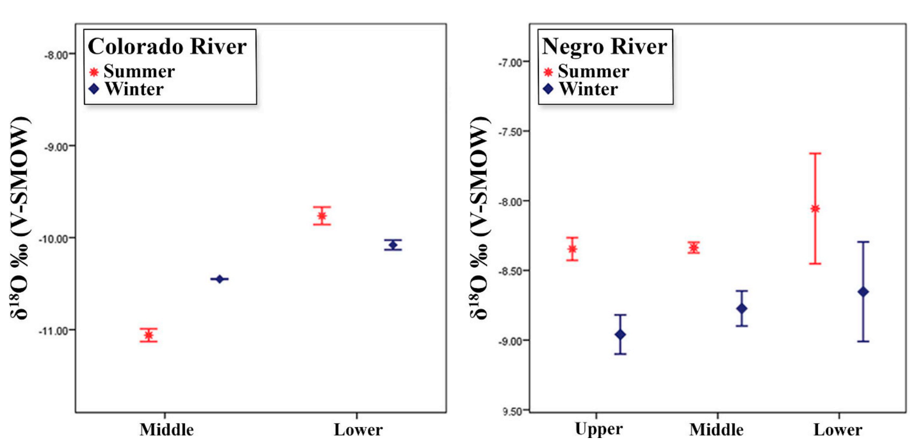

Fig. 3. Colorado and Negro basins δ^{18}O distribution according to the season.

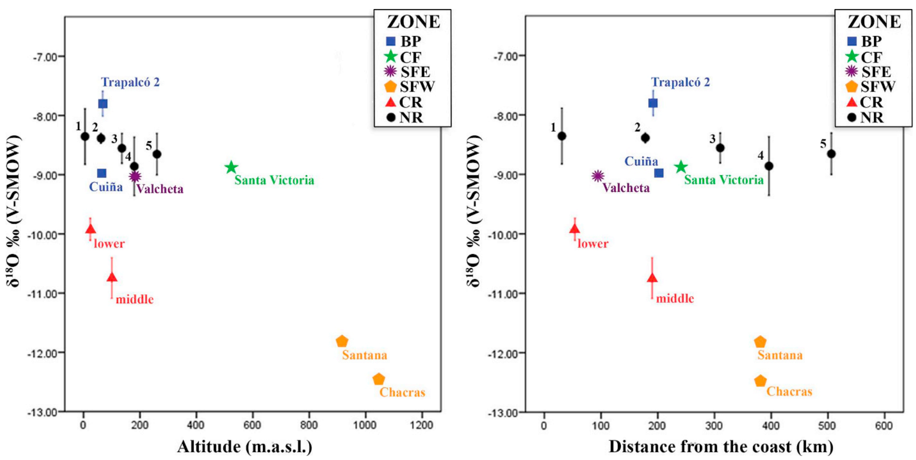

Fig. 5. Water samples $\delta^{18}O$ (mean ± σ and individual values) in relation to the altitude (left) and distance from the coast (right), according to the geographic zone. BP: Closed Basins and Plains; CF: Cuy Foothills; SFE: Somuncurá Foothills East; SFW: Somuncurá Foothills West; CR: Colorado River; NR: Negro River. The numbers correspond to the sectors of the Negro basin: 1-lower, 2-mid-low, 3-middle, 4-mid-up, 5-upper.

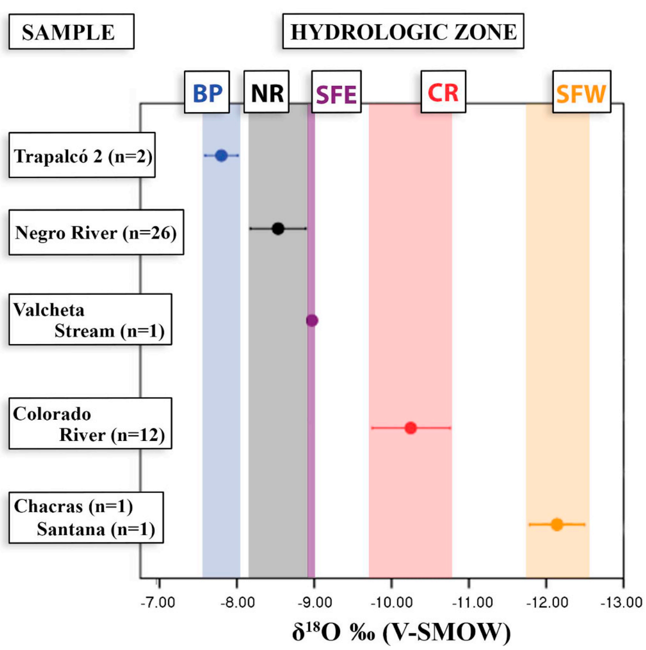

Fig. 6. $\delta^{18}O$ hydrologic zones (mean ± SD and individual values). BP: Closed Basins and Plains; NR: Negro River; SFE: Somuncurá Foothills East; CR: Colorado River; SFW: Somuncurá Foothills West.

