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**Linking Danube River Activity to Alpine Ice-Sheet Fluctuations during the Last Glacial (*ca.* 33-17 ka BP): insights into the continental signature of Heinrich Stadials**

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## ABSTRACT

Offshore archives retrieved from marine/lacustrine environments receiving sediment from large river systems are valuable Quaternary continental records. In the present study, we reconstruct the Danube River activity at the end of the last glacial period based on sedimentological, mineralogical and geochemical analyses performed on long-piston cores from the north-west Black Sea margin. Our data suggest that the Danube River produced hyperpycnal floods throughout the *ca.* 33-17 ka period. Four main periods of enhanced Danube flood frequency, each of 1.5-3 kyr duration, are recorded at *ca.* 32.5-30.5 ka (equivalent to the first part of Heinrich Stadial -HS- 3), at *ca.* 29-27.5 ka (equivalent to Greenland Stadial 4), at *ca.* 25.3-23.8 ka (equivalent to HS 2) and at *ca.* 22.3-19 ka. Based on mineralogical and geochemical data, we relate these events to enhanced surface melting of the Alpine Ice Sheet (AIS) that covered ~50,000 km<sup>2</sup> of the Danube watershed at the Last Glacial Maximum (LGM). Our results suggest that (i) the AIS growth from the inner Alps to its LGM position in the northern Alpine foreland started from *ca.* 30.5 ka, ended no later than *ca.* 25.3 ka, and was interrupted by a melting episode *ca.* 29-27.5 ka; (ii) the AIS volume drastically decreased from *ca.* 22.3 to 19 ka, as soon as summer insolation energy at the AIS latitude increased; and (iii) HSs strongly impacted the AIS mass balance through enhanced summer surface melt. This, together with evidence of severely cool winters and the rapid expansion of sea ice in the North Atlantic, implies strong seasonality in continental Europe during stadials.

Keywords: Danube River, floods, hyperpycnites, Alpine Ice Sheet, Heinrich Stadials, seasonality, Black Sea.

## 1. INTRODUCTION

Fluvial landforms and sediments can be used to reconstruct past hydrological conditions since rivers are highly sensitive to environmental changes (Blum and Törnqvist, 2000; Macklin et al., 2012). Prominent examples from the British Isles (Macklin et al., 2005; Macklin and Lewin, 2003) and continental Europe (Busschers et al., 2007; Gábris and Nádor, 2007; Gibbard, 1988; Kasse et al., 2010; Macklin et al., 2006; Mol et al., 2000; Pastre et al., 2003; Starkel et al., 2015; Vandenberghe, 2002) point to the impact of Quaternary climate change on river system dynamics. However, the riverine archives are highly complex (*e.g.* spatially-dependant types and rates of sedimentation, hiatus and erosion, difficulties in dating; Blum and Törnqvist, 2000; Coulthard et al., 2005; Macklin and Lewin, 2003; Vandenberghe, 2003). Hence the use of sediment archives retrieved offshore of large river systems can provide complementary high-frequency sediment flux records resulting from river discharge and climate variability (Simpson and Castelltort, 2012). This has been demonstrated in various geographic and climatic settings (*e.g.* Bonneau et al., 2014; Covault et al., 2011; Goodbred and Kuehl, 2000; Revel et al., 2010), including the giant glaciated catchments of European rivers, *i.e.* the Channel River (Ménot et al., 2006; Toucanne et al., 2009; Zaragosi et al., 2001) and the Dniepr River (Soulet et al., 2013). These studies provided remarkable new insights into the paleohydrology of Europe during past glacial intervals as well as into the impact of the Fennoscandian Ice Sheet (FIS) in amplifying, pacing and driving regional and global climate changes (*e.g.* Boswell et al., 2019; Soulet et al., 2013; Toucanne et al., 2015).

It is well-known that the Black Sea received vast amounts of sediment and meltwater from the Alpine Ice Sheet (AIS) and FIS during glacial periods (Bahr et al., 2006; Gorlach et al., 2017; Grosswald, 1980; Major et al., 2006; Mangerud et al., 2004; Soulet et al., 2013, 2011a; Tudryn et al., 2016). The Red Layers (*e.g.* Bahr et al., 2006), deposited in the north-west Black Sea by the Dniepr River and the retreating FIS between *ca.* 17.2 and  $15.7 \pm 0.3$  cal ka BP (Soulet et al., 2013), constitute a prominent example further highlighting the ability for ice-sheets to impact sediment routing systems down to the deep sea ‘sink’ (Jaeger and Koppes, 2016). By comparison, the impact of AIS fluctuation on the Danube River discharge and Black Sea sedimentary system has received much less attention. Yet, the entire north-east part of the AIS (~50,000 km<sup>2</sup>) drained into the Black Sea and sediment accumulated in the Danube delta (Giosan et al., 2006, 2005; Panin et al., 1983) and deep-sea fan (Constantinescu et al., 2015; Lericolais et al., 2013; Popescu et al., 2001; Wong et al., 1994) must have recorded Danube River activity and AIS fluctuations during the late Quaternary. Here, we



provide a continuous, high-resolution record of the Danube River discharge into the Black Sea ‘Lake’ at the end of the last glacial period (from *ca.* 33.5 ka) through the study of two sediment cores retrieved midway between the Danube delta and the Danube deep-sea fan (Fig. 1). Millennial-scale variability in sediment input and source observed in our multi-proxy records suggests rapid regional-scale glacier fluctuations in the north-eastern Alps over the course of the AIS to its maximum extent and during the deglaciation.

## 2. GEOLOGICAL AND ENVIRONMENTAL SETTINGS

### 2.1 The north-west Black Sea and the Danube sediment routing system

Since its opening during the Upper Cretaceous, the Black Sea basin acts as a large sediment catchment area that accumulated a thick sedimentary cover of ~19 km in its western margin. Since the Pliocene, a 2.5 to 3 km-thick prograding depositional wedge has been formed by Danube and Dniepr sedimentary supplies (de Leeuw et al., 2018; Nikishin et al., 2003). This geological feature, forming the wide north-west Black Sea continental shelf (<140-170 m water depth) and margin, is incised by numerous erosive canyons (Riboulot et al., 2017) including the Danube Canyon which deeply incises the shelf over 26 km down to -110 m water depth (Popescu et al., 2004) (Figs 1 and 2). The Danube Canyon has fed an extensive, deep depositional (turbidite) system (*i.e.* Danube deep-sea fan) at water depths greater than 2000 m (Lericolais et al., 2013; Popescu et al., 2001; Wong et al., 1994). It has been active during the last glacial period and until *ca.* 11.7 ka (Constantinescu et al., 2015) when the Black Sea was a giant freshwater lake and lowstand conditions (~90-120 m below the present sea level) dominated (Deuser, 1972; Lericolais et al., 2011; Ross et al., 1970; Ryan et al., 1997; Yanchilina et al., 2017). The Danube deep-sea fan thus represents the final sink of the Danube sediment routing system (Matenco and Andriessen, 2013) (Fig. 1), *i.e.* the integrated dynamical system connecting erosion in mountain catchments of western Central Europe to downstream deposition in the Black Sea.

The Danube River is the second longest river in Eurasia, with a length of 2,900 km from the Black Forest (Germany) to its delta (Romania) (Fig. 1), a drainage area of 801,000 km<sup>2</sup> and a modern sediment load estimated at 42-67 Mt per year (sediment yield: ~80 t.km<sup>2</sup>.yr<sup>-1</sup>). Today, it constitutes by far the main sediment source for the northern Black Sea considering the low sediment load, estimated to be less than 10 Mt per year in total, for the rivers Dniestr (sediment yield: ~30 t.km<sup>2</sup>.yr<sup>-1</sup>), Dniepr and Don (sediment yield: ≤5 t.km<sup>2</sup>.yr<sup>-1</sup>). This difference is due to the mountainous character of the Danube drainage area in comparison to

the lowlands-dominated Dniestr, Dniepr and Don watersheds (Milliman and Farnsworth, 2013; Milliman and Syvitski, 1992; Panin and Jipa, 2002; Sommerwerk, 2009). Indeed, the Danube River collects water and sediment from the mountainous areas of the northern and eastern Alps (up to 4,049 m high at Piz Bernina; Fig. 3), the Dinarides and the Carpathians (Fig. 1), mainly formed of Variscan metamorphic and magmatic basements and Mesozoic calcareous rocks (Asch et al., 2005). In detail, from its source to its mouth, the Danube River receives the following: on its right bank, the tributaries Iller, Inn, Enns, Drava and Sava, originating in the Alps and the Dinarides, as well as the Velika Morava flowing from the Balkans; and on its left bank, the major tributaries Morava, rising in the Bohemian Forest, and Tisza, originating in the Carpathians, and the Jiu, Olt, Siret and Prut rivers, with sources in the Carpathians (Garnier et al., 2002) (Fig. 1). In addition to the riverine material from the mountainous areas, the glacial loess deposits constitute a significant additional sediment supply to the Danube sediment load, especially in the Pannonian (Middle Danube) and Dacic (Lower Danube) basins (Marković et al., 2015).

## **2.2 Paleoenvironmental changes in the Danube drainage area**

The loess and glacial sequences preserved in the Danube catchment provide information on the intensity of Alpine glaciations (Marković et al., 2015), with the first major glaciation in the Alps dating from Marine Isotope Stage (MIS) 22 (Muttoni et al., 2003). The (local) Last Glacial Maximum (LGM) in the Alps (Upper Würmian, MIS 2; Chaline and Jerz, 1984, Figs 1 and 3) lasted from not later than *ca.* 24 to 20-19 ka (*e.g.* Ivy-Ochs, 2015; Ivy-Ochs et al., 2008; Monegato et al., 2017), with glaciers extending tens of kilometers out onto forelands and coalescing into huge (Rhine, Isar/Loisach, Inn, Salzach and Traun, from west to east) piedmont lobes (up to 700 m in thickness) in the Molasse Basin, Upper Danube (Ivy-Ochs, 2015; van Husen, 2000) (Fig. 3). This time period is coeval with peak loess accumulation rates in the Pannonian Basin (Fitzsimmons et al., 2012) and the development of the Duttendorf loess sequence in the Salzach palaeoglacier area (Starnberger et al., 2011). Recent compilation of <sup>10</sup>Be surface exposure dating and radiocarbon ages demonstrate that if LGM glaciers vanished rapidly from the foreland at *ca.* 20-18 ka, the initiation of retreat from their maximum positions occurred as soon as *ca.* 24-22 ka (Ivy-Ochs, 2015; Ivy-Ochs et al., 2004; Reber et al., 2014). Note that the timing for the growth and demise of cirque and cirque-valley glaciers in the western and southern Carpathians mountains matches that recognized in the northern Alps (Makos et al., 2018, 2014, 2013; Reuther et al., 2007).

In contrast, little is known about pre-LGM ice fluctuations (Ivy-Ochs et al., 2008). The Middle Würmian (MIS 4 and 3; Chaline and Jerz, 1984) is poorly preserved in the Alps due to the excavation of valleys by the LGM ice advance soon after *ca.* 33-32 ka (*i.e.* onset of the Upper Würmian, as redefined by Spötl et al., 2013). However, recent <sup>10</sup>Be data from the western Alps and the Rivoli-Avigliana end-moraine system (northern Italy) yield a minimum age of  $41.2 \pm 1.9$  ka (MIS 3) for pre-LGM glacier advance(s) in the Pô Valley (Ivy-Ochs et al., 2018). This age fits well with the deterioration of climatic conditions over the Alps after *ca.* 45 ka (Heiri et al., 2014; Starnberger et al., 2013). These results, including recent ones from the Berchtesgaden basin (Germany; Mayr et al., 2019), corroborate numerical simulations suggesting that a highly dynamic ice-sheet occupied the inner-Alpine regions throughout the last glacial period, with possible recurrent short-lived advances onto the forelands well before the LGM (Seguinot et al., 2018).

### 3. MATERIALS

#### 3.1. Sediment cores

This study focuses on the Calypso long-piston cores GAS-CS01 (44°05.1'N, 30°47.5'E) and MD04-2790 (44°12.8'N, 30°59.6'E) retrieved from the north-west Black Sea margin (Figs 1 and 2; Table 1). Core GAS-CS01 (33.4 m long) was collected at 240 m water depth from the shelf margin, ~40 km east of the Danube Canyon head, during the GHASS oceanographic cruise (R/V Pourquoi Pas?; Ker and Riboulot, 2015). Core MD04-2790 (30.3 m long) is located at 352 m water depth, ~19.6 km east of GAS-CS01. This core was collected during the ASSEMBLAGE-1 cruise (R/V Marion Dufresne; Lericolais, 2004) and has been extensively studied over the last few years for chronostratigraphical and paleoenvironmental studies (*e.g.* Soulet et al., 2011a,b; Sanchi et al., 2014). Note that the length of cores GAS-CS01 (this study) and MD04-2790 (Soulet et al., 2011a) has been corrected after removal of coring gaps (corrected length of 32.1 and 27.8 m, respectively). Core depths discussed in this study thus correspond to depth (cm, m) below sea floor (bsf), and corresponding depths (raw - cm- versus -cmbsf-) are presented in Tables 2 and 3.

