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North Atlantic influence on Holocene flooding in the southern Greater Caucasus

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Abstract

In the context of global climate change, flooding becomes an increasingly serious threat to modern societies, and future flooding can only be understood using long-term geological flood records also encompassing Holocene climate variability. Unlike other regions, Holocene flooding in the Caucasus region is only poorly understood so far: Whereas some rivers originating from the Lesser Caucasus were investigated, no studies exist about rivers originating from the Greater Caucasus. This study investigated the Holocene fluvial dynamics of the upper Alazani River in the southern Greater Caucasus using chronostratigraphic and sedimentologic methods applied to a fluvial sediment-paleosol sequence. By comparing these data with other paleoenvironmental and regional recent hydroclimatic data, we aimed to identify the main drivers of Holocene flooding in the southern Greater Caucasus. Our study shows a link between fluvial sedimentation around 7.3, 5.4, 3.8–2.9 and around 1.7 cal. ka BP and North Atlantic Bond events. Although probably caused by a discharge maximum during spring, fluvial sedimentation is coeval with low regional spring precipitation. As supported by recent hydroclimatic data, intensified floods during Bond events could possibly be explained with more intensive precipitation but also a prolonged snow season during colder winters. The latter would lead to more spring meltwater and thus more intensive spring discharge. Consequently, given increasing annual temperatures because of human-caused global warming, the flood maxima of pluvio-nival rivers in the southern Greater Caucasus may be expected to decrease during the next decades. Our findings underline the need of geological flood records to understand future flood patterns of rivers in mountain regions with complex runoff regimes.
Introduction

In the context of global climate change, flooding and related hazards become an increasingly serious threat to modern societies (Arnell and Gosling, 2016; Brázdil et al., 2006). Generally, future regional and global flood activity can only be understood supported by long-term flood records (Benito et al., 2015a; Gregory et al., 2006) that also encompass the effects of multi-decadal to millennial climate variability, such as solar cycles (Steinilber et al., 2012), Holocene Rapid Climate Change events (RCCs; Mayewski et al., 2004) or Bond events (Bond et al., 2001; Bond et al., 1997) on flood dynamics. Worldwide instrumental flood records cover only locally more than 100 years, and apart from singular exceptions, historical flood records comprise only a few centuries (Brázdil et al., 2006; Jones et al., 2012; Machado et al., 2015; Schmocker-Fackel and Naef, 2010; Schulte et al., 2015). Thus, to obtain longer flood records that also contain multicientennial and millennial flood patterns, geological records from fluvial or lacustrine archives need to be investigated (Brázdil et al., 2006; Jones et al., 2012; Schulte et al., 2015; Wirth et al., 2013; Wolf and Faust, 2015). During the last years, several studies tried to establish regional Holocene flood patterns in Europe and in the Mediterranean region and linked these with long-term climatic patterns and anthropogenic influences (e.g. Benito et al., 2015a, 2015b; Hoffmann et al., 2008; Panin and Matlakhova, 2015; Starkel et al., 2006; Wirth et al., 2013; Zielhofer and Faust, 2008; Zielhofer et al., 2008b). In contrast, the Holocene flood dynamics of the Caucasus region located in the transitional area between Euro-Mediterranean and Asian climates with seasonal influences of the Westerlies, the Siberian High, and far-reaching impacts of the Indian Summer Monsoon (Joannin et al., 2014; Lydolph, 1977) are poorly understood so far. Furthermore, because of strong vertical environmental gradients, this mountainous region can be expected to be especially sensitive to climatic changes that can potentially cause intensive floods (Allam et al., 2009; Schulte et al., 2015). Consequently, there is a need to better understand the future flood dynamics of this densely settled region located in a climatic transition area, and this can possibly be obtained by investigating the Holocene flood history. Yet, the regional Holocene flood dynamics were only investigated for some rivers originating from the Lesser Caucasus (Ollivier et al., 2015, 2016; von Suchodoletz et al., 2015). However, geomorphic and environmental conditions in the Greater Caucasus clearly differ from those in the Lesser Caucasus which is expected to influence local hydroclimatic conditions: Whereas the Lesser Caucasus is characterized by volcanic plateaus with maximal altitudes of 2000–3000 m a.s.l. that were not glaciated since the late Pleistocene (Messager et al., 2013), the bivergent Greater Caucasus mountain belt shows altitudes up to >5000 m a.s.l. and a corresponding higher local relief, is partly glaciated today and generally receives higher precipitation (Forte et al., 2014).

To fill this gap, we studied the Holocene fluvial dynamics of the upper Alazani River in eastern Georgia. The largest part of the catchment of the upper Alazani encompasses the southern slope of the Greater Caucasus (Figure 1a and b). Thus, the fluvial dynamics of this river are controlled by paleohydrological changes in the Greater Caucasus. We performed chronostratigraphic and sedimentologic investigations of a fluvial sediment-paleosol sequence naturally outcropping over several 100 m along its right bank to investigate the Holocene fluvial dynamics of the upper Alazani River. By comparison to other global and regional paleoenvironmental and to regional recent hydroclimatic data, we aim to identify the main drivers of Holocene fluvial dynamics in the southern Greater Caucasus and thereby the main triggers of Holocene hydroclimatic variations in this climatic transition area.

