Implications of S1 tephra findings in Dead Sea and Tayma palaeolake sediments for marine reservoir age estimation and palaeoclimate synchronisation

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Abstract

Here we report on the first findings of a cryptotephra in the Holocene lacustrine sediment records of the Dead Sea and Tayma palaeolake (NW Arabian Peninsula). The major element glass composition of this rhyolitic tephra is identical to the distal ‘S1’ tephra layer identified in the Yammoûneh palaeolake (Lebanon), in a marine sediment record from the SE Levantine basin and in the Sodmein Cave archaeological site in Egypt. The S1 tephra corresponds to the early Holocene ‘Dikkartın’ dome eruption of the Erciyes Dağ volcano in central Anatolia (Turkey) and has been dated in the marine record at 8830 ± 140 cal yr BP. We present new age estimates of the S1 tephra based on radiocarbon dating of terrestrial plant remains and pollen concentrates revealing ages of 8939 ± 83 cal yr BP in the Dead Sea sediments and 9041 ± 254 cal yr BP in Tayma. The precise date from the Dead Sea allows refining the early Holocene marine reservoir age in the SE Levantine Sea to ca. 320 ± 50 years. Synchronisation of marine and terrestrial palaeoclimate records in the eastern Mediterranean region using the S1 tephra further suggests a time-transgressive expansion of the early Holocene humid period.

Keywords: Early Holocene; Tephrochronology; S1 Tephra; Palaeoclimate; Lake Sediments; Eastern Mediterranean/Levant; NW Arabian Peninsula; Marine Reservoir Age
1 Introduction

Identifying tephra (volcanic fall material) in lacustrine and marine sediment records is important for the dating and regional synchronization of palaeoclimate and archaeological records (e.g., Davies et al., 2012; Lane et al., 2013; Lowe et al., 2015; Wulf et al., 2013). Methodological advances enabled the finding of cryptotephras (non-visible by naked eye) in the sediments, which have multiplied the number of identified tephras and greatly extended the area of tephra dispersal (e.g., Davies, 2015; Lane et al., 2014; Satow et al., 2015; Wulf et al., 2013; 2016).

In the eastern Mediterranean region, three main volcanic centres build the source regions for late Quaternary tephra dispersals, i.e. the Italian volcanic provinces, the Hellenic Arc and central and eastern Anatolia (Fig. 1). Comprehensive tephrostratigraphical efforts led to a profound knowledge of the chemical compositions and spatial distribution of late Quaternary Mediterranean tephra deposited in marine and terrestrial sediment records (e.g., Bourne et al., 2010; Federman and Carey, 1980; Keller et al., 1978; Narcisi and Vezzoli, 1999; Tomlinson et al., 2015; Wulf et al., 2002; 2004; Zanchetta et al., 2011). So far, there is no evidence for regionally significant volcanic eruptions and tephra dispersal from the Arabian volcanic province including the Harrat Ash Shaam field (Fig. 1) of the southern Levant and northern Arabia in the late Quaternary (e.g., Hamann et al., 2010; Zanchetta et al., 2011). Furthermore, there is still a lack of tephra studies in the Levant, which is crucial for synchronising palaeoenvironmental records and archaeological events in this climatically sensitive area.

The only exception is the finding of the early Holocene ‘Dikkartın’ tephra from the Erciyes volcano in central Anatolia, Turkey, that has been reported as a visible layer in the Yammoûneh palaeolake in Lebanon (Develle et al., 2009) and in the marine core SL112 from the SE Levantine basin (Hamann et al., 2010), and as a cryptotephra at the Sodmein Cave archaeological site in Egypt (Barton et al., 2015) (Fig. 1). This so-called S1 tephra has been dated in the marine record at 8365 ± 65 14C yr BP, resulting in a calibrated age range of 8970 – 8690 cal yr BP when applying a reservoir age of 400 years (Hamann
et al., 2010). The apparently widespread distribution of the S1 tephra encouraged us to search for evidence of this tephra in the lacustrine records from the Dead Sea and the Tayma palaeolake in NW Arabia (Fig. 1). Identification of the S1 cryptotephra in these records is expected to (1) extend the S1 tephra distribution, (2) refine the marine reservoir age for the early Holocene in the SE Levantine Sea, and (3) discuss lead and lag phase relationships of precipitation patterns in the Levantine-Arabian region.