#### 3.2. Danube River and tributary samples

In order to track paleoenvironmental changes in the Danube sediment routing system, eighteen riverine sediments have been recovered from the Danube River and its tributaries to compare their mineralogy with that of the GAS-CS01 core samples (Fig. 1; Table 3). Fine-

grained river sediments constitute the mineralogical diversity of catchment areas. As such they can provide a reliable average mineralogical and geochemical composition of their corresponding drainage basin (Chamley, 1989; Sionneau et al., 2008). Considering the disproportionate role of mountains and glaciers in the erosion processes and the river sediment yields (Hallet et al., 1996; Milliman and Syvitski, 1992; Comiti et al., 2019), we mainly collected river sediments from tributaries draining the northern Alps (*e.g.* Iller, Inn;  $n=5$ ), the western Carpathians (*e.g.* Tisza, Morava;  $n=5$ ) and the Dinarides (Sava and Velika Morava rivers;  $n=2$ ). Additional samples were collected along the Danube mainstream ( $n=6$ ) from Passau (Germany) to the Black Sea, near the confluence with the Tisza and Velika Morava rivers. At every location where fine sediment was available, samples were collected from fresh bank deposit.

## 4. METHODS

### 4.1 Core-scale analyses

The sedimentological analysis of GAS-CS01 is firstly based on visual description and X-ray images that allow the recognition of the main sedimentary facies and sequences, and the manual counting of laminae/beds (see Part 6.1; Fig. 4). 2D X-ray images were acquired with a Geotek XCT system (Ifremer, Brest) from split core sections. Detailed identification of the sedimentary facies was completed by hyperspectral reconstructed images. The hyperspectral spectroscopy technique consists of combining reflectance spectroscopy and digital imaging. A high-resolution digital image of a sediment core and a reflectance measurement for a specified wavelength range are collected in each pixel of the digital image. The wavelength intervals at which absorption features appear in their reflectance spectra are determined by the chemical compounds (*e.g.* organic matter) and physical properties (*e.g.* grain size) of the sediments (Debret et al., 2018; Jacq et al., 2019; Van Exem et al., 2019). In this study, we present high-resolution reconstructed images (spatial resolution of  $\sim 61 \mu\text{m}/\text{pixel}$ ) in order to highlight the sedimentary features in GAS-CS01. Hyperspectral images were acquired at the University of Rouen (France) using a Specim (Spectral Imaging Ltd.) sediment-core-scanning system in combinations with a Specim PFD-xx-V10E camera (Specim Ltd., Finland) and OLES22.5 lens. These high-resolution images were acquired in the VNIR (Visible and Near InfraRed) spectral range between 400 to 1000 nm. The colour obtained with the VNIR is a false (highlighted) colour constructed after selecting three bands in the RGB system (697.94, 548.87 and 469.59 nm).

The bulk intensity of major elements for core GAS-CS01 was analysed using an Avaatech XRF core-scanner (Ifremer, Brest). XRF data were collected every 1 cm along the entire core, with a count time of 10 s, by setting the voltage to 10 kV (no filter) and 30 kV (Pd thick filter) and the intensity to 600 mA and 1000 mA, respectively. Millimetre-scale analyses were also conducted on some sections (30 s., 10-30 kV, 1000 mA) (Fig. 5).

The global mineralogy composition of GAS-CS01 sediments was determined by X-ray diffractometry (XRD), using a D8 Advance BRUKER model, a device type Bragg-Brentano equipped with a Cu X-ray tube, a primary Soller slit of 0.6 mm, a nine-position sample holder and the VANTEC-1 Position Sensitive Detector with a nickel filter (Ni 0.5). Prior to analysis, dried sediment powders were inserted into the sample holder, and flattened. Measurements were made from 5 to 70° with a resolution of 0.01° lasting 1 s. The voltage and amperage were set up to 40 kV and 30 mA, respectively.

The clay mineralogy composition of GAS-CS01 sediments was determined by X-ray diffractometry (XRD), using a D2 PHASER BRUKER model, a Bragg-Brentano device type equipped with a Cu X-ray tube, a primary Soller slit of 0.6 mm and the LYNXEYE Detector with a nickel filter (Ni 0.5). In order to separate the clays from the rest of the matrix, the sample undergoes several stages of preparation: various chemical treatments to eliminate carbonates, iron oxides as well as the organic matter and a settling. Finally, these clays are plated on an oriented glass slide which is analysed in three different ways: without treatment, after saturation with ethylene glycol and after calcination at 490°C. Measurements were made from 2 to 30° with a resolution of 0.02° lasting 1 s. The voltage and amperage were set to 30 kV and 10 mA, respectively. The illite crystallinity (ICr) index, or Kübler index (Kisch, 1991), was measured from XRD results. The illite crystallinity ( $^{\circ}\Delta 2\theta$ ) is the full width at half maximum (FWHM) determined from the illite peak at 10 Å (Chamley, 1989). The clay mineralogy and the ICr of the riverine samples ( $n=18$ ) were also determined.

Finally, the geographical provenance of the GAS-CS01 sediments was determined measuring Nd isotopic ratios of fine-grained detrital fraction ( $n=14$ ). The neodymium isotopic composition ( $\epsilon_{Nd}$ ) of terrigenous sediment is a powerful tracer for geographical provenance because the  $\epsilon_{Nd}$  signature of detrital sediment is retained during continental weathering and subsequent transport (Bayon et al., 2015; Goldstein and Jacobsen, 1988). We focus on the <63 µm fraction because the meltwaters of ice margins predominantly transport the clay and silt fractions of continental detritus (Boswell et al., 2018; Brown and Kennett, 1998). All bulk samples were first treated to quantitatively remove any biogenic, authigenic and organic compounds (Bayon et al., 2002) and residual detrital fractions were digested by alkaline

fusion prior to isolation of the Nd by ion-exchange chromatography (Bayon et al., 2009). Nd isotope measurements were performed on a Thermo Scientific Neptune MC-ICP-MS at the Pôle Spectrométrie Océan, Brest (France). Mass bias corrections on Nd were made with the exponential law, using  $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ . Nd isotopic compositions were determined using sample-standard bracketing, by analysing an internal standard solution every two samples. Mass-bias corrected values for  $^{143}\text{Nd}/^{144}\text{Nd}$  were normalized to a JNdi-1 value. Replicate analyses of the JNdi-1 standard solution during the course of this study gave  $^{143}\text{Nd}/^{144}\text{Nd} = 0.512113 \pm 0.000013$  (2SD,  $n=28$ ), in full agreement with the reference value (0.512115) of Tanaka et al. (2000), and corresponding to an external reproducibility of  $\sim \pm 0.25$   $\epsilon$ -units, taken as the estimated uncertainty on our measurements. In this study, both measured  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios and literature data are reported in  $\epsilon_{\text{Nd}}$  notation  $[(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}}/(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} - 1] \times 10^4$ , using  $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512638$  (Jacobsen and Wasserburg, 1980). The  $\epsilon_{\text{Nd}}$  dataset available in core MD04-2790 for the *ca.* 28-10 ka period (Soulet et al., 2013) was produced using this procedure.

## 4.2 Laminae-scale analyses

To characterize the sedimentary facies of core GAS-CS01, and the laminations in particular (see Part 6.1), we performed microscopic observations of 8 cm long thin-sections of impregnated sediments. The thin sections were prepared at the Université de Bordeaux (UMR EPOC) according to the protocol described in Zaragosi et al. (2006). The studied intervals were selected from well-preserved and representative sedimentary facies (Fig. 4). The thin sections were then used for geochemical characterization of the sediments via scanning electron microscopy (SEM) at Ifremer, using a Thermo Scientific Quanta 200 equipped with an OXFORD Instrument X-MAXN silicon drift detector energy-dispersive X-Ray spectrometer (detector size: 80 mm<sup>2</sup>). The thin sections were coated with a thin layer of carbon and X-ray EDS maps acquired on an area of about 300 mm<sup>2</sup> to investigate the microscale geochemical distribution/composition of the sediments. Each area is composed of over 30 fields (3.47  $\times$  3.04 mm each) scanned with a resolution of 1024  $\times$  896 pixels (1 pixel = 3.4  $\mu\text{m}$ ). Additionally, morphological analyses of lithic grains were performed through SEM (Quanta 200). The grains were deposited onto SEM sample holders covered with a gold layer, and the SEM images were obtained with the secondary electron detector mode and an accelerating voltage of 5.0 kV.

Finally, we performed a total of 398 grain-size analyses using a Malvern Mastersizer 3000 laser diffraction particle size analyser (0.1-2000  $\mu\text{m}$ ) coupled with a Hydro LV wet dispersant



unit. Ultrasonic dispersion (2 minutes) was conducted before analysis. The average resolution is 0.25 cm along the studied intervals.

## 5. AGE-DEPTH MODEL

Core chronologies based on radiocarbon dates are hampered in the Black Sea by unconstrained reservoir age (Guichard et al., 1993; Jones and Gagnon, 1994; Kwiecien et al., 2008; Soulet et al., 2011a, 2019). To circumvent the reservoir age issue, it is possible to reconstruct a meaningful environmental proxy from the core sediment and align it to a calendrically-dated reference record (*e.g.* Bard et al., 2004, 2013; Heaton et al., 2013; Skinner et al., 2010, 2020; Soulet et al., 2011a) in order to obtain the calendar age-depth model without using radiocarbon dates. Original calendar age-depth model of core MD04-2790 is based on the alignment approach (Soulet et al., 2011a): TEX86-derived Lake Surface Temperature (LST) reconstructed from MD04-2790 sediments was aligned to Hulu Cave  $\delta^{18}\text{O}$  speleothem record (Wang et al., 2001). The GAS-CS01 calendar age-depth model is based on that of core MD04-2790, through the alignment of their respective XRF-Ca records (Fig. 5) which provided calendar ages at tie-points. The striking resemblance of XRF-Ca records confirms that the XRF  $\text{Ca}_{\text{norm}}$  is a basin-wide stratigraphic proxy in the Black Sea (Bahr et al., 2005, 2008; Kwiecien et al., 2008; Soulet et al., 2011a; Constantinescu et al., 2015). Original MD04-2790 age-depth model was performed until to the depth 1990 cmbsf (31.3 ka), corresponding to depth 1595 cmbsf in core GAS-CS01. Below, MD04-2790 LST record did not provide reliable tie-point compared to Hulu Cave record (Soulet et al., 2011a). Hence, below the depth 1990 cmbsf in MD04-2790 and the depth 1595 cmbsf in GAS-CS01, the chronology is based on radiocarbon dates (*Dreissena* sp. and bulk organic matter; Table 2). We implemented these chronological constraints (Supplementary Table S1 and S2) in Oxcal (Bronk Ramsey, 2001, 2008, 2009; Bronk Ramsey and Lee, 2013; Bronk Ramsey, 2017). As reservoir ages are unconstrained for this period, we applied a vague reservoir age correction following a uniform distribution spanning 0 to 5000  $^{14}\text{C}$  yrs correction. Oxcal codes and further details of the modelled age-depth models are available in Supplementary Online Material. We found that the base of both cores consistently dates back to ~33.5 ka (Supplementary Figs. S1 and S2).

## 6. RESULTS

## 6.1 GAS-CS01 lithology

The analysis of core GAS-CS01 shows the classic regional lithological succession consisting of the microlaminated coccolith ooze (MCO; 0-0.04 mbsf), the black-brown microlaminated sapropel (0.04-0.52 mbsf), and the carbonate-rich layers (down to 2.53 mbsf) overlying the Red Layers (2.53-4.26 mbsf). Below the Red Layers down to the bottom (32.1 mbsf), the GAS-CS01 lithology consists of even, parallel clastic rhythmites. This lithological succession, found from the north-west Black Sea upper slope (Bahr et al., 2005; Major et al., 2002; Soulet et al., 2011a) to the Danube deep-sea fan (Constantinescu et al., 2015), corresponds to ‘marine’ Unit I (MCO,  $<2,720 \pm 160$  cal a BP) and Unit II (sapropel, deposited from  $8,080 \pm 250$  cal a BP) and below, ‘lacustrine’ Unit III (glacial ‘banded lutites’) described by Ross and Degens (1974). Consequently, the carbonate-rich silty clay is interpreted as the upper part of the ‘lacustrine’ Unit III (Dean and Arthur, 2011). This sediment deposited during the late glacial (including the Bølling-Allerød Interstadial and the Younger Dryas cold event), from  $15.7 \pm 0.3$  ka BP (upper part of the Red Layers; Soulet et al., 2011a) and the Initial Marine Inflow *ca.* 9 ka (Soulet et al., 2011b). Based on the above, core GAS-CS01 therefore consists of a very high-resolution record of the glacial (*sensu stricto*,  $> 16$  ka) ‘lacustrine’ Unit III (until *ca.* 33.5 ka; see Part 5 ‘Age-depth model’).