Study area

The Alazani River originates from the southern slope of the central Greater Caucasus in eastern Georgia at ca. 2800 m a.s.l., and flows ca. 240 km from its source into the Kura River draining into the Caspian Sea (Figure 1a and b). After leaving its uppermost N-S directed course with a length of ca. 40 km, the Alazani joins with its main tributary Ilto that also originates from the southern slope of the central Greater Caucasus at an altitude of ca. 2300 m a.s.l. (Figure 1b). Both the upper Alazani and Ilto Rivers show a braided character. During the following ca. 160 km, the river runs from NW to SE in the Alazani Basin, a thrust top basin located between the southern foothills of the Greater Caucasus (Adamia et al., 2010) and the southwesterly advancing Kura fold-andthrust-belt (Kura-FTB; Forte et al., 2010; Figure 1b). The investigated exposure is located at the northern tip of the Alazani Basin, ca.
10 km downstream from the confluence of the upper Alazani and the Iltu Rivers, and has a sub-
catchment of ca. 1100 km². From the confluence onward, the Alazani Basin has a width between 4 and
12 km, widening toward the east, and the width of the braided Alazani River varies between 250 and
600 m. The river is deflected in its course by frontal folds and thrusts of the Greater Caucasus
bivergent orogen (Forte et al., 2014). It is limited toward the south by a scarp of 5–7 m that crops out
mostly fine-grained overbank deposits, whereas the northern slope is rather gradual (Figure 2). The
outcropping overbank sediments in the south form part of a mainly flat surface with a maximal width
of ca. 10 km, gradually rising toward the Kura-FTB in the south. The surface is dissected by several
small creeks that originate from the Kura-FTB in the south and are canalized today: one of these
canals extends ca. 600 m east of the investigated section. However, the robustness of the outcropping
stratigraphy over several 100 m demonstrates that the investigated section is only influenced by
sedimentation of the main Alazani River. Given the large width of the recent floodplain of some 100 m
and the size of the catchment of ca. 1100 km², flooding of the surface on top of the outcropped
overbank sediments in the south that is found between 5 and 7 m above the recent river bed is not
possible under the recent geomorphological and climatological conditions. At the southern margin of
the valley, at some places, a small gravelly terrace with a maximal height of 2 m is found that often
merges with the recent floodplain. It is stratigraphically younger than the outcropped overbank sediments
and probably of (sub-)recent age. Ca. 5–9 km west of the studied site, well recognizable
gravelly former river channels sub-parallel to the recent river course are recognized in the today’s
landsurface. These extend up to 1.5 km northward from the recent river bed (Figure 2).

The Alazani and the Iltu catchments in the Greater Caucasus are formed by folded and metamorphosed
Jurassic flysch and molasse deposits consisting of black shales, sand-/siltstones, and volcanic rocks,
followed downstream by Cretaceous sand-/silt- and limestones. The Kura-FTB that forms the southern
part of the subcatchment of the investigated outcrop reaches altitudes of 2000 m, and is formed by
folded and overthrust Cretaceous sand-/silt- and limestones as well as Paleogene to Quaternary
sandstones, siltstones, and conglomerates (Gamkrelidze, 2003; Figure 2). The catchments of the upper
Alazani and the Iltu Rivers were not glaciated during the Holocene (Gobejishvili et al., 2011). Annual
precipitation in the catchment varies from up to 2000 mm/a in the Greater Caucasus down to ca. 720
mm/a in Akhmeta close to the investigated site (unpublished precipitation map W. Bagrationi,
Geographic Institute Tbilisi; Figure 1c). Most precipitation falls in spring and early summer during
convective events (Lydolph, 1977). Accordingly, the discharge maximum within the Alazani River
between April and June (Figure 1c) is dominated by both snow melt in the Greater Caucasus and the
annual precipitation maximum, that is, the runoff regime has a pluvio-nival character. Subalpine
meadows in the highest parts of the site-subcatchment are followed by mixed beech and hornbeam
forests in the mid-mountain belt and elm-oak-vine forests in the semi-humid parts of the upper Alazani
Basin (Connor and Kvavadze, 2008).

Methods

Stratigraphic mapping and sampling

The stratigraphy of the naturally exposed mostly fine-grained overbank sediments along the southern
bank of the Alazani River was mapped over a distance of ca. 250 m to obtain a robust stratigraphy of
the investigated sequence. Subsequently, a representative profile (42°02′17.7″N, 45°21′18.7″E; 550 m
a.s.l.; Figure 3) was cleaned and described in detail. The sediments between 700 and 800 cm depth
were retrieved by hand drilling using an Eijkelkamp hand driller. The profile was sampled for (1)
sedimentary analyses that included the determination of carbonate and organic carbon contents, pH
values, grain size distributions, and environmental magnetic properties, and (2) for radiocarbon dating.

Sedimentary analyses

Paleosols indicate past periods of stable geomorphological conditions without significant fluvial
sedimentation. To distinguish in situ formed paleosols from allochthonously formed soil sediments
that often have a similar appearance, vertical patterns of sedimentological proxies through possible
paleosol horizons were used: It is assumed that systematically decreasing carbonate and pH values and
increasing values of total organic carbon (TOC) and mass-specific magnetic susceptibility (χ) from bottom to top indicate in situ pedogenesis (Zielhofer et al., 2009). In total, 49 samples were analyzed.

**Carbonate contents** were measured following Scheibler in an Eijkelkamp Calcimeter apparatus, that is, based on the CO₂ volume produced during the reaction of the sediments with 10% HCl. TOC was determined by measuring C_{total} with a Vario EL cube elemental analyzer, and subtracting inorganic carbon (taken from carbonate measurements) from C_{total}. pH values were measured in a 0.1 M KCl suspension using a pH-meter 196 (WTW) after 2 h of soaking. Mass-specific magnetic susceptibility (χ) was measured using a Bartington MS3 magnetic susceptibility meter equipped with a MS2B dual frequency sensor. First, volume magnetic susceptibility (κ) was measured with a frequency of 4.65 kHz, and χ was obtained by relating κ to the mass of the sample. Furthermore, to investigate sedimentological changes throughout the section, the grain size was measured using a Malvern Mastersizer S. Prior to measurement, carbonate was removed using 10% and 30% HCl and organic matter using 10% and 35% H₂O₂, respectively. Subsequently, the samples were dispersed in sodium pyrophosphate solution (Na₄P₂O₇) followed by ultrasonic treatment for 45 min. Given the systematic underestimation of the clay fraction by the laser method, we chose a clay/silt limit of 6 μm (Konert and Vandenberghe, 1997).