2 Regional setting

The sedimentary record of the hypersaline Dead Sea is one of the most important palaeoclimate archives in the Levantine region (e.g., Enzel et al., 2006; Neugebauer et al., 2014; Stein, 2001). The Dead Sea is situated at the boundary between Mediterranean climate and the hyperarid Saharo-Arabian desert belt, where small changes in precipitation over its drainage area are sensitively recorded through changing lake levels and in sedimentation (e.g., Enzel et al., 2003; Stein et al., 2010). The Dead Sea is located ca. 250 km southeast of marine coring site SL112 (Fig. 1) and in a favourable wind position to Eastern Mediterranean volcanic sources. The Tayma palaeolake is located within the Saharo-Arabian desert belt in the hyperarid interior of the NW Arabian Peninsula and north of the oasis of Tayma within an area of rich archaeological heritage (Dinies et al., 2015; Eichmann et al., 2006; Engel et al., 2012).

3 Materials and methods

3.1 Sediment sampling

The selective search for the S1 tephra used radiocarbon-dated sediments of the ca. 21 m long core DSEn from the western margin of the Dead Sea (Migowski et al., 2004), and the ca. 6 m long core Tay253 from the eastern part of the Tayma palaeolake. The latter is correlated to the dated cores Tay220 and Tay254 (Dinies et al., 2015) using macroscopic sediment marker layers. The early Holocene
Dead Sea (DSEn) and Tayma (Tay253) sediment sections were sampled in continuous 5 cm increments with sample volumes of 5 cm$^3$ at around the expected sediment depth of the S1 tephra.

From the overlapping Dead Sea core sections DSEn-A13u and DSEn-B11u, 21 samples were taken from 17.11 to 18.16 m composite depth covering the time interval between 8758 ± 71 and 9138 ± 73 cal yr BP. The sediments consist of finely laminated, greyish, brownish or black detrital marls that alternate with primary aragonite or gypsum laminae and thicker layers of intraclast breccias.

Twelve samples were collected from Tayma (core sections Tay253_4.5-5.3 and Tay253_5-5.3) at 5.38-5.95 m composite depth covering the time interval from 8541 ± 41 to 9406 ± 411 cal yr BP. The sediments are grey, clayey-silty siliciclastics and brownish marls with carbonaceous laminations. The lowermost ca. 0.5 m include reworked bedrock (Ordovician clay and sandstone).

3.2 Tephrochronological methods

We identified glass shards using physical-chemical treatment of sediments, i.e. rinsing in deionised water to remove highly saline pore waters, adding 10% HCl and 15% H$_2$O$_2$ solutions to remove carbonates and organic matter, wet-sieving into a 20-100 µm grain size fraction, and SPT liquid density mineral separation (2-2.65 g/cm$^3$). The residual light fraction was inspected for glass shards under a transmitted light microscope. Identified glass shards were hand-picked into single-hole-stubs, embedded in Araldite 2020 epoxy resin, sectioned and polished for electron probe microanalyses (EPMA).

The major-element composition of individual glass shards was measured using a JEOL JXA-8500F electron microprobe with operating conditions of 15 kV voltage, 10 nA beam current, and a beam size of 5 µm. Exposure times for each analysis were 20 s for the elements Fe, Cl, Mn, Ti, Mg and P, as well as 10 s for Si, Al, K, Ca and Na. Instrumental calibration used natural mineral and glass standards, i.e. Lipari obsidian (Hunt and Hill, 1996; Kuehn et al., 2011) and MPI-Ding glasses ATHO-G, StHs-6-80-G and GOR-132-G (Jochum et al., 2006) (Supplement A).
3.3 Chronologies

We calculated age-depth models for the DSEn and Tay253 sediment profiles based on radiocarbon ages published by Migowski et al. (2004) and Dinies et al. (2015). These were introduced into a P-Sequence deposition model in OxCal v4.2 (Bronk Ramsey, 2008) applying the IntCal13 calibration curve (Reimer et al., 2013).