In detail, the glacial sequence of GAS-CS01 is fully laminated (total of 3815 laminae/beds revealed by X-ray imagery) and consists of even, parallel clastic rhythmites that resemble the varve pattern of Degens et al. (1978). They consist of silty to sandy graded deposits, evolving gradually from millimetre- (*i.e.* laminae; Scandinavian-type lacustrine varves of Degens et al., 1978) to decimetre-thick deposits (*i.e.* beds; megavarves of Degens et al., 1978) from the Red Layers (2.53-4.26 mbsf) to the core bottom (32.1 mbsf) (Figs 4 and 6). They are separated by muddy intervals of similar thickness (*i.e.* mm- to dm-thick from top to bottom). Interestingly, the grain-size of the deposits is positively correlated with the thickness of the clastic intervals, from silt-sized grains in the Red Layers to sand-sized particles in the deeper part of GAS-CS01. High-resolution grain-size analyses reveal the superposition of (basal) coarsening-up and (top) fining-up sequences throughout GAS-CS01. This pattern is mirrored by the millimetre-scale XRF Zr/Rb, well-known as a semi-quantitative grain-size indicator (Dypvik and Harris, 2001) (Figs 7, 8 and 9). All these results (and the inversely graded facies in particular) suggest that the clastic rhythmites in GAS-CS01 are hyperpycnite sequences, *i.e.* deposits that are related to quasi-steady hyperpycnal (*i.e.* ‘above a density threshold’) turbidity currents (Mulder et al., 2003; Mulder and Alexander, 2001). The detailed observation of X-ray images and thin-sections reveal similar sedimentary deposits (including



4841 laminae) at site MD04-2790 (Soulet et al., 2011a), ~20 km east of GAS-CS01. This highlights the regional imprint of hyperpycnal flows in the north-west Black Sea 'Lake' during the last glacial period.

## 6.2 Hyperpycnal turbidites frequency

The frequency of hyperpycnal turbidite deposits ( $\text{hyp.250 yr}^{-1}$ ) has been estimated for cores GAS-CS01 and MD04-2790 (Fig. 10b and Supplementary Online Material). The two cores present similar trends. A peak in hyperpycnal turbidite frequency ( $>100 \text{ hyp.250 yr}^{-1}$ ) is observed from *ca.* 32.5-30.5 ka with a maximum (up to  $380 \text{ hyp.250 yr}^{-1}$  at GAS-CS01) estimated at *ca.* 31 ka. A rapid decrease is observed thereafter ( $<50 \text{ hyp.250 yr}^{-1}$ ), followed by a new increase centred at 29-27.5 ka ( $\sim 150 \text{ hyp.250 yr}^{-1}$  at MD04-2790). Then, a ~10-kyr-long period (from *ca.* 27 to 17 ka) of moderate hyperpycnal turbidite frequency fluctuating between ~15 and  $50 \text{ hyp.250 yr}^{-1}$  is observed. During this interval, two main episodes of increasing frequency (up to  $\sim 50\text{-}70 \text{ hyp.250 yr}^{-1}$ ) are observed, at *ca.* 25.3-23.8 ka and at *ca.* 22.3-19 ka. A last peak in hyperpycnal turbidite frequency is observed at the two sites at 17.2- 16 ka, *i.e.* during the deposition of the Red Layers. The evolution of sediment accumulation rates calculated for the two cores strictly follows that of the hyperpycnal turbidite frequency (Fig. 10c).

## 6.3 Sedimentary geochemistry

### 6.3.1 Clay mineralogy

The GAS-CS01 clay mineralogy ( $n=146$ ; Fig. 10d,e and Supplementary Online Material) reveals the dominance of both smectite ( $39 \pm 13\%$ ) and illite ( $36 \pm 10\%$ ) over chlorite ( $15 \pm 7\%$ ) and kaolinite ( $11 \pm 5\%$ ). From *ca.* 31 to 17.2 ka, the abundance of clay minerals fluctuates (up to ~15-20% variability for smectite) but is relatively stable in comparison to the  $>31$  ka (peak in hyperpycnal turbidite frequency; see Part 6.2) and  $<17.2$  ka (Red Layers) intervals. Indeed, the oldest interval of GAS-CS01 is characterised by high abundances in illite (~40-45%) and chlorite (~20-30%), but low abundances in smectite (~30%) and kaolinite (~5-10%), in comparison to the subsequent *ca.* 31-17.2 ka interval. Similarly, the smectite content abruptly decreases from *ca.* 17.2 ka ( $<10\%$ ) while the content in kaolinite (up to 25%) and illite (up to 60-70%) significantly increases. Note that despite a general low abundance, kaolinite is the only clay mineral to show a distinct signature for the three time periods considered.

The Danube River tributary samples ( $n=12$ ) were divided into three groups corresponding to the north-eastern Alps ( $n=5$ ), the Carpathians ( $n=5$ ) and the Dinarides ( $n=2$ ) mountainous areas (Table 3). River samples from the Alps show an illite content of ~40-50% while chlorite represents ~25% of the clayey fractions. Kaolinite and smectite contents are low, usually <5%, except in the Iller River where the smectite content reaches 40%. Smectite (~55%) dominates over illite (~25-30%) in the Carpathians, while an inverse pattern is observed in the Dinarides (~45-50% and ~20-25% of illite and smectite, respectively). The clay mineralogy of these zones contrasts with that of the north-eastern Alps with significant abundance of kaolinite (~10-20%) and very low content of dolomite (<2%). Dolomite abundance is significant (up to 17%) in the Alpine rivers.

### 6.3.2 Illite crystallinity index

The illite crystallinity index (ICr) in GAS-CS01 ( $n=146$ ) shows values ranging from ~0.25° at the core bottom to ~0.60° in the Red Layers interval (Fig. 10f and Supplementary Online Material). In between, the ICr values range from ~0.30° to 0.40°. The increase in ICr values in the Red Layers is abrupt (at *ca.* 17.2 ka). In contrast, the ICr decreases gradually from *ca.* 33.5 ka to *ca.* 30 ka.

The ICr values of the Danube tributary river samples ( $n=12$ ) are more homogeneous than in core GAS-CS01. ICr values reach a maximum of only 0.34-0.39° in the northern Alps (Iller River and the Danube at Passau). The ICr minimum (0.20°) is found in the (upper) Drava River sample, eastern Alps (Table 3).

### 6.3.3 $\epsilon_{Nd}$

The  $\epsilon_{Nd}$  dataset used in this study is from core MD04-2790 ( $n=84$ ) for the *ca.* 28-10 ka period (Soulet et al., 2013) and from core GAS-CS01 ( $n=14$ ) for the *ca.* 32-28 ka period (Fig. 10g). GAS-CS01 sediment show an  $\epsilon_{Nd}$  value of -11.0 at *ca.* 28 ka, in line with the  $\epsilon_{Nd}$  signature of -11.3 ± 0.3 ( $n=53$ ) for the glacial MD04-2790 sediment (except Red Layers, see the discussion below).  $\epsilon_{Nd}$  values gradually change towards the core bottom, reaching about -9.4 ± 0.1 ( $n=9$ ) over the *ca.* 33.5-31 ka period (Supplementary Online Material).

## 7. DISCUSSION

### 7.1 Linking GAS-CS01 and MD04-2790 sedimentary sequences to Central European rivers

### 7.1.1 Sedimentological evidence for direct connection between Central European sediment sources and Black Sea depositional sinks

The sedimentological investigation of core GAS-CS01 shows that the continuous, rhythmic layering observed throughout the core (Fig. 4) consists in the succession of coarsening-upward and fining-upward units characteristics of hyperpycnal turbidity-current deposits (*i.e.* hyperpycnites; Mulder et al., 2003), at both the decimetre (lower part of GAS-CS01; Fig. 7) and millimetre scales (upper part of the core; Fig. 9). Thus, we attribute each clastic rhythmite (laminae or beds; Fig. 4) in GAS-CS01 to a discrete hyperpycnal flow. Hyperpycnal flows are defined as a bottom-riding flow evolving from a hyperpycnal plume, which descends to the basin floor as a result of excess density generated by its sediment load. Hyperpycnal flows are thus seen as the continuation of river discharges (*i.e.* floods) in subaqueous environments (Mulder et al., 2003, 1998; Mulder and Syvitski, 1995; Normark and Piper, 1991; Piper and Normark, 2009). They are frequent in freshwater lakes where only small particle density thresholds are necessary to trigger plunging (Mulder et al., 2003). The brackish water conditions prevailing in the Black Sea during the last glacial period (Deuser, 1972; Marret et al., 2009; Ross et al., 1970; Schumilovskikh et al., 2014; Soulet et al., 2010; Wegwerth et al., 2016), combined with the increased surface runoff and sediment load of the western Black Sea rivers (*e.g.* Danube, Dniepr) due to the combination of reduced vegetation cover and widespread permafrost (Kasse et al., 2010; Sidorchuk et al., 2011, 2008) likely favoured the generation of hyperpycnal flows and their deposits at the GAS-CS01 site. The recording of these events, and by extension of the river floods, was also favoured by the proximity to the GAS-CS01 site and the delta of the western Black Sea rivers. Indeed, the glacial lowstand conditions (lake level ~90-120 m below the present sea level; Figs 1 and 2) significantly reduced the river mouth distance to the shelf margin. This evidence suggests that the 3815 hyperpycnites of core GAS-CS01 provides a high-resolution flood record in the north-west Black Sea over the *ca.* 33.5-15 ka interval. Similarities in both the stratigraphy (see the XRF  $Ca_{norm}$  for a prominent example; Fig 5) and sedimentology (*i.e.* rhythmic layering throughout; Fig. 6) observed in core MD04-2790 (Soulet et al., 2011a), ~20 km east of GAS-CS01, highlights the regional imprint of river floods in the north-western part of the glacial Black Sea ‘Lake’.

Reconstructed hyperpycnite frequency indicates that both GAS-CS01 and MD04-2790 sites experienced five main periods of increased flood frequency ( $F_x$ ), each of 1.5-3 kyr duration, over the *ca.* 33.5-15 ka interval. These periods occurred simultaneously at both sites, at

*ca.* 32.5-30.5 ka ( $F_5$ ), corresponding to the peak maximum, and at *ca.* 29-27.5 ka ( $F_4$ ), *ca.* 25.3-23.8 ka ( $F_3$ ), *ca.* 22.3-19 ka ( $F_2$ ) and *ca.* 17.2-15.7 ka ( $F_1$ ) (Fig. 10b,c). The simultaneity of each of the  $F_x$  periods at sites GAS-CS01 and MD04-2790 importantly suggests a common forcing and, potentially, a unique river source at the two sites. This is fittingly demonstrated for the  $F_1$  period, *ca.* 17.2-15.7 ka. Geochemical investigations of MD04-2790 sediment and the use of neodymium isotopic composition show that the sediment composing the Red Layers originated from the decaying FIS and were transported to the Black Sea by the Dniepr River (Soulet et al., 2013). Thus, the  $F_1$  peak in GAS-CS01 and MD04-2790 result from the increase in the frequency of hyperpycnal flows generated at the Dniepr River mouth in response to ice mass fluctuations ~1,200 km northward (Fig. 1; Soulet et al., 2013). The available geochemical dataset reveals that the unique, major river source for hyperpycnal flows over the *ca.* 33.5-17.2 ka interval and the  $F_2$ ,  $F_3$ ,  $F_4$  and  $F_5$  peaks was the Danube River (see Part 7.1.2). The concomitance between climate, sediment input and geochemical signature changes (smectite abundance,  $\epsilon_{Nd}$ , etc.; Fig. 10) excludes both an autogenic origin (*i.e.* internally generated ‘noise’) and lake-level changes (highly debated over the last glacial period; see Lericolais et al., 2011 and Vidal et al., 2010 for contrasting conclusions) for the observed variability in the sedimentary record at the GAS-CS01 and MD04-2790 sites (see Part 7.2).