**Chronology**

The chronology of the section is based on radiocarbon datings of nine charcoal pieces: Seven samples were taken from fluvial sediments and two samples from archaeological contexts in the upper parts of paleosols Ahb1 and Ahb6 (stratigraphic positions see Figure 3). Radiocarbon measurements were performed in the radiocarbon dating laboratories of Mannheim/Germany and Glasgow/UK using the AMS technique. The respective ages were calibrated using Calpal_A (applying the Intcal13-curve; Reimer et al., 2013). In floodplains, fluvial sediments are deposited during periods with increased flooding activity, and soils are formed during periods of decreased flooding and/or incision of the river bed (Faust et al., 2004; Zielhofer et al., 2008). Accordingly, a relative Soil Development Index (SDI) was furthermore applied to the (paleo) soils to support the radiocarbon-based age model. Unlike other similar indices that only use relative field parameters (e.g. Birkeland, 1999), our index includes the values of the four independently measured pedogenetic proxies carbonate, pH, TOC, and χ: The differences of the analytical values of these proxies between the upper part of a soil and its underlying parent material were normalized for each proxy between 0 and 1, and the SDI for every soil was subsequently calculated using the formula SDI = ((−ΔpH) + (−Δcarbonate) + ΔTOC + ΔΔχ)/4. The intensity of soil formation depends on topography, parent material, climate, organisms, and time (Jenny, 1994): The factor topography can be regarded as equal for all (paleo)soils of the investigated fluvial sediment sequence. Likewise, despite smaller differences also the parent material was similar for all (paleo)soils, showing loamy to clayey grain sizes (50–85% clay) and carbonate contents between 20% and 40% (Figure 3). Reconstructed paleoprecipitation in Georgia indicates an average of up to 180 mm additional annual precipitation compared with today during some periods of the Holocene (Connor and Kvavadze, 2008). Accordingly, precipitation at the investigated site should have increased from ca. 720 to ca. 900 mm/a, what should not fundamentally have changed the local soil moisture regime. Furthermore, biomarker analyses (unpublished results by M. Bliedtner, Bern) indicate that the local vegetation at the investigated site was dominated by natural and anthropogenic grassland throughout the Holocene. Thus, the most important factor that influenced Holocene soil formation intensity must have been its duration. Accordingly, the duration of the recent soil forming period (Ah) served as a reference to calculate approximate durations of soil formation for the paleosols.

**Statistical analysis of recent hydroclimatic data**

To better understand general hydroclimatic properties of the Alazani catchment, we statistically analyzed regional climate data from the station Tbilisi located ca. 40 km south (41°41′N, 44°57′E; data retrieved from the National Centers for Environmental Information, n.d.), and discharge data of the
Alazani River for the station Chiaura located ca. 70 km downstream from the investigated site (41°40′N, 46°04′E; data retrieved from the NCAR/UCAR Research Data Archive and the National Environmental Agency of Georgia) for the period 1926–1992 (missing discharge data for 1929–1932, 1937, and 1945; missing precipitation data for 1930). We performed correlation analyses using the Spearman correlation coefficient (rho) of first differences. First differences are used to counter effects of auto-correlated time series. The significance of the trends is tested by verifying against re-sampled (bootstrapped) surrogate datasets with the same spectral (frequency) properties and a similar autocorrelation structure, using methods described and implemented in Baddouh et al. (2016), Ebisuzaki (1997), and Meyers (2014). The amount of data is re-drawn from the original dataset with re-drawing. The R-script of the calculations is documented in the Supporting Online Material SOM-1, and the data used for the statistical analyses in SOM-2.

Results

Stratigraphy supported by sedimentary analyses

The investigated section mostly consists of fine-grained loamy to clayey overbank sediments (generally >40% clay; Figure 3) that are intercalated with six blackish-grayish to reddish in situ paleosols. A well-developed soil (Ah) is present at today’s surface. Between paleosols Ahb5 and Ahb4, Ahb2 and Ahb1, as well as between paleosol Ahb1 and the recent soil Ah, the sediments are partly sandy and contain singular pebbles (Figure 3). All paleosols could be followed over distances >100 m along the outcrop, showing their extent and significance. The sedimentological analyses revealed that the upper parts of the (paleo)soils are characterized by reduced carbonate and pH values and enhanced TOC and χ values compared with their underlying parent materials (Figure 3). Three intensively developed paleosols (Ahb1, Ahb5, Ahb6) were characterized by distinct upper but gradual lower limits. Three weaker developed paleosols (Ahb2, Ahb3, Ahb4) showed only gradual upper and lower limits. Prehistoric artifacts such as potsherds, bones, or obsidian tools were found in the upper part of Ahb6, between Ahb5 and Ahb4, within Ahb4, and in the upper part of Ahb1. A detailed stratigraphic description is given in Table 1.

Chronology

All radiocarbon datings of charcoal pieces gave Holocene ages, ranging between 10.34 ± 0.07 and 1.68 ± 0.04 cal. ka BP (Figure 3). Most ages are in the stratigraphic order; however, samples MAMS-29341 (385 cm depth; 6.33 ± 0.04 cal. ka BP) and MAMS-29339 (190 cm depth; 10.34 ± 0.07 cal. ka BP) gave ages that are significantly older than samples from lower stratigraphical units and are thus older than the real deposition age. Details of the radiocarbon datings are given in Table 2. According to the radiocarbon date from the uppermost sediment layer below the recent soil (Ah), the maximum duration to form this well-developed soil was estimated to be ca. 1.6 ka (Figure 3). Consequently, the calculated approximate durations of soil formation were ca. 1.2 ka for Ahb1, ca. 0.05 ka for Ahb2, ca. 0.7 ka for Ahb3, ca. 0.4 ka for Ahb4, ca. 1 ka for Ahb5, and ca. 1.7 ka for Ahb6. Although only being approximate estimations, apart from Ahb6 where no radiocarbon age was determined from underlying fluvial sediments, all SDI-based durations of soil formation fit well between the non-overestimated radiocarbon ages, that is, there are no soil formation periods that are significantly longer than the time between their bracketing radiocarbon ages from fluvial sediment layers (Figure 4a).