4 Results and discussion

4.1 Tephra identification

Volcanic glass shards were detected in both, the Dead Sea and Tayma sediments (Fig. 2). In the Dead Sea, the highest glass shard concentration (n >50) with the largest grain size (d_max = 70 µm) occurred within finely laminated sediments between 17.61 and 17.56 m depth (Fig. 2a). Glass shards have been further detected within intraclast breccias up to 45 cm above the primary layer at 17.61 m depth (Fig. 2a). Identified glass shards in the Tayma record spread over 38 cm from 5.86 to 5.48 m sediment depth with the first and highest glass shard concentration (n = 12) appearing in the lowermost sample at 5.86-5.81 m depth (mean isochron depth at 5.835 m), close to the bottom of the sediment core (Fig. 2a).

All glass shards are colourless and rich in elongated vesicles (Figs. 2b, 2c). The chemical composition of the shards both in the primary cryptotephra and in reworked layers in the Dead Sea and Tayma cores is identical and homogeneous rhyolitic (Fig. 3a). The normalised silica concentrations of glass shards range between 75.6 and 77.8 wt.%, while calcium and iron values show variations of 0.9-1.5 and 1.0-1.6 wt.%, respectively (Table 1, Fig. 3, Supplement A).

Only few rhyolitic tephras of Holocene age are reported in the Mediterranean region, i.e. the Minoan (Z-2) Ash from Santorini (Keller et al., 1978), the Gabellotto-Fiumebianco eruption of Lipari (Aeolian Islands) (Siani et al., 2004), and the early Holocene Dikkartın, Perikartın and Karagüllü eruptions of
Erciyes Dağ in the central Anatolian volcanic province (Hamann et al., 2010, and references therein; Zanchetta et al., 2011) (Fig. 1). Rhyolitic tephra from the Arabian volcanic province have not been evidenced for the Holocene period. Comparing the glass compositions of the Dead Sea and Tayma cryptotephras with the compositional fields of rhyolitic Holocene tephras from the eastern Mediterranean (Fig. 3) excludes Italian and Aegean Arc volcanoes as a source, but implies an origin from the Erciyes volcano in central Anatolia (Figs. 1, 3). The best match is given with distal S1 tephra occurrences in the Levantine Sea (Hamann et al., 2010) and Yammoûneh (Develle et al., 2009) (Table 1, Fig. 3). Slight offsets of SiO2, Al2O3 and Na2O glass compositions of the S1 tephra in Sodmein cave (Egypt; Barton et al., 2015) and the proximal S1 correlative, the ‘Dikkartın’ tephra (Hamann et al., 2010) (Fig. 3), are likely related to analytical problems such as sodium migration during EPMA measurements of the glass shards of the Dead Sea and Tayma cryptotephras with a 5 µm beam (see also Lipari reference glass data for comparison, Supplement A). The small size of glass shards in our records (~50-70 µm) hindered trace element to further support the correlation with distal S1 tephra occurrences and Erciyes proximal tephras.

4.2 S1 tephra and marine reservoir age constraints

P-Sequence deposition models for the Dead Sea (DSEn) and Tayma (Tay253) cores revealed modelled ages of 8939 ± 83 cal yr BP and 9041 ± 254 cal yr BP for the samples with the earliest glass shard occurrence, respectively (Table 2; Fig. 2). These ages are within the 2σ error range of published calibrated radiocarbon ages (Develle et al., 2009; Hamann et al., 2010) (Table 2). The larger uncertainty of the Tayma age is due to age extrapolation from the oldest radiocarbon date to the earliest tephra occurrence 32 cm below. Since the age uncertainty of ± 850 yrs in the Yammoûneh record is even larger (Develle et al., 2009), the age of 8939 ± 83 cal yr BP from the Dead Sea is the most precise terrestrial date for the S1 tephra.
This allows refining the marine reservoir age through calculating the difference of the marine (8365 ± 65 14C yr BP; Hamann et al., 2010) and the terrestrial 14C age. The resulting marine reservoir age of ca. 320 ± 50 years (Table 2) is younger than the common Mediterranean Sea 14C reservoir age estimate of ~400 years (Siani et al., 2001). This can be explained by higher freshwater flux from the Nile River to the Levantine Sea during the humid early Holocene. Variations in marine reservoir ages have been previously demonstrated from other regions (Bard, 1988; Bard et al., 1994; Çağatay et al., 2015; Kwiecien et al., 2008).