## 7.1.2 Deciphering river sources

### 7.1.2.1 Neodymium isotopic composition

The sediment of the Red Layers deposited between *ca.* 17.2 and 15.7 ka ( $F_1$  peak) are characterised by  $\epsilon_{Nd}$  values of  $-14.5 \pm 1.3$  (down to -18;  $n=24$ ; Soulet et al., 2013). These values are close to  $\epsilon_{Nd}$  values of sediments of the Baltic area (Boswell et al., 2019, 2018; Soulet et al., 2013; Toucanne et al., 2015) and the rivers Dniepr ( $\epsilon_{Nd}$  of -14; Soulet et al., 2013) and Don (-14.3, this study) showing that the Red Layer sediments originated from the decaying FIS (Soulet et al., 2013). In contrast, the glacial sediment in MD04-2790, from *ca.* 28 ka (oldest  $\epsilon_{Nd}$  analysis in Soulet et al., 2013) to 11.7 ka shows homogenous but very distinct  $\epsilon_{Nd}$  values of  $-11.3 \pm 0.3$  ( $n=53$ ). The latter signature is close to that of the European Alps (Soulet et al., 2013), for which the northern and eastern parts (sediment yield of 160 and 220 t.km<sup>2</sup>.yr<sup>-1</sup>, respectively; Hinderer et al., 2013) represent a substantial contribution to the Danube sediment discharge (mean sediment yield: 83 t.km<sup>2</sup>.yr<sup>-1</sup>). Indeed, Alpine rivers carry fine-grained sediments with modern  $\epsilon_{Nd}$  values ranging from about -10.8 for the Pô River

(south-western Alps; Frost et al., 1986),  $-10.2 \pm 0.5$  for the Rhône River (western Alps; Henry et al., 1994) and -9 for the Rhine (north-western Alps; Bayon et al., 2015) and the Danube (north-eastern Alps) rivers (Soulet et al., 2013). The Nd isotopic values of these Alpine rivers match those of the Quaternary loess of the Carpathian Basin ( $-10.4 \pm 0.6$ ,  $n=20$ ; Újvári et al., 2012). Based on the above, we thus assume that the glacial  $\epsilon_{Nd}$  signatures of  $-11.3 \pm 0.3$  at site MD04-2790 are mainly ascribed to the Alps-Carpathian realm and, by extension, to the Danube River. Nevertheless, it cannot be ruled out that a slight contribution from the Dniepr and Don ( $\epsilon_{Nd}$  of about -14) at the studied sites possibly lowered the ‘pure’  $\epsilon_{Nd}$  signature (-11/-9) of the Danube to its lower (*i.e.* unradiogenic) end members during the last glacial period. This mixing is supported by the significant increase in  $\epsilon_{Nd}$  at *ca.* 33-31 ka ( $F_5$  peak), with radiogenic values (about -9) that undoubtedly excludes all river sources except that of the Danube at that time.

#### 7.1.2.2 Clay mineralogy and illite crystallinity

Illite and smectite dominates the clay mineralogy of GAS-CS01 sediment (Fig. 10d). Illite is a good marker for northern river sources in the Black Sea (Stoffers and Müller, 1978; Tudryn et al., 2016), including the Danube River (Table 3), but it shows a maximal abundance (>40%) both during the ‘non-radiogenic’ Dniepr-sourced  $F_1$  peak (*i.e.* Red Layers;  $\epsilon_{Nd}$  down to -18) and the ‘radiogenic’ Danube-sourced  $F_5$  peak ( $\epsilon_{Nd}$  about -9). Thus, it cannot be used to decipher sediment sources in the north-west Black Sea. The combination of illite with kaolinite at GAS-CS01 during the  $F_1$  peak is a peculiar mineralogical signature over the studied interval. Illite and kaolinite are typical of (paleo)soils, pedogenic processes and long-term weathering of sedimentary basins (Chamley, 1989) and appear as the main clay minerals in the Volga and Dniepr-Donets basins (Makshaev and Svitoch, 2016; Misch et al., 2018; Vishnevskaya and Sedaeva, 2000), thus supporting an eastern Central European origin for the deposition of Red Layers (Soulet et al., 2013). On the other hand, the low kaolinite content (~5-15%), together with the  $\epsilon_{Nd}$  values, exclude this source (at least as a major contributor) for glacial sediment input to the margin between *ca.* 33.5 and 17.2 ka.

Smectite content is low to absent in the Red Layers but represents 30 to 60% of the *ca.* 33.5-17.2 ka interval at GAS-CS01. Smectite is well-represented in the Danube drainage basin (Tudryn et al., 2016), particularly in the Dinarides (~20-25%) and the Carpathians (~45-55%; Table 3). In the latter region, it probably originates from chemical hydrothermal weathering of volcanic rocks (Chamley, 1989), as shown in the Tisza basin (*e.g.* Tokaj Mountains and the

Calimani-Gurghiu-Hargita volcanic chain in the northern and eastern Carpathians, respectively; Bobos, 2012; Kiss et al., 2010). In the Alpine domain of the Danube, smectite is restricted to the western part of the Molasse Basin as shown by mineralogical data from the Iller River (40%; Table 3) and the neighbouring Lake Constance (Müller and Quakernaat, 1969). It is important to note that the Rhine glacier covered Lake Constance during the last glacial period and drained into the Danube (Ellwanger et al., 2011), thus providing a significant source of Alpine smectite, with chlorite and dolomite (Müller and Quakernaat, 1969; Schmieder et al., 2004), to the north-west Black Sea. Chlorite content is also significant (~25-30%) in the Alpine domain, in rivers draining both the northern (*e.g.* Inn, Enns) and central eastern Alps (*i.e.* eastern Dolomites and the Drava River; Table 3), where physical weathering at high altitudes dominates (Chamley, 1989; Müller and Quakernaat, 1969). Chlorite content in GAS-CS01 reaches a maximum (~25%) during the F<sub>5</sub> peak together with an increase in dolomite (~10%), a common mineral in the inner Alps (Upper Rhine, Iller, Enns, Drava; Table 3). Thus, the mineralogical signature for the F<sub>5</sub> peak suggests a significant Alpine contribution. More generally, these data together with ε<sub>Nd</sub> support our contention that the Danube River was the main source for sediment input to the margin over the *ca.* 33.5-17.2 ka interval.

This is further supported by the ICr values of  $0.32 \pm 0.04$  at GAS-CS01 (Fig. 10f) that fit well with those obtained for the Danube river samples ( $0.3 \pm 0.05$ ; Table 3). Such values are indicative of illite formation in low-grade metamorphic, high anchizonal conditions (*i.e.* lower greenschist facies; Kubler, 1967), typical of the central and eastern Alps (Bousquet et al., 2008; Krumm et al., 1988; Rantitsch, 1997; Schramm, 1991), the Dinarides (Rainer et al., 2002) and the Carpathians (Árkai, 1991; Árkai et al., 2003; Ciulavu et al., 2008). Interestingly, the F<sub>5</sub> peak shows ICr values down to 0.23° pointing to anchizonal-epizonal thermal conditions (*i.e.* blueschist facies). Such metamorphic conditions are restricted to the western (Gemer and Bükk units; Árkai, 1991) and southern parts (Danubian window; Ciulavu et al., 2008) of the Carpathians, and the Engadine (up to 4,049 m at Piz Bernina) and Taurin (up to 3,898 m at Grossglockner) windows of the eastern and central Alps, respectively (Bousquet et al., 2008; Nievergelt et al., 1996; Schramm, 1991) (Figs 1 and 3). The dolomite enrichment at *ca.* 33-31 ka together with the above ICr values, points to these high alpine sources for the F<sub>5</sub> peak sediments, and by extension to the Drava (and its tributaries; *e.g.* Mur), that drain both the Dolomites and the southern Taurin window, and the Inn rivers. The Inn River has its source in the Bergell Massif (upper Engadine valley) and the Salzach River, its main tributary, drained the northern part of the Taurin window (Fig. 3). Interestingly, the



$\epsilon_{Nd}$  values of the  $F_5$  peak ( $-9.4 \pm 0.1$ ,  $n=9$ ) match with the lower range of  $\epsilon_{Nd}$  values ( $-8.0 \pm 1.4$ ,  $n=9$ ) determined for recent granites in the high Central Alps, including the Bergell Massif (Juteau et al., 1986). Altogether, the neodymium isotopic composition, the clay mineralogy and the ICr highlight the imprint of the Danube River on the north-west Black Sea margin sedimentation during the *ca.* 33.5-17.2 ka interval.

The above finding is in line with geomorphological aspects from the continent interior to the deep Black Sea. First, the modern contribution of the ‘mountainous’ Danube to the Black Sea exceeds that of the ‘lowlands’ Dniester, Dnieper and Don by a factor 8 (see Part 2.1; Milliman and Farnsworth, 2013; Milliman and Syvitski, 1992; Panin and Jipa, 2002), and this difference was strongly enhanced during the last glacial period by the presence of the AIS and Carpathian glaciers in the Danube drainage area (see Part 2.2). In contrast, the southern limb of the FIS entered the Upper Dniepr basin only during a short period *ca.* 20-17 ka (Hughes et al., 2016; Karabanov and Matveyev, 2011; Soulet et al., 2013) (Fig. 1). Moreover, the physiography of the shelf-penetrating Danube Canyon (Popescu et al., 2004) and the high sediment inputs into the Danube deep-sea fan (Constantinescu et al., 2015) are indicative of both a high Danube sediment discharge and of a good connectivity between the Danube delta and the shelf margin where cores GAS-CS01 and MD04-2790 are located (Fig. 2). As such, the physiography of the margin suggests that the GAS-CS01 site was located at only ~20-30 km (~50 km for MD04-2790) from the Danube paleo-delta front (~130 km today) during glacial lowstand conditions (Popescu et al., 2004). Sediments deposited at sites GAS-CS01 and MD04-2790 between *ca.* 33.5 and 17.2 ka therefore record the evolution of the Danube basin.

## 7.2 From the Black Sea sedimentary record to the Alpine Ice Sheet

The above geochemical constraints, together with the timing for the  $F_5$  (*ca.* 32.5-30.5 ka),  $F_4$  (*ca.* 29-27.5 ka),  $F_3$  (*ca.* 25.3-23.8 ka) and  $F_2$  (*ca.* 22.3 to 19 ka) peaks, suggest increased Danube sediment discharges during Heinrich Stadial (HS; *i.e.* a stadial which contains a Heinrich event as defined by Sanchez Goñi and Harrison, 2010) 3 (*ca.* 32-29 ka), Greenland Stadial 4 (*ca.* 28.6-27.8 ka), HS 2 (*ca.* 26-23.5 ka) and during the period corresponding to the transition from the global LGM to the Northern Hemisphere deglaciation, respectively. This correlation highlights a close relationship between Danube River flood frequency and climate change. Nowadays, severe floods in the Alpine-Carpathian range are triggered by heavy, late spring to fall precipitation (*e.g.* Blöschl et al., 2013; Parajka et al., 2010). Holocene

reconstructions of the flood frequency in the central and north-eastern Alps reveal centennial- to millennial-scale fluctuations in flood activity, with high flood activity correlated with cool periods (lows in solar activity, global/alpine glacier advances; Glur et al., 2013; Swierczynski et al., 2013; Wirth et al., 2013).

During the last glacial period, the continental ice build-up promoted cooling and enhanced aridity over Europe (Wu et al., 2007). Precipitation over the Alps and Carpathians occurred mainly in the form of snow. This inhibited the generation of floods from high-altitude sites, but participated in the build-up of glaciers over the Carpathians and of the AIS during the last glacial period (Figs 1 and 3). The AIS built up, reached its glacial maximum extent (covering up to ~50,000 km<sup>2</sup> of the Danube watershed) and thickness (up to 2,500 m in the Hohe Tauern region; Figs 3 and 11), and eventually shrunk over the *ca.* 33-17.2 ka interval (*e.g.* Ivy-Ochs, 2015; Ivy-Ochs et al., 2008; Seguinot et al., 2018). Pullback of glaciers from their maximum extent (*i.e.* Alpine LGM) was underway by *ca.* 24-22 ka, and glaciers had receded within the mountain front by *ca.* 19-18 ka (*e.g.* Ivy-Ochs, 2015; Monegato et al., 2017; Reitner, 2007). This has been widely documented in the north-eastern Alps, including the Rhine, Iller, Isar, Loisach, Inn, Salzach, Enns and Drava valleys (Ellwanger et al., 2011; Habbe and Rögner, 1989; Heiri et al., 2014; Ivy-Ochs et al., 2008; Preusser, 2004; Reitner, 2012; Spötl et al., 2013; Starnberger et al., 2009; Van Husen, 2004). Thus, glaciers retreat must have supplied meltwater and sediment to the to the Danube sediment routing system.

The timing for the Alpine (and Carpathian; Makos et al., 2018, 2013) deglaciation closely matches that of enhanced Danube river floods from *ca.* 22 to 19 ka (F<sub>2</sub> peak). This strongly suggests that the F<sub>2</sub> peak originated from the rapid downstream transport of glacial sediments by AIS (and possibly Carpathian glaciers) meltwater, an assumption supported by the glacial character of the lithic grains found in GAS-CS01 (Supplementary Fig. S3). Onshore ice retreat and concomitant offshore sediment pulse is a common feature of large-scale glaciated sediment routing systems (*e.g.* Mississippi in North America, Kolla and Perlmutter, 1993; Channel River, Rhine, Rhone or Po in Europe, Bonneau et al., 2017; Hinderer, 2001; Pellegrini et al., 2017; Toucanne et al., 2010). In these systems, sediment production greatly exceeds transport capacity, hence downstream release of sediments occurs through pulses of meltwater floods when glaciers retreat. Thus, the F<sub>2</sub> peak (sourced by the AIS-Danube system) as well as the F<sub>1</sub> peak (sourced by the FIS-Dniepr system; Soulet et al., 2013) share the common genetic origin of ice melting. Together with the geochemical results, we thus suggest that the F<sub>5</sub> (*ca.* 32.5-30.5 ka), F<sub>4</sub> (*ca.* 29-27.5 ka) and F<sub>3</sub> (*ca.* 25.3-23.8 ka)



peaks, just like F<sub>2</sub>, were triggered by significant ice melting and meltwater production in the Danube basin.