Statistical analysis of recent hydroclimatic data

The spring discharge maximum of the Alazani River between April and June shows a negative correlation with preceding winter temperature (January–March) of rho = −0.20. Preceding winter temperature shows a negative correlation with winter precipitation of rho = −0.34. Preceding winter precipitation has a positive correlation with spring discharge of rho = 0.35, and the ratio winter precipitation/winter discharge has a negative correlation with winter temperature of rho = −0.25 (Figure 5). All trends are significant with >95% confidence using the described method.
### Discussion

**Chronostratigraphy**

The two radiocarbon ages of charcoal pieces that were taken from archaeological contexts in the upper parts of paleosols Ahb1 (SUERC-28293/GU26198) and Ahb6 (MAMS-17442) most probably directly date human activity on these paleosurfaces and thus fall within the periods of soil formation (Figure 3). In contrast, the radiocarbon ages of charcoal pieces taken from the fluvial sediment layers could potentially overestimate their burial age in case that the charcoal was reworked after longer storage in the catchment (Lang and Hönscheid, 1999). For the two overestimated samples MAMS-29341 and MAMS-29339, this must have been the case, and thus these were discarded from any further interpretation. Given that apart from these two ages all other ages are in the stratigraphical order, the remaining five ages from fluvial sediment layers were assumed to be not significantly overestimated. In the following, the chronostratigraphy is discussed in detail (Figure 4a): The oldest radiocarbon age of 8.1 ± 0.05 cal. ka BP from an archaeological context in the upper part of Ahb6 probably falls into the formation period of that paleosol. An age of 7.34 ± 0.05 cal. ka BP was obtained from fluvial sediments between paleosols Ahb6 and Ahb5, and an age of 5.35 ± 0.07 cal. ka BP from fluvial sediments between paleosols Ahb5 and Ahb4. According to the SDI, the intercalated paleosol Ahb5 developed during ca. 1 ka. Given that the development of paleosol Ahb6 could theoretically have lasted until 7.34 cal. ka BP and the onset of development of Ahb5 could not be determined, the exact duration of fluvial sedimentation between Ahb6 and Ahb5 around 7.34 cal. ka BP is not known. Taken together, according to the SDI paleosols, Ahb4 and Ahb3 developed during ca. 1 ka. Both paleosols are sandwiched between the radiocarbon ages of 5.35 ± 0.07 cal. ka BP and 3.80 ± 0.05 cal. ka BP, showing a temporal separation of ca. 1.5 ka. On one hand, this indicates that fluvial sedimentation between Ahb5 and Ahb4 around 5.35 cal. ka BP could not have been significantly longer than that date although it might have started earlier. On the other hand, fluvial sedimentation between Ahb4 and Ahb3 must have been of quite short duration and was thus of minor importance. An age of 3.80 ± 0.05 cal. ka BP was obtained from fluvial sediments above Ahb3, and an age of 2.91 ± 0.04 cal. ka BP from fluvial sediments just below paleosol Ahb1. Given that the intercalated paleosol Ahb2 had only developed during some decades, it is interpreted that these sediments originate from one main fluvial sedimentation phase that was only shortly interrupted by a period of floodplain stability. The age of 3.8 cal. ka BP indicates the start of this sedimentation phase (see above). Overlying paleosol Ahb1 developed during ca. 1 ka what fits well between the underlying fluvial sedimentation age of 2.91 ± 0.04 cal. ka BP and the overlying fluvial sedimentation age of 1.68 ± 0.04 cal. ka BP. Consequently, fluvial sedimentation between Ahb3 and Ahb1 occurred between ca. 3.8 and 2.9 cal. ka BP. An age of 2.55 ± 0.12 cal. ka BP was obtained from an archaeological context in the upper part of Ahb1 and fits well into the time of its formation. The start of fluvial sedimentation following Ahb1 should have developed during 1.7 cal. ka BP. However, given that the fluvial archive became inactive shortly after this date (see below), the end of this phase could not be determined.

**Deciphering the fluvial archive**

Mostly fine-grained overbank deposits were deposited at the investigated site during the Holocene. The base of these fluvial sediments is formed by paleosol Ahb6 that overprints the upper part of clayey to silty fine-grained overbank deposits. According to the SDI, this soil should have developed during >1.5 ka. However, the SDI is only applicable for relatively stable climatic phases such as most of the Holocene (Jenny, 1994; Zielhofer et al., 2009). Thus, given that the development of Ahb6 already started prior to ca. 9 cal. ka BP when regional annual precipitation was significantly lower compared with younger periods (Connor and Kvavadze, 2008; Joannin et al., 2014), soil development should have been significantly longer than the approximate duration calculated by the SDI. Similar to other Northern Hemisphere rivers (Bridgland and Westaway, 2008; Hetzel et al., 2006; Howard et al., 2004), the river network in the southern Caucasus was strongly incised around the Pleistocene/Holocene transition because of a more continuous and sediment-limited discharge regime in a stabilized landscape. This led to abandonment of the generally elevated sedimentation niveaus of the late Pleistocene that often form elevated terraces today (Ollivier et al., 2015, 2016; von Suchodoletz et
activity, (4) internal river processes, or (5) depend on several by stable periods with soil formation in the floodplain between ca. 3.8 and 2.9, and Holocene periods of fluvial aggradation occurred along the upper Alazani River and possible causes such as (1) sea level fluctuations, (2) human activity, (3) tectonic activity, (4) internal river processes, or (5) climalight changes (Bridgland and Westaway, 2008; Erkens et al., 2009; Leopold et al., 1964):

1. The Alazani River breaks through the Kura-FTB in a narrow gorge ca. 160 km downstream from the studied site. Thus, our sedimentary section is hydraulically decoupled from the Caspian Sea so that sea-level fluctuations can clearly be ruled out as a cause for the observed changes of the fluvial dynamics.