4.3 Palaeoclimatic implications

During the early Holocene, enhanced summer insolation led to a stronger African monsoon, resulting in enhanced Nile River runoff and the formation of a sapropel layer (S1) in the eastern Mediterranean Sea (deMenocal et al., 2000; Rossignol-Strick, 1985). At the same time, stronger cyclone activity in the Mediterranean promoted increased winter rains in the eastern Mediterranean (Kutzbach et al., 2014). In the SE Levantine Sea, the S1 tephra occurs in the lower part of the sapropel ('S1a'; ca. 9.6-8.2 cal ka BP; Hamann et al., 2010; Kuhnt et al., 2008) coinciding with a period of higher inland rainfall at Yammoûneh (ca. 9.5-8.5 cal ka BP; Develle et al., 2010) and a higher lake level of the Dead Sea (ca. 10-8.6 cal ka BP; Migowski et al., 2006) (Fig. 4). In contrast, the S1 tephra in the Tayma record appears 3-4 centuries before the main humid phase commencing in this region with grassland expansion and initial lake development at ca. 8.6 ka BP (Dinies et al., 2015). This indicates a lagged onset of the humid phase on the Arabian Peninsula compared to the SE Mediterranean and Levant. This time-transgressive expansion of regional humidity further suggests that the main moisture source at Tayma (NW Arabia) was the Mediterranean (Enzel et al., 2015), while moist early Holocene conditions in the southern Arabian peninsula were related to an intensified monsoon (Fleitmann et al., 2007).

5 Conclusions
1) The early Holocene S1 tephra is the first volcanic ash identified in the Dead Sea sediment record and the first Mediterranean tephra ever documented on the Arabian Peninsula, located 1240 km to the south of the volcanic source. This extends the spread of Mediterranean tephra further to the Southeast.

2) Based on $^{14}$C dating from the Dead Sea, we provide a precise terrestrial S1 tephra age of $8939 \pm 83$ cal yr BP, which allows to refine the marine reservoir age for the early Holocene SE Levantine Sea to $320 \pm 50$ years.

3) The occurrence of the S1 tephra in the Dead Sea and Tayma sediments makes this tephra a key isochron for synchronising marine and terrestrial climate records in a large region reaching from the SE Mediterranean-Levantine to the NW Arabian Peninsula. Synchronising palaeoclimate records from the Levantine Sea, Yammoûneh, the Dead Sea and Tayma using the S1 tephra suggests a time-transgressive expansion of humid conditions over a few centuries from the southeastern Mediterranean and the Levant towards the northern Arabian Peninsula during the early Holocene.

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Table 1: Mean normalised (volatile-free) major-element glass data in wt.% with 1σ standard deviation (SD) and original totals of proximal pumices from the Dikkartın eruption of Erciyes Dağ volcano and distal S1 tephra correlatives (Yammoûneh, core SL112 from the SE Levantine Sea, Sodmein Cave, Dead Sea and Tayma); references: #1 Hamann et al. (2010); #2 Tomlinson et al. (2015); #3 Develle et al. (2009); #4 Barton et al. (2015).