Little is known about ice fluctuation in the Alps before the deglaciation (*e.g.* Ivy-Ochs et al., 2008). The Tagliamento end moraine system of the south-eastern Alps (Fig. 3) provides insights into AIS fluctuations at the LGM. Two phases of glacial culmination occurred from 26.5 ka and at *ca.* 23-21 ka, separated by a distinct recession which ended no later than *ca.* 23 ka (Monegato et al., 2017) (Fig. 12e). This recession, also recognized in the neighbouring Brenta glacier region (Rossato and Mozzi, 2016; Fig. 3), is concomitant with enhanced Danube river floods and the F<sub>3</sub> peak, while the two glacier advances are coeval with the minima in the Danube River discharge at *ca.* 27-25.3 ka and from *ca.* 23.8 ka (Fig. 12b). The Tagliamento and Brenta glaciers were draining south into the Adriatic, suggesting that the timing of ice fluctuations in the Danube basin was similar to that of the south-eastern Alps with a significant melting episode at *ca.* 25.3-23.8 ka (F<sub>3</sub> peak). The hypothesis is supported by the fact that both the Tagliamento and Brenta glaciers in the south-eastern Alps, and the Inn, Salzach, Traun, Mur and Drava glaciers in the Danube basin were all draining the Hohe Tauern region and the Engadine ice dome in the central eastern Alps (Florineth and Schlüchter, 2000; Seguinot et al., 2018; Fig. 3). Geochemical data characterizing the GAS-CS01 sediment during F<sub>5</sub> peak (high  $\epsilon_{Nd}$  and dolomite content, and low ICr; see Part 7.1 and Fig. 10) also suggest this region as the source of sediment.

F<sub>5</sub> peak in hyperpycnal flow frequency suggests a significant retreat of the AIS at *ca.* 32.5-30.5 ka, with significant remobilization (via meltwater) of the glacial sediment from the Engadine (*e.g.* upper Inn) and Tauern (*e.g.* upper Salzach and Drava) valleys. By extension, this indicates that ice caps were mainly restricted to the high Central Alps before *ca.* 32 ka, with a configuration that could resemble that proposed in Figure 11. This is consistent with the radiocarbon chronology of the Inn Valley that indicates that glaciers were upstream of (but close to) the Innsbruck region (*i.e.* ~100 km of the Alpine foreland) *ca.* 33-32 ka (Spötl et al., 2013). It is noteworthy that the sediment flux at the GAS-CS01 and MD04-2790 sites was greater during the F<sub>5</sub> peak than during the F<sub>2</sub>, F<sub>3</sub>, and F<sub>4</sub> peaks. The amplitude of the F<sub>5</sub> peak could be explained by very low lake-level conditions, basinward migration of the Danube delta or even lateral changes in sediment dispersal and depocenter considering the fact that the F<sub>5</sub> peak is higher at site GAS-CS01 than at site MD04-2790 (Fig. 10). However, these forcings alone cannot explain the alpine geochemical signature (*i.e.* high Central Alps) observed at that time. As a result, we propose that the F<sub>5</sub> peak (as well as the F<sub>4</sub> peak, see below) is the result of a short term glacier retreat event that occurred in the course of the

longer term growing phase of the AIS toward its LGM extent. Resulting pulse in meltwater flood flushed a huge amount of glacial material (*i.e.* no longer available) accumulated in the high Alps since the previous glaciation (MIS 4 or even before; Ivy-Ochs et al., 2008; van Husen, 2000). In addition, our results suggest that the rapid ice growth that led to the invasion of the northern Alpine foreland by the major AIS ice-streams at the LGM started after *ca.* 30.5 ka, *i.e.* after the retreat interval mirrored by the F<sub>5</sub> peak, and ended no later than *ca.* 25.3 ka (onset of the F<sub>3</sub> peak). Such timing is consistent with that reported for the western Carpathians (Makos et al., 2018) and the northern foreland of the Swiss Alps (*e.g.* Rhine glacier; Preusser et al., 2011) (Fig. 12d). Finally, the F<sub>4</sub> peak suggests that the LGM ice advance was interrupted (with glaciers melting in the north-eastern Alps) between *ca.* 29 and 27.5 ka. Our results imply that the ice front in the Inn Valley (Spötl et al., 2013) advanced to its LGM position at a rate of about 20-45 m/year (~100 km in a maximum of 2,200-5,200 years).

This rapid change in AIS volume and extent modified the regional pattern of erosion (*e.g.* Jiao et al., 2018), which may explain the mineralogical changes observed at the GAS-CS01 site after the F<sub>5</sub> peak. This includes in particular the dominance of smectite in clay assemblages (Fig. 10d) and the disappearance of the signature (ICr,  $\epsilon_{Nd}$ ; Fig. 10f,g) of the Engadine and Tauern source areas. These two regions were rapidly covered by ice (>1 km in thickness at the LGM; Florineth and Schlüchter, 2000) and the sediment-laden meltwater reaching the north-west Black Sea then mainly originated from the ice-streams located in the northern Alpine foreland, including those of the Rhine (Ellwanger et al., 2011) and the Iller valleys (Habbe and Rögner, 1989) that remobilize smectite-rich sediments (Müller and Quakernaat, 1969; Table 3) (Fig. 11).

It is striking how Danube River flood activity *ca.* 33-17 ka closely resembles that of the Po River, Italy (Fig. 12c). Indeed, the F<sub>5</sub>, F<sub>4</sub>, F<sub>3</sub> and F<sub>2</sub> Danube peaks are coeval with rapid increases in sediment flux (up to 200 km<sup>3</sup>/kyr) in the Po delta, which accumulated ~350 m of sediment from the southern AIS *ca.* 32-14 ka (Pellegrini et al., 2018, 2017). This result indicates that the sediment flux of the Danube and Po rivers was not controlled by basin base level changes (*i.e.* water-level fluctuations) because they evolved independently in the Mediterranean and Black Seas during glacial periods (Badertscher et al., 2011). Hence, AIS mass-balance changes must explain the similar sediment flux pattern observed offshore of the Po and Danube rivers during glacial times. The close resemblance between AIS fluctuations in the Danube catchment and the regional precipitation pattern (*i.e.* 7H  $\delta^{18}O$  record of the

Sieben cave system; Luetscher et al., 2015), that partly controlled glacier mass balance (Fig. 12d), suggests that the AIS as a whole was responding to large-scale climate forcing.

### 7.3 Enhanced surface melting of the Alpine Ice Sheet during periods of North Atlantic cooling

In the above discussion, we have showed that the F<sub>5</sub> (*ca.* 32.5-30.5 ka), F<sub>4</sub> (*ca.* 29-27.5 ka), F<sub>3</sub> (*ca.* 25.3-23.8 ka) and F<sub>2</sub> (*ca.* 22.3 to 19 ka) peaks in hyperpycnal flow frequency, resulting from increased Danube sediment discharges, likely resulted from enhanced surface melting of the AIS during HS 3 (*ca.* 32-29 ka), Greenland Stadial 4 (*ca.* 28.6-27.8 ka), HS 2 (*ca.* 26-23.5 ka) and the very end of the global LGM. The timing of these episodes of AIS melting has important implications, at regional to continental scales. The last episode of AIS melting recorded in this study (*i.e.* F<sub>2</sub> peak in the Danube record, and its equivalent in the Po delta) occurred *ca.* 22-19 ka. This time interval corresponds to the onset of the widespread retreat of the Northern Hemisphere ice sheets in response to high northern latitude summer insolation forcing (Carlson and Clark, 2012; Clark et al., 2009). This event is well-documented for the FIS that rapidly retreated in the Baltic lowlands at that time (Hughes et al., 2016; Larsen et al., 2009; Stroeve et al., 2016; Toucanne et al., 2015, 2010). In contrast, the rapid acceleration in the FIS retreat from *ca.* 18 ka (HS 1; Hughes et al., 2016; Toucanne et al., 2015, 2010; Zaragosi et al., 2001) has no AIS equivalent in the north-west Danube and Po sedimentary records. Sediment flushed by AIS fluctuations younger than *ca.* 19 ka (*i.e.* Alpine Lateglacial, including the Gschnitz stadial *ca.* 17-16 ka; Ivy-Ochs, 2015) could have been trapped in glacier-shaped lake basins that formed during deglaciation (Hinderer, 2001). This is consistent with the idea that glaciers receded back into the mountain front during the *ca.* 22-19 ka melting event, especially in the north-eastern Alps (Ivy-Ochs, 2015; Reitner, 2007; van Husen, 2000). By extension, this suggests that the HS 1 melting episode (*ca.* 18-16.5 ka; Toucanne et al., 2015), if it existed at all in the Alps (*e.g.* Ravazzi et al., 2014; Wirsig et al., 2016; Lehmann et al., 2020), impacted only a restricted area and volume of ice. This implies that the forcing that caused the initial retreat of the FIS at *ca.* 22-19 ka was important enough to cause a catastrophic collapse of the AIS. The rapid AIS collapse from *ca.* 22 ka was likely forced by the summer energy of Huybers (2006) that increase from 23.5 ka at 45°N and precede the summer insolation by 1.5 kyr (22 ka at 45°N; Laskar et al., 2004) (Fig. 12e). This suggests that summer energy is a better indicator of glacial variability than insolation intensity

(Huybers, 2006), and confirms that the AIS was a highly dynamic ice-sheet (Seguinot et al., 2018).

AIS melting events during HS 3 and HS 2 in both the Danube and Po catchments (Fig. 12c) and the substantial FIS recession in the Baltic lowlands (Boswell et al., 2019; Hughes et al., 2016; Toucanne et al., 2015) are concomitant within age-model uncertainties. HS 3 and HS 2 in the North Atlantic correspond to cold air and sea-surface temperatures, widespread sea ice and increased iceberg calving and dispersal from surrounding marine-terminating ice margins (Bard et al., 2000; Barker et al., 2009). This is in contrast to the expectation that ice sheets should expand in colder climates and shrink in warmer climates, given constant precipitation. Considering that ablation of land-based ice in mid-latitudes occurs primarily during summer months (Oerlemans et al., 2001), the melting of both the AIS and the FIS during HS 3 and HS 2 (as well as during Greenland Stadial 4 for the AIS) requires a high seasonality in western Europe. High seasonality implies enhanced summertime melting, and thus rising summer temperature, despite severely cool winters and the rapid expansion of sea ice in the North Atlantic as previously discussed by Buizert et al. (2014) and Denton et al. (2005, 2010) for HS 1 and the Younger Dryas cold reversal. High-resolution climate simulations and proxy evidence, including the disintegration of Alps glaciers and the Scottish ice cap, support the occurrence of warm European summers during HS 1 (*i.e.* the so-called Gschnitz stadial in the Alps; Lehmann et al., 2020) and the Younger Dryas (Bromley et al., 2018; Schenk et al., 2018; Magyari et al., 2019). The atmospheric blocking of cold westerly winds over western Europe would be a key mechanism according to Schenk et al. (2018) and such a forcing would have prevailed during HSs of the last glacial period. A strong shift in atmospheric circulation in Europe and over the Alps during and around the LGM (Becker et al., 2016; Beghin et al., 2015; Florineth and Schlüchter, 2000; Ludwig et al., 2016), including millennial-scale climate changes (Luetscher et al., 2015; Monegato et al., 2017), partly support this assumption. Whatever the mechanism responsible for driving stadial decay of the AIS, our results, together with those obtained for the FIS for the last (Toucanne et al., 2015) and penultimate glacial periods (Boswell et al., 2019), strengthens the idea that warm European summers are likely a recurring feature of North Atlantic stadials.