Table 2
Hydrologic zones of central-eastern North Patagonia.

Area	Source	*n*	Mean $\delta^{18}O$	SD	Mean ± SD
BP	spring	2	−7.80	0.22	−7.58/−8.02
NR	river	26	−8.53	0.36	−8.17/−8.89
SFE	stream	1	−9.03	–	–
CR	river	12	−10.25	0.49	−9.76/−10.74
SFW	spring	2	−12.15	0.47	−11.68/−12.62

BP: Closed Basins and Plains; NR: Negro River; SFE: Somuncurá Foothills East; CR: Colorado River; SFW: Somuncurá Foothills West.

enriched signals. The negative $\delta^{18}O$ values for the Colorado River are expected because of the influence of rainfall and melt water from high altitudes (Hoke et al., 2009; Ugan et al., 2012).

Some of our results agree (Fig. 5), to a certain degree, with the expectations for precipitation according to the Dansgaard's effects (Rozanski et al., 1993; Gat, 2005): as the altitude or distance from the coast increases, the waters become progressively ^{18}O-depleted. This is because the preferential removal of the heavy isotopes during condensation, as the air masses move inland and loss moisture, is responsible for both altitude and distance effects (Rozanski et al., 1993). Nevertheless, the different types of the water sources make these isotopic shifts not completely straightforward to compare all the samples against each other. For example, some samples come from similar altitude/distance from the coast (~150–200 m/~200–250 km) but their isotopic variation is large (middle Colorado River *vs.* Valcheta Stream ~1.7‰, or the former *vs.* Trapalcó 2 spring ~3‰, Fig. 5). Unlike the Colorado River that remarkably changes along its course (~0.8‰/100 m/150 km), the Negro River slightly varies (~0.1–0.2‰/100 m/200 km) (see similar trends in altitude for Central-western Argentina in Ugan et al., 2012: 5 and in distance from the coast for western Europe in Rozanski et al., 1993: 6). Water in rivers tends to retain the signals from the catchment area (Mook, 1970, 2002; Dutton et al., 2005), unless some other influxes (e.g. tributaries) or evaporation deviate the isotopic values. As pointed above, the observed isotopic shifts along the Colorado River could be explained by a stronger input of groundwater in the lower basin, but a conclusive answer remains elusive. The Negro River lacks of tributaries along its course, shallow aquifer recharge and or strong evaporation, therefore it is not surprising to find constant values along its course (see an example from an arid environment in Aravena and Suzuki, 1990).

The groundwater samples from western Somuncurá Foothills -Chacras and Santana-have both negative $\delta^{18}O$ values compatible with a geographic location higher and farther from the coast (Fig. 5). It is likely that those springs are being recharged by local rainfall infiltration, common in the study area (Vogel et al., 1975). When the recharge is local, the isotopic composition of groundwater usually represents the bulk of the rainfall (Gonfiantini et al., 1998; Darling, 2004). Nevertheless, it is possible that a particular region presents a "multi-aquifer" arrangement, with several units that can have different isotopic compositions owing to different sources of recharge (Aravena, 1995; Darling, 2004). This kind of complexity seems to be present in the Closed Basins and Plains area. While some groundwater samples have different altitudes and similar $\delta^{18}O$ values -Cuiña and Santa Victoria-, other samples are nearby but are different in oxygen composition -Trapalcó 2 and Santa Victoria- (Fig. 5). Although rainfall is usually the main recharge agent, hydraulic connections with lakes and leakage from rivers can also contribute; so that groundwater will correspond to the trends exhibited by those water bodies (Meyers et al., 1993; Gonfiantini et al., 1998; Darling, 2004). For instance, this situation has been observed in the province of San Juan -Argentina-, where recharge occurs by rivers that descend from the mountains, generating ^{18}O-depleted water inside the aquifers with respect to the local rainfall (Albero et al., 1987).