<table>
<thead>
<tr>
<th>Site (reference)</th>
<th>Erciyes Dağ (Dikkartın [#1, 2])</th>
<th>Yammoûneh [#3]</th>
<th>SE Levantine Sea [#1]</th>
<th>Sodmein Cave [#4]</th>
<th>Dead Sea (this study)</th>
<th>Tayma (this study)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>n = 12 SD</td>
<td>n = 10 SD</td>
<td>n = 11 SD</td>
<td>n = 16 SD</td>
<td>n = 24 SD</td>
<td>n = 16 SD</td>
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<tr>
<td>SiO₂</td>
<td>76.01 ± 0.47</td>
<td>77.07 ± 0.49</td>
<td>77.17 ± 0.49</td>
<td>75.71 ± 0.27</td>
<td>76.68 ± 0.56</td>
<td>76.35 ± 0.23</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.20 ± 0.04</td>
<td>0.24 ± 0.12</td>
<td>0.19 ± 0.03</td>
<td>0.21 ± 0.04</td>
<td>0.21 ± 0.03</td>
<td>0.20 ± 0.03</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>13.40 ± 0.32</td>
<td>13.37 ± 0.42</td>
<td>13.15 ± 0.35</td>
<td>13.60 ± 0.17</td>
<td>13.57 ± 0.24</td>
<td>13.66 ± 0.09</td>
</tr>
<tr>
<td>FeO</td>
<td>1.31 ± 0.08</td>
<td>1.34 ± 0.19</td>
<td>1.19 ± 0.12</td>
<td>1.22 ± 0.08</td>
<td>1.19 ± 0.09</td>
<td>1.25 ± 0.05</td>
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<td>MnO</td>
<td>0.05 ± 0.04</td>
<td>0.07 ± 0.07</td>
<td>0.07 ± 0.04</td>
<td>0.03 ± 0.04</td>
<td>0.05 ± 0.04</td>
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<td>MgO</td>
<td>0.28 ± 0.03</td>
<td>0.24 ± 0.11</td>
<td>0.29 ± 0.03</td>
<td>0.29 ± 0.03</td>
<td>0.27 ± 0.02</td>
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<tr>
<td>CaO</td>
<td>1.48 ± 0.08</td>
<td>1.37 ± 0.27</td>
<td>1.47 ± 0.10</td>
<td>1.46 ± 0.07</td>
<td>1.45 ± 0.13</td>
<td>1.46 ± 0.04</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.76 ± 0.16</td>
<td>2.89 ± 0.41</td>
<td>2.95 ± 0.61</td>
<td>3.81 ± 0.32</td>
<td>3.23 ± 0.39</td>
<td>3.39 ± 0.21</td>
</tr>
<tr>
<td>K₂O</td>
<td>3.50 ± 0.12</td>
<td>3.40 ± 0.30</td>
<td>3.53 ± 0.18</td>
<td>3.48 ± 0.10</td>
<td>3.31 ± 0.14</td>
<td>3.34 ± 0.08</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>n.a.</td>
<td>n.a.</td>
<td>n.a.</td>
<td>0.04 ± 0.08</td>
<td>0.04 ± 0.02</td>
<td>0.05 ± 0.02</td>
</tr>
<tr>
<td>Original total</td>
<td>96.95 ± 1.68</td>
<td>94.96 ± 0.99</td>
<td>91.87 ± 0.92</td>
<td>95.76 ± 1.01</td>
<td>94.46 ± 0.73</td>
<td>95.02 ± 0.69</td>
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Table 2: Age estimates of the S1 tephra derived from radiocarbon chronologies.

<table>
<thead>
<tr>
<th>Location (sediment core)</th>
<th>S1 tephra age (¹⁴C yr BP)</th>
<th>Calibrated S1 tephra age (cal yr BP)</th>
<th>Dated material</th>
<th>Age estimation</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dead Sea (DSEn-A13u)</td>
<td>8049 ± 40</td>
<td>8939 ± 83</td>
<td>Terrestrial macro-remains</td>
<td>Deposition model (Bayesian) *</td>
<td>Migowski et al. (2004), this study</td>
</tr>
<tr>
<td>Tayma, NW Arabia (Tay253_5-5.3)</td>
<td>-</td>
<td>9041 ± 254</td>
<td>Pollen concentrates</td>
<td>Extrapolation of deposition model (Bayesian) *</td>
<td>Diness et al. (2015), this study</td>
</tr>
<tr>
<td>SE Levantine Sea (GeoTü SL112)</td>
<td>8365 ± 65</td>
<td>8830 ± 140</td>
<td>Planktonic foraminifera</td>
<td>Directly at tephra position; reservoir age estimation: 400 yrs</td>
<td>Hamann et al. (2010)</td>
</tr>
<tr>
<td>Yammoûneh, Lebanon (TR02; YAM04C)</td>
<td>-</td>
<td>~8600 ± 850</td>
<td>Detrital wood fragments</td>
<td>Linear regression</td>
<td>Develle et al. (2009)</td>
</tr>
</tbody>
</table>