## 8. CONCLUSION

The GAS-CS01 and MD04-2790 glacial sedimentary sequences of the Black Sea ‘Lake’ provide a continuous high-resolution record of the Danube River discharge over the *ca.* 33-

17 ka period. The main results, based on sedimentological, mineralogical and geochemical analysis, are the following:

1. The Danube sediment routing system produced hyperpycnal discharges into the north-west Black Sea throughout the study interval, and their deposits (*i.e.* hyperpycnites) at the GAS-CS01 and MD04-2790 sites provide a continuous, high-resolution paleoflood record of the Danube River at the end of the last glacial period (*ca.* 33-17 ka);

2. Four main periods of enhanced Danube flood frequency ( $F_x$ ), each of 1.5-3 kyr duration, are recorded at *ca.* 32.5-30.5 ka ( $F_5$ ), at *ca.* 29-27.5 ka ( $F_4$ ), at *ca.* 25.3-23.8 ka ( $F_3$ ) and at *ca.* 22.3-19 ka ( $F_2$ ). The sediment of the  $F_2$  period are typical of the Alpine foreland, whereas that deposited during the  $F_5$  period shows mineralogical and geochemical signatures typical of the inner Alps (*i.e.* Engadine and Tauern windows). This suggests that the Danube River activity over the *ca.* 33-17 ka interval provides a valuable record of paleoenvironmental changes in the European Alps, and by extension of the north-eastern AIS, from the second part of MIS 3 to the deglaciation;

3. Our data and continental morphostratigraphical evidence, suggest that the AIS volume has drastically decreased at *ca.* 19 ka (and as soon as *ca.* 22 ka) since no fluctuation of the AIS is recorded at the studied sites after this date. On the other hand, millennial-scale fluctuations of the north-eastern AIS are recorded before the deglacial 22-19 ka event ( $F_2$ ), with three significant melting episodes, able to remobilize glacial sediment to the Black Sea, at *ca.* 32.5-30.5 ka ( $F_5$ ), at *ca.* 29-27.5 ka ( $F_4$ ) and at *ca.* 25.3-23.8 ka ( $F_3$ ). Ice advance occurred between these periods and both the mineralogical and geochemical data suggest that the rapid AIS growth from the inner Alps to its LGM position started from *ca.* 30.5 ka and ended no later than *ca.* 25.3 ka;

4. The  $F_5$  (*ca.* 32.5-30.5 ka),  $F_4$  (*ca.* 29-27.5 ka) and  $F_3$  (*ca.* 25.3-23.8 ka) events indicates enhanced surface melting of the AIS during HS 3 (*ca.* 32-29 ka), Greenland Stadial 4 (*ca.* 28.6-27.8 ka) and HS 2 (*ca.* 26-23.5 ka) when severely cold winters and the rapid expansion of sea ice prevailed in the North Atlantic. This new finding is in line with those recently reported for the southern FIS and reinforces the emerging view that warm summers and hence strong seasonality are key climatic features of western Europe during North Atlantic stadials.

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## DATA AVAILABILITY

Datasets related to this article can be found at <https://doi.org/10.17882/70660>, via the SEANOE open-source online data repository.

## FIGURE CAPTIONS

**Fig. 1:** Paleogeography of the Danube sediment routing system at the Last Glacial Maximum (LGM) including the Alpine Ice-Sheet (AIS, ~130,000 km<sup>2</sup>; Ehlers and Gibbard, 2004), the river course (~2,900 km; ~801,000 km<sup>2</sup>) from the Black Forest (Bf) to the (modern) Danube delta (Dd), and the Danube deep-sea fan (Df; ~20,000 km<sup>2</sup>; Danube Canyon and channel-levees in orange) in the Black Sea 'Lake' (level ~90-120 m below the present sea level; paleoshoreline at -120 m in the Black and Adriatic Seas in light blue). A significant part of the Rhine glacier (~14,000 km<sup>2</sup>; red arrows) drained into the Danube River at the LGM (Ellwanger et al., 2011), thus increasing the AIS surface in the Danube watershed to ~50,000 km<sup>2</sup>. Filled circles indicate the location of the modern river samples, with the colors referring to the Alpine (white), Dinaric (purple) and Carpathians (green) sources. Orange circles refer to the samples collected along the Danube mainstream (Table 3). Tagl.: Tagliamento glacier; EW: Engadine window; TW: Tauern window (Alps); GW: Gemer-Bürkk window; DW: Danubian window (Carpathians). White stars in the Black Sea indicate the location of cores GAS-CS01 and MD04-2790. Black stars in the north-western Alps and the Adriatic Sea refer to the Sieben Hengste cave system (7H; Luetscher et al., 2015) and the PRAD1-2 borehole (Pellegrini et al., 2017, 2018) respectively. The limits of the Fennoscandian Ice Sheet (FIS) at 22 ka (continuous white line) and 17 ka (dashed white line) are also shown (Hughes et al., 2016). The retreat of the FIS in the upper Dniepr from *ca.* 17 ka caused the deposition of the Red Layers into the Black Sea (Soulet et al., 2013). Note the absence of an ice-sheet on the Carpathians at the LGM.

**Fig. 2:** Detailed map of the north-western Black Sea margin showing the location of the long-piston cores GAS-CS01 and MD04-2790 in front of the modern Danube delta. The numerous canyons (including the shelf-penetrating Danube Canyon; Popescu et al., 2004) visible on the continental slope are indicative of high Danube sediment discharge and of a good connectivity between the Danube delta and the shelf margin during the last glacial period, when the Black Sea was a giant freshwater lake and lowstand conditions dominated (see Part 2 for details). Danube paleovalleys from Popescu et al. (2004).

**Fig. 3:** Detailed map of the eastern Alps showing (i) the main geographical and geological features discussed in the main text, including the Tauern (TW) and Engadine (EW) windows (Bousquet et al., 2008); (ii) the main rivers (blue lines) draining into the Danube River (except



the Rhine and Adige); (iii) the Alpine Ice-Sheet (AIS) at the LGM (Ehlers and Gibbard, 2004). The thick white line shows the AIS ice-divide at the LGM (based on Seguinot et al., 2018). Star labeled 7H: Sieben Hengste cave system (Luetscher et al., 2015).

**Fig. 4:** Digital photographs and X-ray images of representative sections of the glacial sequence of core GAS-CS01 (*i.e.* ‘banded lutites’ -Unit III- of Ross and Degens, 1974). Note that the sequence is fully laminated, with clastic rhytmities evolving gradually downward from millimetre- (*i.e.* laminae; Scandinavian-type lacustrine varves of Degens et al., 1978) to decimetre-thick deposits (*i.e.* beds; megavarves of Degens et al., 1978). Clastic rhytmities in GAS-CS01 are interpreted as hyperpycnite sequences. See Figs 7, 8 and 9 for details.

**Fig. 5:** Stratigraphic correlation between cores GAS-CS01 (magenta line) and MD04-2790 (black line; Soulet et al., 2011a) based on XRF-Can<sub>orm</sub> profiles. See section 5 ‘Age-depth model’ and Supplementary Online Material for details, and Table 2 for the radiocarbon dates.

**Fig. 6:** Depth-depth relationship between cores GAS-CS01 and MD04-2790 based on the stratigraphic correlations shown in Figure 5. See section 5 ‘Age-depth model’ and Supplementary Online Material for details, and Table 2 for the radiocarbon dates. The laminae / bed thickness and the Red Layers interval are also shown. Note that sediment accumulation rates (SAR) at site GAS-CS01 are significantly higher than at site MD04-2790 before *ca.* 31 ka (*i.e.* below depth 1595 cmbsf in core GAS-CS01 and depth 1990 cmbsf in MD04-2790), consistent with the significant thickness difference of hyperpycnal beds observed at the two sites on the same time interval. The decimeter-scale hyperpycnite beds recognized in the lower part of core GAS-CS01 are detailed in Figure 7.

**Fig. 7:** High-resolution analysis of the pluricentimeter-thick clastic rhytmities, interpreted as hyperpycnites (H), observed in the basal part of core GAS-CS01 (>28.5 mbsf; here Section 34/34, 3139-3214 cmbsf; *ca.* 33 ka). **a**, Digital photograph; **b**, X-ray image and grey-level measurements (yellow line); **c**, Hyperspectral image highlighting changes in the sediment texture and/or composition. Colored circles refer to the grain-size analyses shown in **e** and **g** (sedimentary sequences S1, S2 and S3); **d**, Schematic lithological succession based on grain-size (**e**) and XRF (**f**) measurements. *Ha* and *Hb* correspond to the (basal) coarsening-up (*i.e.* waxing period of river discharge) and (top) fining-up (*i.e.* waning period of discharge) sequences of the hyperpycnites (Mulder et al., 2003) respectively. C: clay. S: silt-sand. **e**,



Grain-size measurements (median or D50, *i.e.* grain-size - $\mu\text{m}$ - at which 50% of sample is finer). Mean sampling resolution of 0.25 cm ( $n=218$ ). Black arrows highlight some representative coarsening-up and fining-up sequences. Colored circles refer to the grain-size analyses shown in **c** and **g**; **f**, XRF Zr/Rb ratio, here used as a semi-quantitative grain-size indicator (Dypvik and Harris, 2001); **g**, Particle size distribution of discrete sediment samples (colored circles in **c** and **e**) from the coarsening-up sequences S1, S2 and S3.

**Fig. 8:** High-resolution analysis of the centimeter-thick clastic rhythmites, interpreted as hyperpycnites (H), observed in the lower part of core GAS-CS01 (from ~22 to 28.5 mbsf; here Section 30/34, 2804-2859 cmbsf; *ca.* 32 ka). **a**, Digital photograph; **b**, X-ray image and grey-level measurements (yellow line); **c**, Hyperspectral image highlighting changes in the sediment texture and/or composition. Colored circles refer to the grain-size analyses shown in **e** and **g** (sedimentary sequences S1, S2, S3 and S4); **d**, Schematic lithological succession based on grain-size (**e**) and XRF (**f**) measurements. Ha and Hb correspond to the (basal) coarsening-up (*i.e.* waxing period of river discharge) and (top) fining-up (*i.e.* waning period of discharge) sequences of the hyperpycnites (Mulder et al., 2003), respectively. C: clay. S: silt-sand; **e**, Grain-size measurements (median or D50, *i.e.* grain-size - $\mu\text{m}$ - at which 50% of sample is finer). Mean sampling resolution of 0.25 cm ( $n=149$ ). Black arrows highlight some representative coarsening-up and fining-up sequences. Colored circles refer to the grain-size analyses shown in **c** and **g**; **f**, XRF Zr/Rb ratio, here used as a semi-quantitative grain-size indicator (Dypvik and Harris, 2001); **g**, Particle size distribution of discrete sediment samples (colored circles in **c** and **e**) from the coarsening-up sequences S1, S2 and S3.

**Fig. 9:** High-resolution analysis of the subcentimeter-thick clastic rhythmites, interpreted as hyperpycnites (H), observed in core GAS-CS01 (down to ~22 mbsf; here Section 22/34, 2031-2083 cmbsf; *ca.* 31 ka). **a**, Digital photograph; **b**, X-ray image and grey-level measurements (yellow line); **c**, Hyperspectral image highlighting changes in the sediment texture and/or composition; **d**, XRF Zr/Rb ratio, here used as a semi-quantitative grain-size indicator (Dypvik and Harris, 2001); **e**, Focus on the XRF Zr/Rb ratio. The Zr/Rb variability, similar to that observed for thicker clastic rhythmites in GAS-CS01 (Figs 7 and 8), suggest coarsening-up and fining-up sequences; **f**, Thin section photograph (2039-2047.5 cmbsf) showing subcentimeter-thick clastic rhythmites. White circles refer to the grain-size analyses shown in **g**. Mean sampling resolution of 0.25 cm ( $n=31$ ); **g**, SEM-EDX image of a coarsening-up sequences (see the red square in **f** for location). The color variability highlights

the sediment geochemistry (see the color boxes). Coarse grains are mainly quartz (Si) and dolomite grains (Mg, Ca). Note their angular character. Particle size distribution (white boxes) for clayey ( $n=9$ ) and silty ( $n=5$ ) intervals are shown. A schematic lithological succession is proposed on the right. Ha and Hb according to Mulder et al. (2003).

**Fig. 10: a**,  $\delta^{18}\text{O}$  from GISP2 on the GICC05 timescale and the INTIMATE event stratigraphy. Greenland Stadials (GS) and Greenland Interstadials (GI) are the Greenland expressions of the characteristic Dansgaard-Oeschger events that represent cold and warm phases of the North Atlantic region, respectively (Rasmussen et al., 2014). The Greenland synthetic  $\delta^{18}\text{O}$  record (GLT\_syn) of Barker et al. (2011), placed on the absolutely dated Chinese speleothem record (Hulu Cave on this time interval), is also shown; **b**, **c**, Flood frequency and sediment accumulation rates (SAR) off the Danube River, respectively; **d**, **e**, **f**, Clay mineralogy, dolomite abundance and illite crystallinity (three-points average) at core GAS-CS01, respectively. The green arrow highlights the increase in smectite content to higher values, interpreted as the signature of the western Molasse Basin and Rhine glacier in response to the rapid growth of the AIS (see Fig. 11 for details); **g**, Neodymium isotopic composition (expressed in  $\epsilon\text{Nd}$ ) in core MD04-2790 (*ca.* 28-10 ka; Soulet et al., 2013) and GAS-CS01 (*ca.* 33.5-28 ka; this study) as a proxy for sediment provenance. The vertical light blue bars highlight the timing for the  $F_5$  (*ca.* 32.5-30.5 ka),  $F_4$  (*ca.* 29-27.5 ka),  $F_3$  (*ca.* 25.3-23.8 ka),  $F_2$  (*ca.* 22.3- 19 ka) and  $F_1$  (*ca.* 17.2-15.7 ka) peaks -based on both the flood frequency (**b**) and the SAR (**c**)- interpreted as periods of enhanced flux of sediment-laden meltwater on the north-west Black Sea margin. HS refers to Heinrich Stadials.