After ca. 1.6 cal. ka BP, the river channel must have been strongly incised again since no fluvial sediments were deposited on the floodplain surface. Instead, the recent soil Ah was formed. Thus, since that time, the fluvial sediment archive had lost its ability to record flood periods (Figure 6d). Ca. 5–9 km west of the investigated section, the current landscape surface exposes former river channels extending up to 1.5 km north of the recent Alazani river bed (Figure 2). This suggests a significant southward shift of the river channel during the past. Accordingly, given that the river bed was located in some distance from the investigated section prior to ca. 1.6 cal. ka BP but is located next to the section today, this indicates a shift of the river bed from a position at least some 100 m further in the north toward its recent southern position since that date (Figure 6d). Looking at the possible causes for the observed changes, recent earthquakes in the vicinity of the study area, high steepness indices (ksn), and high local relief in the uppermost Alazani Basin (Figure 7) in conjunction with GPS data provide evidence for recent deformation and surface uplift resulting from the continued convergence between Arabia and Eurasia (Reilinger et al., 2006). Thus, southward river channel migration and incision could have been caused by southward propagation and related surface uplift of the Greater Caucasus and the Kura-FTB, leading to a southward translation, deformation, and uplift of the intervening Alazani thrust top basin. Southward advancement of the Greater Caucasus in the north and uplift in the Kura-FTB in the south will increasingly confine the course of the Alazani River, leading to its southward shift. Locally, this process may be counteracted by sediment-flux steering (von Suchodoletz et al., 2016) of sediments derived from the Kura-FTB. GPS data indicate shortening rates along strike of the southern margin of the Greater Caucasus from 4 mm/a in the west to 14 mm/a in the east (Forte et al., 2010) which is mirrored in increasing surface uplift rates toward the east (Forte et al., 2016). This surface uplift gradient might have led to a wave of incision (Holbrook and Schumm, 1999; Yanites et al., 2010).

**The Holocene fluvial dynamics of the upper Alazani River and possible causes**

Holocene periods of fluvial aggradation occurred along the upper Alazani River around 7.3 and 5.4, between ca. 3.8 and 2.9, and around 1.7 cal. ka BP. These phases of fluvial activity were interrupted by stable periods with soil formation in the floodplain (Figure 4a). The fluvial dynamics of a river depend on several possible factors such as (1) sea level fluctuations, (2) human activity, (3) tectonic activity, (4) internal river processes, or (5) climatic changes (Bridgland and Westaway, 2008; Erkens et al., 2009; Leopold et al., 1964):

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1. The Alazani River breaks through the Kura-FTB in a narrow gorge ca. 160 km downstream from the studied site. Thus, our sedimentary section is hydraulically decoupled from the Caspian Sea so that sea-level fluctuations can clearly be ruled out as a cause for the observed changes of the fluvial dynamics.
2. As can be seen by Neolithic artifacts on the surface of Ahb6, the studied site was settled since ca. 8 cal. ka BP (Figure 3). Thus, an influence of human activity on the fluvial dynamics cannot be ruled out. However, only parts of the high-mountaint catchment in the Greater Caucasus were settled since the Kura-Araxes culture from ca. 4.5 cal. ka BP (Akhundov, 2004), whereas most parts of the catchment are not settled even today. Furthermore, with >1000 km², the sub-catchment of the studied site is so large that (local) human activity should only have had a minor influence on the fluvial dynamics recorded in our archive (Hoffmann et al., 2007).

3. The region between Greater Caucasus and Kura-FTB is tectonically very active (Figure 7), and such processes obviously significantly influenced the working of the fluvial sediment archive by a southward migration of the Alazani river course after ca. 1.6 cal. ka BP (see above). Thus, further tectonic effects on the Holocene pattern of fluvial aggradation and floodplain stability cannot be ruled out.

4. Given that only one sedimentary section was studied during our investigations, an influence of the Bond events on the fluvial dynamics of the Alazani River recorded at the studied site cannot be excluded. However, given that an outcrop with a lateral extent of 250 m was studied and most overbank sheets show a large lateral extent, only large-scale internal river processes could have influenced our sedimentary record.