* the Dead Sea and Tayma age-depth models were regenerated for this study, applying a P-sequence deposition model in OxCal v4.2 (Bronk Ramsey, 2008) and the IntCal13 calibration curve (Reimer et al., 2013)
Fig. 1: Schematic map showing eastern Mediterranean volcanic provinces and the occurrence of the early Holocene S1 tephra originating from the Erciyes Dağ volcano in central Anatolia, Turkey: Dead Sea and Tayma palaeolake, NW Arabia (this study; red stars); Yammoûneh palaeolake, Lebanon (Develle et al., 2009); marine core GeoTü SL112 from the SE Levantine Sea (Hamann et al., 2010); and Sodmein Cave, Red Sea Mountains, Egypt (Barton et al., 2015) (black crosses).
Fig. 2: a) Dead Sea (core DSEn; zoomed section A13u) and Tayma (core Tay253; zoomed sections 4.5-5.3 and 5-5.3) lithological profiles with sediment core images, radiocarbon ages after Migowski et al. (2004) and Dinies et al. (2015) and modelled ages of the S1 tephra isochron (in italic; this study). Samples including glass shards are coloured in yellow and orange; the latter marking samples from which glass chemistry was measured. Note that the DSEn composite depths have been improved and slightly differ from depths of Migowski et al. (2004); shards/cm³ in DSEn are minimum values due to very high siliciclastic remains after liquid density separation. b) BSE (backscattered electron) image of a S1 glass shard from sample DSEnS1-10. c) Microscope image of S1 glass shards from sample DSEnS1-9.
Fig. 3: a) Total alkali-silica diagram (TAS; Le Bas et al., 1986) with glass compositions of the S1 cryptotephra in the Dead Sea and the Tayma palaeolake (normalised, volatile-free data; this study), Yammoûneh palaeolake, Lebanon (Develle et al., 2009), marine core GeoTü SL112 from the SE Levantine Sea (Hamann et al., 2010), Sodmein Cave, Red Sea Mountains, Egypt (Barton et al., 2015) compared with glass data of proximal pumice deposits of the early Holocene Dikkartın, Karagüllü and Perikartın eruptions of Erciyes Dağ, central Anatolia (Hamann et al., 2010; Tomlinson et al., 2015), and widespread Holocene rhyolitic tephras from Santorini (Kwiecien et al., 2008) and the Aeolian Islands/Italy (Siani et al., 2004). b) – d) Bivariate plots of b) Al₂O₃ versus SiO₂; c) MgO versus SiO₂ and d) FeO versus CaO.
Fig. 4: Comparing palaeoclimate records indicating more humid climate (black bars) during the early Holocene in the Levantine/NW-Arabian region based on synchronisation using the S1 tephra (in red): in the Yammoûneh palaeolake, Lebanon, low δ18O values of ostracods between 9.5 and 8.5 ka BP were interpreted to reflect enhanced inland rainfall (Develle et al., 2009; 2010); in the marine core SL112 from the SE Levantine Sea, sapropel S1 (divided into S1a and S1b at ca. 8.2 ka BP) formed from 9.6 to 6.5 ka BP, but grain size changes suggest an increased suspension load of the Nile River already from ~11 ka BP (Hamann et al., 2008; 2010; Kuhnt et al., 2008); in the Dead Sea, a relatively high lake level from 10 to 8.6 ka BP (up to 22 m above the Holocene average) is inferred from the lithology of several correlated cores, exposures and palaeoshorelines (Migowski et al., 2006; this study); in the Tayma palaeolake record, NW Arabia, palynological analyses suggest that sufficiently high moisture to expand grassland into this region was only available from 8.6 to 8 ka BP (Dinies et al., 2015; this study). Note that in the Levantine records (Yammoûneh, the marine core SL112 and the Dead Sea) the ages of the S1 tephra fall well within the periods of highest moisture availability (grey boxes) during the early Holocene, while in NW Arabia (Tayma) this period occurs only ~400 years later than the deposition of the S1 tephra.