**Fig. 11:** Paleogeography of the central and eastern Alps, including the hydrographic network (blue lines) and glacial limits (white shaded areas) at (a) *ca.* 33 ka (*i.e.* before Heinrich Stadial 3 =  $F_5$  peak at the GAS-CS01 and MD04-2790 sites) and (b) during the local Last Glacial Maximum (LGM; Ehlers and Gibbard, 2004), *ca.* 26 ka (*i.e.* before Heinrich Stadial 2 =  $F_3$  peak). The proposed configuration at *ca.* 33 ka is based first on the geochemical evidences (ICr,  $\epsilon\text{Nd}$ ) found at GAS-CS01 for significant inputs of terrigenous material from the Engadine (EW) and Taurin (TW) windows (purple dashed lines; via the Inn, Salzach, Enns, Mur and Drava rivers) into the north-west Black Sea *ca.* 32.5-30.5 ka (*i.e.* during the  $F_5$  peak interpreted as AIS melting; see the main text for details); and second, on the low content of smectite at the GAS-CS01 site before *ca.* 30.5 ka (and until *ca.* 33.5 ka at least; Fig. 10d). The

smectite in the Alpine domain being restricted to the western part of the Molasse Basin (*e.g.* Iller River; [Table 3](#)) and the upper Rhine (Müller and Quakernaat, 1969), the clay mineralogy at GAS-CS01 suggests that these areas were mainly free of ice at *ca.* 33 ka (and until *ca.* 30.5 ka) as shown in (a). Such an ice configuration (with the Iller, Lech and Isar/Loisach rivers not connected to the AIS) is found using an equilibrium line altitude at ~2300 m above sea level (white, high-altitude areas; also shown in (b) for comparison). Note that the invasion of the northern Alpine foreland by the major AIS ice-streams (Alpine LGM) strongly modified the glacial hydrographic pattern of the Alpine Danube. Smectite content at GAS-CS01 suggests that the AIS growth (including the Iller glacier; Habbe and Rögner, 1989) and the subsequent re-routing of the Rhine glacier meltwaters to the Danube (Ellwanger et al., 2011) occurred from *ca.* 30.5 ka. The white arrows show the main sources for AIS meltwater to the Danube. HS refers to Heinrich Stadials.

**Fig. 12:** **a**,  $\delta^{18}\text{O}$  from GISP2 (Rasmussen et al., 2014) and the GLT\_syn record (Barker et al., 2011). See [Fig. 10a](#) for details; **b**, Flood frequency of the Danube River ([Fig. 10b](#)). The dashed lines (between *ca.* 17.2 and 15.7 ka) correspond to the Dniepr River flood (Red Layers; Soulet et al., 2013); **c**, Sediment volume of the Po River lowstand wedge, based on seismic reflection data and the PRAD1-2 borehole ([Fig. 1](#); Pellegrini et al., 2017, 2018); **d**, Ice-mass balance of the north-east AIS based on the Danube sediment records (open circles and black line) and the 7H  $\delta^{18}\text{O}$  record from the Sieben Hengste cave system, Switzerland ([Figs 1 and 3](#); Luetscher et al., 2015). Thick blue arrows highlight AIS melting intervals, meltwater production and sediment-laden meltwater inputs to the north-west Black Sea margin (**b**); **d**, Fluctuations of the Rhine (purple; Preusser et al., 2011) and Tagliamento (green; Monegato et al., 2007) glaciers. Arrows highlight the glaciers fluctuations discussed in the main text; **e**, Summer energy (red line,  $\tau \sim 400$ ) (Huybers, 2006) and 21 June-20 July insolation for 45°N (blue line; Laskar et al., 2004). The vertical light blue bars highlight the timing for the F<sub>5</sub> (*ca.* 32.5-30.5 ka), F<sub>4</sub> (*ca.* 29-27.5 ka), F<sub>3</sub> (*ca.* 25.3-23.8 ka), F<sub>2</sub> (*ca.* 22.3-19 ka) and F<sub>1</sub> (*ca.* 17.2-15.7 ka) flood events. HS refers to Heinrich Stadials.

**TABLE CAPTIONS**

**Table 1:** Core locations.

**Table 2:** New radiocarbon ( $^{14}\text{C}$ ) dates for cores GAS-CS01 and MD04-2790. Analyses were carried out at the Beta Analytic radiocarbon dating laboratory (USA).

**Table 3:** Clay mineralogy of the Danube River and its main tributaries. Geographic ‘mountainous’ clusters (raw -†- and weighted -‡- mean values considering the river drainage surface, with the upper Drava excluded in the latter calculation) correspond to the main sediment sources of the Danube watershed. Note that the weighted mean clay mineralogy of these clusters (*i.e.* Alps + Dinarides + Carpathians) is close to that obtained for the Danube River mainstream.

## REFERENCES

- Aksu, A.E., Hiscott, R.N., Kaminski, M.A., Mudie, P.J., Gillespie, H., Abrajano, T., Yaşar, D., 2002. Last glacial–Holocene paleoceanography of the Black Sea and Marmara Sea: stable isotopic, foraminiferal and coccolith evidence. *Mar. Geol.* 190, 119–149. [https://doi.org/10.1016/S0025-3227\(02\)00345-6](https://doi.org/10.1016/S0025-3227(02)00345-6)
- Árkai, P., 1991. Chlorite crystallinity: an empirical approach and correlation with illite crystallinity, coal rank and mineral facies as exemplified by Palaeozoic and Mesozoic rocks of northeast Hungary. *Journal of metamorphic Geology* 9, 723–734. <https://doi.org/10.1111/j.1525-1314.1991.tb00561.x>
- Árkai, P., Faryad, S.W., Vidal, O., Balogh, K., 2003. Very low-grade metamorphism of sedimentary rocks of the Meliata unit, Western Carpathians, Slovakia: implications of phyllosilicate characteristics. *International Journal of Earth Sciences* 92, 68–85. <https://doi.org/10.1007/s00531-002-0303-x>
- Asch, K., Bellenberg, S., Commission for geological Map of the World, 2005. The 1: 5 million international geological map of Europe and adjacent areas (IGME 5000). Bundesanstalt für Geowissenschaften und Rohstoffe.
- Badertscher, S., Fleitmann, D., Cheng, H., Edwards, R.L., Göktürk, O.M., Zumbühl, A., Leuenberger, M., Tüysüz, O., 2011. Pleistocene water intrusions from the Mediterranean and Caspian seas into the Black Sea. *Nat. Geosci.* 4 (4), 236–239. <https://doi.org/10.1038/NGEO1106>
- Bahr, A., Arz, H.W., Lamy, F., Wefer, G., 2006. Late glacial to Holocene paleoenvironmental evolution of the Black Sea, reconstructed with stable oxygen isotope records obtained on ostracod shells. *Earth Planet. Sci. Lett.* 241 (3–4), 863–875. <https://doi.org/10.1016/j.epsl.2005.10.036>
- Bahr, A., Lamy, F., Arz, H., Kuhlmann, H., Wefer, G., 2005. Late glacial to Holocene climate and sedimentation history in the NW Black Sea. *Mar. Geol.* 214 (4), 309–322. <https://doi.org/10.1016/j.margeo.2004.11.013>
- Bahr, A., Lamy, F., Arz, H.W., Major, C., Kwiecien, O., Wefer, G., 2008. Abrupt changes of temperature and water chemistry in the late Pleistocene and early Holocene Black Sea. *Geochem. Geophys. Geosyst.* 9. <https://doi.org/10.1029/2007GC001683>
- Bard, E., Rostek, F., Turon, J.-L., Gendreau, S., 2000. Hydrological impact of Heinrich events in the subtropical northeast Atlantic. *Science* 289, 1321–1324. <https://doi.org/10.1126/science.289.5483.1321>
- Bard, E., Rostek, F., Ménot-Combes, G., 2004. Radiocarbon calibration beyond 20,000 <sup>14</sup>C yr BP by means of planktonic foraminifera of the Iberian Margin. *Quat. Res.* 61(2), 204–214. <https://doi.org/10.1016/j.yqres.2003.11.006>
- Bard, E., Ménot, G., Rostek, F., Licari, L., Böning, P., Edwards, R. L., Cheng, H., Wang, Y., Heaton, T. J., 2013. Radiocarbon calibration/comparison records based on marine sediments from the Pakistan and Iberian margins. *Radiocarbon*, 55(4), 1999–2019. [https://doi.org/10.2458/azu\\_js\\_rc.55.17114](https://doi.org/10.2458/azu_js_rc.55.17114)
- Barker, S., Diz, P., Vautravers, M.J., Pike, J., Knorr, G., Hall, I.R., Broecker, W.S., 2009. Interhemispheric Atlantic seesaw response during the last deglaciation. *Nature* 457 (7233), 1097. <https://doi.org/10.1038/nature07770>

1030 Barker, S., Knorr, G., Edwards, R. L., Parrenin, F., Putnam, A. E., Skinner, L. C., Wolff, E.,  
1031 Ziegler, M., 2011. 800,000 years of abrupt climate variability. *Science*, 334 (6054),  
1032 347–351. <https://doi.org/10.1126/science.1203580>

1033 Bayon, G., Barrat, J.A., Etoubleau, J., Benoit, M., Bollinger, C., Révillon, S., 2009.  
1034 Determination of rare earth elements, Sc, Y, Zr, Ba, Hf and Th in geological samples  
1035 by ICP-MS after Tm addition and alkaline fusion. *Geostandards and Geoanalytical*  
1036 *Research* 33, 51–62. <https://doi.org/10.1111/j.1751-908X.2008.00880.x>

1037 Bayon, G., German, C.R., Boella, R.M., Milton, J.A., Taylor, R.N., Nesbitt, R.W., 2002. An  
1038 improved method for extracting marine sediment fractions and its application to Sr  
1039 and Nd isotopic analysis. *Chem. Geol.* 187, 179–199. [https://doi.org/10.1016/S0009-](https://doi.org/10.1016/S0009-2541(01)00416-8)  
1040 [2541\(01\)00416-8](https://doi.org/10.1016/S0009-2541(01)00416-8)

1041 Bayon, G., Toucanne, S., Skonieczny, C., André, L., Bermell, S., Cheron, S., Dennielou, B.,  
1042 Etoubleau, J., Freslon, N., Gauchery, T., Germain, Y., Jorry, S.J., Ménot, G., Monin,  
1043 L., Ponzevera, E., Rouget, M.-L., Tachikawa, K., Barrat, J.A., 2015. Rare earth  
1044 elements and neodymium isotopes in world river sediments revisited. *Geochimica et*  
1045 *Cosmochimica Acta* 170, 17–38. <https://doi.org/10.1016/j.gca.2015.08.001>

1046 Becker, P., Seguinot, J., Jouvet, G., Funk, M., 2016. Last Glacial Maximum precipitation  
1047 pattern in the Alps inferred from glacier modelling. *Geographica Helvetica* 71, 173–  
1048 187. <https://doi.org/10.5194/gh-71-173-2016>

1049 Beghin, P., Charbit, S., Dumas, C., Kageyama, M., Ritz, C., 2015. How might the North  
1050 American ice sheet influence the northwestern Eurasian climate? *Clim. Past*  
1051 *Discussions* 11, 1467–1490. <https://doi.org/10.5194/cp-11-1467-2015>

1052 Blöschl, G., Nester, T., Komma, J., Parajka, J., Perdigão, R.A., 2013. The June 2013 flood in  
1053 the Upper Danube Basin, and comparisons with the 2002, 1954 and 1899 floods.  
1054 *Hydrology and Earth System Sciences* 17, 5197–5212. [https://doi.org/10.5194/hess-](https://doi.org/10.5194/hess-17-5197-2013)  
1055 [17-5197-2013](https://doi.org/10.5194/hess-17-5197-2013)

1056 Blum, M.D., Törnqvist, T.E., 2000. Fluvial responses to climate and sea-level change: a  
1057 review and look forward. *Sedimentology* 47, 2–48. [https://doi.org/10.1046/j.1365-](https://doi.org/10.1046/j.1365-3091.2000.00008.x)  
1058 [3091.2000.00008.x](https://doi.org/10.1046/j.1365-3091.2000.00008.x)

1059 Bobos, I., 2012. Characterization of smectite to NH<sub>4</sub>-illite conversion series in the fossil  
1060 hydrothermal system of Harghita Băi, East Carpathians, Romania. *American*  
1061 *Mineralogist* 97, 962–982. <https://doi.org/10.2138/am.2012.3801>

1062 Bonneau, L., Jorry, S.J., Toucanne, S., Silva Jacinto, R., Emmanuel, L., 2014. Millennial-  
1063 scale response of a western Mediterranean river to late Quaternary climate changes: a  
1064 view from the deep sea. *J. Geol.* 122, 687–703. <https://doi.org/10.1086/677844>

1065 Bonneau, L., Toucanne, S., Bayon, G., Jorry, S.J., Emmanuel, L., Jacinto, R.S., 2017. Glacial  
1066 erosion dynamics in a small mountainous watershed (Southern French Alps): A  
1067 source-to-sink approach. *Earth Planet. Sci. Lett.* 458, 366–379.  
1068 <https://doi.org/10.1016/j.epsl.2016.11.004>

1069 Boswell, S.M., Toucanne, S., Creyts, T.T., Hemming, S.R., 2018. Continental-scale transport  
1070 of sediments by the Baltic Ice Stream elucidated by coupled grain size and Nd  
1071 provenance analyses. *Earth Planet. Sci. Lett.* 490, 143–150.  
1072 <https://doi.org/10.1016/j.epsl.2018.03.017>