5. Looking at possible climatic triggers, it appears that our observed phases of fluvial aggradation are similar with Bond events throughout the investigated Holocene period (Figure 4b). Thus, although also tectonic activity and large-scale internal river processes could have influenced our sediment record, this would argue for a dominant climatic trigger of the large-scale deposition of mostly fine-grained overbank sediments on the upper Alazani floodplain during the Holocene. The Holocene Bond events are characterized by peaks of ice rafted debris (IRD) in the North Atlantic that result from the south- and eastward advection of cold ice-bearing surface waters from the Nordic and Labrador Seas (Bond et al., 1997, 2001). Based on a compilation of several temperature reconstructions from North America and Eurasia (Wanner et al., 2015; Figure 4c), it seems that the Bond events were mostly characterized by a relative decrease of annual temperatures in the northern hemisphere. The influence of the Bond events on the hydroclimatic conditions varied between different periods and regions (Wanner et al., 2011) but the evidences for dryness prevail: For example, from Ireland (Turney et al., 2005), the European Alps (Mangini et al., 2007), southeastern Arabia (Parker et al., 2006), and the eastern Mediterranean Basin (Bar-Matthews and Ayalon, 2011), generally drier conditions are reported during Bond events, and also the Indian summer monsoon was weakened during these periods (Gupta et al., 2005). In contrast, middle and late Holocene Bond events were reported as humid from the Spanish Pyrenees (Pélach et al., 2011). Similarly, wet late Holocene Bond events were also reported from the Moroccan Middle Atlas; however, early and middle Holocene Bond events were characterized by dryness in that region (Zielhofer et al., 2017). The Bond events also influenced the hydroclimatic conditions in and near the Caucasus region: Dust input reconstructed from a sediment core that was taken in a bog nearby Lake Neor in northwestern Iran was systematically increased during Bond events and was connected to regional dryness (Sharifi et al., 2015; Figure 4e). Maximal precipitation in northwestern Iran occurs during spring, followed by a secondary maximum during autumn. Likewise, despite its low temporal resolution, it appears that the maxima of a pollen-based millennial-scale master curve of spring precipitation in Georgia are not correlated with Bond events, that is, regional springs were even dry during some of these events (Connor and Kavvadze, 2008, Figure 4g). Thus, large-scale overbank sedimentation on the upper Alazani floodplain during Bond events obviously occurred during periods with generally drier regional spring conditions. Consequently, variations of spring precipitation cannot be the cause of higher spring discharge maxima that most likely were responsible for fluvial aggradation on the floodplain (Figure 1c). Instead, given the generally lowered annual temperatures in the northern hemisphere during Bond events (Wanner et al., 2015; Figure 4c), we assume an influence of the Bond events on the catchment hydrology during winter as the cause for Holocene overbank sedimentation along the upper Alazani River.
To test this hypothesis of an influence of the regional hydroclimatic conditions during winter on the Holocene flood dynamics of the upper Alazani River, we looked at the correlations between regional hydroclimatic data of the 20th century that are shown in Figure 5: The significant negative correlation of the spring discharge maximum of the Alazani River between April and June with preceding winter temperature between January and March indicates that spring discharge is higher when preceding winter temperatures were lower (Figure 5a). Accordingly, the significant negative correlation between preceding winter temperature and winter precipitation indicates that there is more precipitation during colder winters (Figure 5b). Thus, the observed negative correlation between preceding winter temperature and spring discharge during the 20th century is at least partly caused by higher winter precipitation (Figure 5c). However, looking into the past, the picture of regional winter precipitation is less clear: Whereas Bar-Matthews and Ayalon (2011) observed drier conditions during Bond event 4 in a speleothem in the eastern Mediterranean Basin where winter rain dominates (Figure 4f), based on a speleothem from NE-Turkey covering the last 500 years, Jex et al. (2011) inferred higher regional winter precipitation (October–January) during large parts of Bond event 0 (Figure 4g). Thus, it is not clear whether there was higher winter precipitation in the region throughout the Holocene. Looking further into the recent hydroclimatic data, the significant negative correlation between the ratio winter precipitation/winter discharge and winter temperature suggests that during colder winters, a smaller relative share of winter precipitation directly contributes to winter river discharge because of prolonged snow accumulation (Figure 5d). Instead, this additional water contributes to higher flood peaks during following springs. Thus, similar to Pèlachs et al. (2011) for the Pyrenees, we propose that the observed link between periods of Holocene overbank sedimentation along the upper Alazani River and North Atlantic Bond events may at least partly be caused by lowered winter temperatures during the Bond events.

Consequently, given that annual temperatures in the Caucasus region showed an increasing trend during the last years because of the recent global warming (Keggenhoff et al., 2014; Mamedov et al., 2009), our study suggests that flood maxima of pluvio-nival rivers in the southern Greater Caucasus during spring might have lower magnitudes during the next decades. Likewise, similar observations were also made by recent hydrological studies for Scandinavia (Arnheimer and Lindström, 2015; Vormoor et al., 2015) and California (Dettinger and Cayan, 1995).

Conclusion

This study is the first that documents the Holocene fluvial dynamics of a river originating from the Greater Caucasus Mountains. Our investigation shows that despite documented influences of tectonic activity and possible effects of large-scale internal river processes, Holocene overbank sedimentation along the upper Alazani River was concomitant with North Atlantic Bond events. This indicates a dominant climatic control of the fluvial dynamics recorded in our fluvial sediment archive. Although characterized by a discharge maximum during spring, Holocene overbank sedimentation along the upper Alazani River also occurred during periods with low regional spring precipitation. Thus, the fluvial dynamics of the upper Alazani River must have been controlled by the hydroclimatic conditions during winter. Recent climate and discharge data show a significant negative correlation between preceding winter temperature and spring discharge of the Alazani River, and a positive correlation between winter temperature and winter precipitation. Thus, the influence of the Bond events on the flood dynamics of the upper Alazani River could be explained with more intensive precipitation during cold winters. However, regional data of Holocene winter precipitation are rather contradictory so that it is not clear whether there was more regional winter precipitation throughout the Holocene. Another possible effect of the Bond events on regional hydroclimatic conditions during winter could have been a prolonged snow season during such years. This would have led to a larger volume of spring meltwater that contributes to a more intensive discharge peak during following spring. A similar link was also found in recent hydroclimatic data. Thus, the established link between North Atlantic Bond events and flooding along the upper Alazani River is proposed to have at least partly been caused by lower winter temperatures and a resulting longer retention of the snow cover until spring during those periods. Consequently, given increasing annual temperatures in the region
because of human-caused global warming, flood maxima of pluvio-nival rivers in the southern Greater Caucasus may be expected to decrease during the next decades.

Our study demonstrates the influence of North Atlantic Bond events on flood phases in the southern Greater Caucasus, and thus the high sensitivity of the hydroclimatology of this mountain range toward global climate variations. The contrasting Holocene flood pattern of the upper Alazani River compared with regional humidity phases can be explained with the pluvio-nival runoff regime of that river. This demonstrates that Holocene flood patterns of rivers in mountain regions with complex runoff regimes can only be reconstructed based on the investigation of geological archives, and cannot be derived solely based on regional paleoclimatic data. Understanding the former flood dynamics of such rivers is further complicated by the fact that the timing of runoff maxima could have changed because of changing runoff regimes during the past.