Among other factors that can affect the groundwater signals (Moser and Stichler, 1972; Sidle, 1998), the ground permeability is relevant, since it determines the speed with which atmospheric water will enter to the underground system and, consequently, how subject to evaporation it may be (Gat, 1996). In this regard, the Valcheta Stream sample from the Somuncurá Foothills constitutes a case of fast entry and retention of the local rainfall signals (Fig. 5). Somuncurá Plateau acts as a natural reservoir of water, since its basaltic mantle allows a fast infiltration of precipitation (Fontana, 2001). Recent research has confirmed the fast introduction of the atmospheric water into the Somuncurá subterranean system with null or scarce evaporation (Parica et al., 2014), and indicated that the water sources that spring up from the plateau are consistent with high elevation zones (Remesal et al., 2016).

4.2. Data integration: towards a baseline for central-eastern North Patagonia

Considering the previous assessments, we defined five "hydrologic zones" based on their stable oxygen values (Fig. 6, Table 2). Each one has a particular $\delta^{18}O_{tsw}$ and represents a geographic location or watercourse: Closed Basins and Plains; Negro River; Somuncurá Foothills East; Colorado River; and Somuncurá Foothills West. Although some of the zones are defined with few samples, the differences among some of them are large enough to make promising the use of this isotopic tracer to assess paleomobility.

5. Conclusions

This paper sought to explore the potential of building a terrestrial surface water baseline to assess paleomobility, characterizing isotopically ($\delta^{18}O$) water sources from central-eastern North Patagonia. Through the comparison between our data and the usually used predicted precipitation values, we highlight the importance of analyzing water sources instead of using just precipitation because one does not exactly mirror the other. In the study area, different processes (evaporation, water mixing and transport) deviate the isotopic signal from the sources respect to the local rain.

Based on the stable oxygen composition of the water sources, we characterized and differentiated five geographic areas as "hydrologic zones": Colorado River, Negro River, Closed Basins and Plains, Somuncurá Foothills East and West. Although the scope of this work is limited by the sample size, the differences among the hydrologic zones make evident the potential of the $\delta^{18}O_{tsw}$ baseline for paleomobility purposes. Finally, we highlight the importance of increasing the number of samples, in order to strengthen the resolution of this baseline and enhance our hydrologic understanding of northern Patagonia.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.quaint.2019.06.008.

Appendix

Table A.1
Isotopic predictions of modern precipitation calculated with OIPC (Bowen, 2018) for the same coordinates as the water samples used in this paper.

Location	Coordinate	Season	Altitude (m.a.s.l.)	δ^2H	$\delta^{18}O$
NR upper	38°58′49.08″S; 68°3′18.42″W	summer	260	−54	−7.4
		winter	–	−63	−9.3
		annual	–	−51	−7.2
NR mid-up	39°6′3.66″S; 66°38′55.86″W	summer	180	−46	−6.6
		annual	–	−51	−7.3
NR middle	39°25′21.1″S; 65°43′6.9″W	summer	136	−45	−6.1
		winter	–	−61	−9.4
		annual	–	−50	−7.2
NR mid-low	40°6′5.61″S; 64°27′19.49″W	summer	62	−55	−7.9
		winter	–	−60	−9.3
		annual	–	−50	−7.2
NR lower	40°48′53.97″S; 62°59′0.06″W	summer	5	−55	−8
		winter	–	−59	−9.2
		annual	–	−49	−7.2
CR middle	38°58′58.5″S; 64°06′0.1″W	summer	100	−43	−5.9
		winter	–	−58	−9.2
		annual	–	−47	−7
CR lower	39°31′33.93″S; 62°39′41.29″W	summer	24	−55	−8
		winter	–	−55	−8.8
		annual	–	−46	−6.9
BP Cuiña	39°39′36.5″S; 66°54′27.5″W	summer	64	−47	−6.3
		annual	–	−52	−7.4
BP Trapalcó 2	39°45′34.33″S; 66°51′46.26″W	summer	68	−45	−6
		annual	–	−51	−7.1
CF Santa Victoria	40°0′11.75″S; 67°42′15.78″W	summer	523	−52	−7.4
		annual	–	−58	−8.1
SFE Valcheta	40°40′49.83″S; 66°10′26.06″W	summer	183	−47	−6.9
		annual	–	−54	−7.6

(continued on next page)

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Table A.1 (*continued*)

Location	Coordinate	Season	Altitude (m.a.s.l.)	δ^2H	$\delta^{18}O$
SFW Santana	41°20′22.56″S; 69°31′40.67″W	summer	916	−56	−7.7
		annual	–	−80	−11.1
SFW Chacras	41°21′23.50″S; 69°32′47.60″W	summer	1044	−58	−8
		annual	–	−82	−11.3

NR: Negro River; CR: Colorado River; BP: Closed Basins and Plains; CF: Cuy Foothills; SFE: Somuncurá Foothills East; SFW: Somuncurá Foothills West.

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