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- NCAR/UCAR Research Data Archive (n.d.): Available at: https://rda.ucar.edu/.


## Table 1

Stratigraphic details of the investigated sediment-paleosol sequence.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Sediment (S)/Paleosol (P)</th>
<th>Grain size</th>
<th>Color</th>
<th>Aggregation (subangular to angular blocky)</th>
<th>Particularities</th>
</tr>
</thead>
<tbody>
<tr>
<td>0–45</td>
<td>P (Ah)</td>
<td>Sandy loam</td>
<td>Blackish-grayish</td>
<td>++</td>
<td>Singular pebbles</td>
</tr>
<tr>
<td>45–215</td>
<td>S</td>
<td>Partly layered: clayey, loamy, and sandy areas</td>
<td>According to layering ocherous-yellowish, brownish, grayish, or reddish</td>
<td>+</td>
<td>Some gravel layers and gravelly channels; singular intact snail shells, charcoal pieces, and carbonatic pseudomycelia</td>
</tr>
<tr>
<td>215–250</td>
<td>P (Ahb1)</td>
<td>Loamy clay</td>
<td>Grayish-reddish</td>
<td>++</td>
<td>Singular carbonatic pseudomycelia; intact and broken snail shells, charcoal pieces, potsherds, and singular bones in upper part</td>
</tr>
<tr>
<td>250–285</td>
<td>S</td>
<td>Loam</td>
<td>Ocherous-reddish</td>
<td>+</td>
<td>Singular Fe and carbonate stains</td>
</tr>
<tr>
<td>285–305</td>
<td>P (Ahb2)</td>
<td>Loamy clay</td>
<td>Reddish-grayish</td>
<td>++</td>
<td>Singular intact snail shells and carbonatic pseudomycelia</td>
</tr>
<tr>
<td>305–365</td>
<td>S</td>
<td>Clayey loam</td>
<td>Ocherous-yellowish to grayish, partly reddish</td>
<td>+</td>
<td>Carbonatic pseudomycelia, singular intact snail shells, and charcoal layer in lower part</td>
</tr>
<tr>
<td>365–385</td>
<td>P (Ahb3)</td>
<td>Silty clay</td>
<td>Grayish</td>
<td>++</td>
<td>Singular intact snail shells, carbonatic pseudomycelia, and charcoal pieces</td>
</tr>
<tr>
<td>385–400</td>
<td>S</td>
<td>Clayey loam</td>
<td>Ocherous-grayish</td>
<td>+</td>
<td>Carbonatic pseudomyc.</td>
</tr>
<tr>
<td>400–450</td>
<td>P (Ahb4)</td>
<td>Clayey loam</td>
<td>Grayishish</td>
<td>+</td>
<td>Charcoal in upper part</td>
</tr>
<tr>
<td>450–535</td>
<td>S</td>
<td>Slightly clayey loam with sandy and gravelly lenses</td>
<td>Grayish-ocherous to yellowish</td>
<td>+</td>
<td>Singular incised pebbly channels, intact and broken snail shells; singular bones and potsherds; charcoal pieces in lower part</td>
</tr>
<tr>
<td>Layer</td>
<td>Type</td>
<td>Texture</td>
<td>Color</td>
<td>BC</td>
<td>Description</td>
</tr>
<tr>
<td>-------</td>
<td>------</td>
<td>---------</td>
<td>-------</td>
<td>----</td>
<td>-------------</td>
</tr>
<tr>
<td>535–565</td>
<td>P (Ahb5)</td>
<td>Clayey loam</td>
<td>Grayish-ocherous</td>
<td>++</td>
<td>Numerous carbonatic pseudomycelia; singular brownish Fe-stains</td>
</tr>
<tr>
<td>565–620</td>
<td>S</td>
<td>Loam</td>
<td>Yellowish-ocherous</td>
<td>+</td>
<td>Small reddish Fe-stains; singular carbonatic pseudomycelia in upper part</td>
</tr>
<tr>
<td>620–655</td>
<td>P (Ahb6)</td>
<td>Clayey loam</td>
<td>Deep black</td>
<td>+++</td>
<td>Singular pebbles; charcoal pieces, potsherds, obsidian artifacts, and prehistoric fireplace in upper part</td>
</tr>
<tr>
<td>655–800</td>
<td>S</td>
<td>Clayey loam</td>
<td>Blackish-grayish to ocherous-yellowish</td>
<td>++</td>
<td>Reddish Fe-stains; singular carbonate concretions; carbonate pseudomycelia</td>
</tr>
</tbody>
</table>

+++: Strong; ++: medium; +: slightly.
Table 2

Results of radiocarbon datings. Ages with MAMS-numbers were measured in Mannheim, and the age with a SUERC-number in Glasgow.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Sample depth (m)</th>
<th>Material</th>
<th>Stratigraphical context</th>
<th>δ(^{13})C (%)</th>
<th>(^{14})C age (ka BP)</th>
<th>Calibrated (^{14})C age (cal. ka BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MAMS-17443</td>
<td>0.8</td>
<td>Charcoal</td>
<td>Sediment layer</td>
<td>−23.7</td>
<td>1.77 ± 0.02</td>
<td>1.68 ± 0.04</td>
</tr>
<tr>
<td>MAMS-29339</td>
<td>1.9</td>
<td>Charcoal</td>
<td>Sediment layer</td>
<td>−33.2</td>
<td>9.17 ± 0.04</td>
<td>10.34 ± 0.07</td>
</tr>
<tr>
<td>SUERC-38293</td>
<td>2.3</td>
<td>Charcoal from archaeological context</td>
<td>Upper part of paleosol</td>
<td>−25.0</td>
<td>2.46 ± 0.03</td>
<td>2.55 ± 0.12</td>
</tr>
<tr>
<td>GU26198</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MAMS-29340</td>
<td>2.5</td>
<td>Charcoal</td>
<td>Sediment layer</td>
<td>−18.7</td>
<td>2.81 ± 0.03</td>
<td>2.91 ± 0.04</td>
</tr>
<tr>
<td>MAMS-18587</td>
<td>3.5</td>
<td>Charcoal</td>
<td>Sediment layer</td>
<td>−27.7</td>
<td>3.53 ± 0.02</td>
<td>3.80 ± 0.05</td>
</tr>
<tr>
<td>MAMS-29341</td>
<td>3.85</td>
<td>Charcoal</td>
<td>Sediment layer</td>
<td>−27.5</td>
<td>5.53 ± 0.03</td>
<td>6.33 ± 0.04</td>
</tr>
<tr>
<td>MAMS-17444</td>
<td>5.2</td>
<td>Charcoal</td>
<td>Sediment layer</td>
<td>−25.2</td>
<td>4.59 ± 0.02</td>
<td>5.35 ± 0.07</td>
</tr>
<tr>
<td>MAMS-29342</td>
<td>5.7</td>
<td>Charcoal</td>
<td>Sediment layer</td>
<td>−21.9</td>
<td>6.40 ± 0.03</td>
<td>7.34 ± 0.05</td>
</tr>
<tr>
<td>MAMS-17442</td>
<td>6.4</td>
<td>Charcoal from archaeological context</td>
<td>Upper part of paleosol</td>
<td>−23.9</td>
<td>7.28 ± 0.03</td>
<td>8.1 ± 0.05</td>
</tr>
</tbody>
</table>
Figures

Figure 1

(a) Location of the study area. The Greater Caucasus is marked with a dashed line, Kura and Alazani River (a) with blue lines, and the study area with a red rectangle. Furthermore, the location of regional paleoenvironmental studies discussed in this paper is shown: 1 = Sharifi et al. (2015), 2 = Connor and Kvavadze (2008), 3 = Bar-Matthews and Ayalon (2011), 4 = Jex et al. (2011). (b) The sub-catchment of the investigated site (Kura-FTB: Kura fold-and-thrust belt). (c) Climate diagram of the climate station Akhmeta (data retrieved from Climate-data.org, n.d.) overlain with monthly mean discharge for the station Chiatura ca. 70 km downstream of the investigated site (data retrieved from the NCAR/UCAR Research Data Archive, n.d.).

DEM-source: ALOS Science Project, Earth Observation Research Center (EORC), Japan Aerospace Exploration Agency (JAXA).
Figure 2
Detailed map of the surroundings of the studied site, with photos of exposed former river channels north of the recent river course (left) and the Alazani River with the investigated section in the background (right). The lines of sight are indicated with white arrows.

DEM-source: ALOS Science Project, Earth Observation Research Center (EORC), Japan Aerospace Exploration Agency (JAXA).
Figure 3
Stratigraphic sketch and photo of the investigated site with calibrated radiocarbon ages, sedimentological proxies, and the Soil Development Index (SDI) with derived approximate durations of soil formation.
Figure 4

The fluvial dynamics of the upper Alazani River (a) compared with regional and global paleoenvironmental data. (b) Bond events (Bond et al., 2001); (c) average land temperature in the northern hemisphere (Wanner et al., 2015); (d) dust input in northern Iran (Sharifi et al., 2015); (e) average precipitation anomaly in Georgia (Connor and Kvavadze, 2008); (f) winter precipitation in Israel (Bar-Matthews and Ayalon, 2011); (g) winter precipitation variations in northeastern Turkey (Jex et al., 2011).
Figure 5
Correlation of recent climate data from the station Tbilisi ca. 40 km to the south (41°41′N, 44°57′E; data retrieved from the National Centers for Environmental Information, n.d.), and discharge data of the Alazani River for the station Chiaura located ca. 70 km downstream from the investigated site (41°40′N, 46°04′E; data retrieved from the NCAR/UCAR Research Data Archive and National Environmental Agency of Georgia) for the period 1926–1992. All correlations are significant at >95% confidence. The data used for these correlations can be found in Supporting Online Material SOM-2: (a) Rho = −0.20 (n = 61), (b) Rho = −0.34 (n = 64), (c) Rho = 0.35 (n = 61), and (d) Rho = −0.25 (n = 61).
Figure 6

Schematic representation of phases of working of the fluvial sediment archive: (a) archive not working because of intensive incision around the Pleistocene/Holocene transition when floods could not reach the abandoned floodplain surface; (b and c) archive working between ca. 8 and 1.6 cal. ka BP when the river bed was so much aggraded that floods could reach the floodplain surface; (d) archive not working after ca. 1.6 cal. ka BP following incision and southward shift of the river bed when floods could not reach the then abandoned surface.
Local relief, steepness index ($ksn$) sensu Kirby and Whipple (2012) and earthquakes since 1978, highlighting areas of deformation and surface uplift. Local relief is calculated as the difference between maximum and minimum elevations measured over a 1-km-diameter moving window.

DEM analysis was carried out using the topotoolbox-software (https://topotoolbox.wordpress.com) and is based on an SRTM-DEM with a resolution of 90 m. A concavity index of 0.45 was used for normalization of steepness indices. Earthquake data were compiled from two regional earthquake catalogues (‘General Catalogue of Earthquakes in North Eurasia (GNRL)’; http://earthquake.usgs.gov/data/russia_seismicity/sourcecatalogs/general.php; ‘Composed Regional Catalogue – Caucasus (CRC)’; http://earthquake.usgs.gov/data/russia_seismicity/regionalcatalogs/caucasus.php), as well as from two global catalogues (‘Havard CMT Catalogue: http://www.globalcmt.org; Seismic Information Service of the GFZ Potsdam’: http://geofon.gfz-potsdam.de/eqinfo/form.php).