- 1073 Boswell, S.M., Toucanne, S., Pitel-Roudaut, M., Creyts, T.T., Eynaud, F., Bayon, G., 2019.  
1074 Enhanced surface melting of the Fennoscandian Ice Sheet during periods of North  
1075 Atlantic cooling. *Geology* 47 (7), 664–668. <https://doi.org/10.1130/G46370.1>
- 1076 Bousquet, R., Oberhänsli, R., Goffé, B., Wiederkehr, M., Koller, F., Schmid, S.M., Schuster,  
1077 R., Engi, M., Berger, A., Martinotti, G., 2008. Metamorphism of metasediments at the  
1078 scale of an orogen: a key to the Tertiary geodynamic evolution of the Alps. *Geological*  
1079 *Society, London, Special Publications* 298, 393–411.  
1080 <https://doi.org/10.1144/SP298.18>
- 1081 Bromley, G., Putnam, A., Borns, H., Lowell, T., Sandford, T., Barrell, D., 2018. Interstadial  
1082 Rise and Younger Dryas Demise of Scotland's Last Ice Fields. *Paleoceanography and*  
1083 *Paleoclimatology* 33, 412–429. <https://doi.org/10.1002/2018PA003341>
- 1084 Bronk Ramsey, C., 2001. Development of the Radiocarbon Calibration Program. *Radiocarbon*  
1085 43, 355–363. <https://doi.org/10.1017/S0033822200038212>
- 1086 Bronk Ramsey, C., 2008. Deposition models for chronological records. *Quat. Sci. Rev.*, 27 (1-  
1087 2), 42–60. <https://doi.org/10.1016/j.quascirev.2007.01.019>
- 1088 Bronk Ramsey, C., 2009. Bayesian analysis of radiocarbon dates. *Radiocarbon*, 51 (1), 337-  
1089 360. <https://doi.org/10.1017/S0033822200033865>
- 1090 Bronk Ramsey, C., 2017. Methods for summarizing radiocarbon datasets. *Radiocarbon*, 59  
1091 (6), 1809–1833. <https://doi.org/10.1017/RDC.2017.108>
- 1092 Bronk Ramsey, C., Lee, S., 2013. Recent and Planned Developments of the Program OxCal.  
1093 *Radiocarbon* 55, 720–730. <https://doi.org/10.1017/S0033822200057878>
- 1094 Brown, P.A., Kennett, J.P., 1998. Megaflood erosion and meltwater plumbing changes during  
1095 last North American deglaciation recorded in Gulf of Mexico sediments. *Geology* 26,  
1096 599–602. [https://doi.org/10.1130/0091-7613\(1998\)026<0599:MEAMPC>2.3.CO;2](https://doi.org/10.1130/0091-7613(1998)026<0599:MEAMPC>2.3.CO;2)
- 1097 Busschers, F.S., Kasse, C., Van Balen, R.T., Vandenberghe, J., Cohen, K.M., Weerts, H.J.T.,  
1098 Wallinga, J., Johns, C., Cleveringa, P., Bunnik, F.P.M., 2007. Late Pleistocene  
1099 evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of  
1100 climate change, sea-level oscillation and glacio-isostasy. *Quat. Sci. Rev.* 26, 3216–  
1101 3248. <https://doi.org/10.1016/j.quascirev.2007.07.013>
- 1102 Carlson, A.E., Clark, P.U., 2012. Ice sheet sources of sea level rise and freshwater discharge  
1103 during the last deglaciation. *Rev. Geophys.* 50. <http://doi.org/10.1029/2011RG000371>
- 1104 Chaline, J., Jerz, H., 1984. Arbeitsergebnisse der Subkommission für Europäische  
1105 Quartärstratigraphie. *Stratotypen des Würm-Glazials. Eiszeitalter und Gegenwart* 35,  
1106 185–206.
- 1107 Chamley, H., 1989. *Clay sedimentology*, Springer-Verlag. ed. New York.
- 1108 Ciulavu, M., Mählmann, R.F., Schmid, S.M., Hofmann, H., Seghedi, A., Frey, M., 2008.  
1109 Metamorphic evolution of a very low-to low-grade metamorphic core complex  
1110 (Danubian window) in the South Carpathians. *Geological Society, London, Special*  
1111 *Publications* 298, 281–315. <https://doi.org/10.1144/SP298.14>
- 1112 Clark, P.U., Dyke, A.S., Shakun, J.D., Carlson, A.E., Clark, J., Wohlfarth, B., Mitrovica, J.X.,  
1113 Hostetler, S.W., McCabe, A.M., 2009. The last glacial maximum. *Science* 325, 710–  
1114 714. <https://doi.org/10.1126/science.1172873>

- 1115 Comiti, F., Mao, L., Penna, D., Dell'Agnese, A., Engel, M., Rathburn, S., Cavalli, M., 2019.  
1116 Glacier melt runoff controls bedload transport in Alpine catchments. *Earth and*  
1117 *Planet.Sci. Lett.* 520, 77-86. <https://doi.org/10.1016/j.epsl.2019.05.031>
- 1118 Constantinescu, A.M., Toucanne, S., Dennielou, B., Jorry, S.J., Mulder, T., Lericolais, G.,  
1119 2015. Evolution of the Danube Deep-Sea Fan since the Last Glacial Maximum: new  
1120 insights into Black Sea water-level fluctuations. *Mar. Geol.* 367, 50–68.  
1121 <https://doi.org/10.1016/j.margeo.2015.05.007>
- 1122 Coulthard, T.J., Lewin, J., Macklin, M.G., 2005. Modelling differential catchment response to  
1123 environmental change. *Geomorphology* 69, 222–241.  
1124 <https://doi.org/10.1016/j.geomorph.2005.01.008>
- 1125 Covault, J.A., Romans, B.W., Graham, S.A., Fildani, A., Hilley, G.E., 2011. Terrestrial  
1126 source to deep-sea sink sediment budgets at high and low sea levels: Insights from  
1127 tectonically active Southern California. *Geology* 39, 619–622.  
1128 <https://doi.org/10.1130/G31801.1>
- 1129 de Leeuw, A., Morton, A., van Baak, C.G.C., Vincent, S.J., 2018. Timing of arrival of the  
1130 Danube to the Black Sea: Provenance of sediments from DSDP site 380/380A. *Terra*  
1131 *Nova* 30, 114–124. <https://doi.org/10.1111/ter.12314>
- 1132 Dean, W.E., Arthur, M.A., 2011. Geochemical characteristics of Holocene laminated sapropel  
1133 (unit II) and underlying lacustrine unit III in the Black Sea (Report n°. 2010–1323),  
1134 Open-File Report. <https://doi.org/10.3133/ofr20101323>
- 1135 Debret, M., Copard, Y., Van Exem, A., Bessereau, G., Haeseler, F., Rouzaud, J.N., 2018. The  
1136 color of refractory organic carbon. *Bulletin de la Société Géologique de France* 189  
1137 (2), 9. <https://doi.org/10.1051/bsgf/2018008>
- 1138 Degens, E.T., Ross, D.A., 1972. Chronology of the Black Sea over the last 25,000 years.  
1139 *Chem. Geol.* 10, 1–16. [https://doi.org/10.1016/0009-2541\(72\)90073-3](https://doi.org/10.1016/0009-2541(72)90073-3)
- 1140 Degens, E.T., Stoffers, P., Golubic, S., 1978. Varve chronology: Estimated rates of  
1141 sedimentation in the Black Sea Basin. In D. A. Ross (Ed.), *Initial report of the deep*  
1142 *sea drilling project* (Vol. 42, Part 2, pp. 499–508). Washington, D.C.: U.S.  
1143 Government Printing Office.
- 1144 Deuser, W.G., 1972. Late Pleistocene and Holocene history of the Black Sea as indicated by  
1145 stable isotope studies. *J. Geophys. Res.* 77, 1071–1077.
- 1146 Dypvik, H., Harris, N.B., 2001. Geochemical facies analysis of fine-grained siliciclastics  
1147 using Th/U, Zr/Rb and (Zr+Rb)/Sr ratios. *Chem. Geol.* 181, 131-146.  
1148 [https://doi.org/10.1016/S0009-2541\(01\)00278-9](https://doi.org/10.1016/S0009-2541(01)00278-9)
- 1149 Eglinton, T.I., Benitez-Nelson, B.C., Pearson, A., McNichol, A.P., Bauer, J.E., Druffel, E.R.,  
1150 1997. Variability in radiocarbon ages of individual organic compounds from marine  
1151 sediments. *Science* 277, 796–799. <https://doi.org/10.1126/science.277.5327.796>
- 1152 Ehlers, J., Gibbard, P.L., 2004. Quaternary glaciations-extent and chronology: part I: Europe.  
1153 Elsevier.
- 1154 Ellwanger, D., Wielandt-Schuster, U., Franz, M., Simon, T., 2011. The Quaternary of the  
1155 southwest German Alpine Foreland (Bodensee-Oberschwaben, Baden-Württemberg,  
1156 Southwest Germany). *Quat. Sci. J.* 60, 306–328. <https://doi.org/10.3285/eg.60.2-3.07>



- 1157 Fitzsimmons, K.E., Marković, S.B., Hambach, U., 2012. Pleistocene environmental dynamics  
1158 recorded in the loess of the middle and lower Danube basin. *Quat. Sci. Rev.* 41, 104–  
1159 118. <https://doi.org/10.1016/j.quascirev.2012.03.002>
- 1160 Florineth, D., Schlüchter, C., 2000. Alpine evidence for atmospheric circulation patterns in  
1161 Europe during the Last Glacial Maximum. *Quat. Res.* 54, 295–308.  
1162 <https://doi.org/10.1006/qres.2000.2169>
- 1163 Freymond, C.V., Peterse, F., Fischer, L.V., Filip, F., Giosan, L., Eglinton, T.I., 2017.  
1164 Branched GDGT signals in fluvial sediments of the Danube River basin: Method  
1165 comparison and longitudinal evolution. *Org. Geochem.* 103, 88–96.  
1166 <https://doi.org/10.1016/j.orggeochem.2016.11.002>
- 1167 Frost, C.D., O’Nions, R.K., Goldstein, S.L., 1986. Mass balance for Nd in the Mediterranean  
1168 Sea. *Chem. Geol.* 55, 45–50. [https://doi.org/10.1016/0009-2541\(86\)90126-9](https://doi.org/10.1016/0009-2541(86)90126-9)
- 1169 Gábris, G., Nádor, A., 2007. Long-term fluvial archives in Hungary: response of the Danube  
1170 and Tisza rivers to tectonic movements and climatic changes during the Quaternary: a  
1171 review and new synthesis. *Quat. Sci. Rev.* 26, 2758–2782.  
1172 <https://doi.org/10.1016/j.quascirev.2007.06.030>
- 1173 Garnier, J., Billen, G., Hannon, E., Fonbonne, S., Videnina, Y., Soulie, M., 2002. Modelling  
1174 the Transfer and Retention of Nutrients in the Drainage Network of the Danube River.  
1175 *Estuarine, Coastal and Shelf Science* 54, 285–308.  
1176 <https://doi.org/10.1006/ecss.2000.0648>
- 1177 Gibbard, P.L., 1988. The history of the great northwest European rivers during the past three  
1178 million years. *Philosophical Transactions of the Royal Society of London. B, Biol.*  
1179 *Sci.* 318, 559–602. <https://doi.org/10.1098/rstb.1988.0024>
- 1180 Giosan, L., Donnelly, J.P., Constantinescu, S., Filip, F., Ovejanu, I., Vespremeanu-Stroe, A.,  
1181 Vespremeanu, E., Duller, G.A., 2006. Young Danube delta documents stable Black  
1182 Sea level since the middle Holocene: Morphodynamic, paleogeographic, and  
1183 archaeological implications. *Geology* 34, 757–760. <https://doi.org/10.1130/G22587.1>
- 1184 Giosan, L., Donnelly, J.P., Vespremeanu, E., Bhattacharya, J.P., Olariu, C., Buonaiuto, F.S.,  
1185 2005. River delta morphodynamics: Examples from the Danube delta. L. Giosan, J.P.  
1186 Bhattacharya (Eds.), *River Deltas - Concepts, Models and Examples*, SEPM Special  
1187 Publication, 83 (2005), pp. 393–411.
- 1188 Glur, L., Wirth, S.B., Büntgen, U., Gilli, A., Haug, G.H., Schär, C., Beer, J., Anselmetti, F.S.,  
1189 2013. Frequent floods in the European Alps coincide with cooler periods of the past  
1190 2500 years. *Sci. Rep.* 3, 2770. <https://doi.org/10.1038/srep02770>
- 1191 Goldstein, S.J., Jacobsen, S.B., 1988. Nd and Sr isotopic systematics of river water suspended  
1192 material: implications for crustal evolution. *Earth Planet. Sci. Lett.* 87, 249–265.  
1193 [https://doi.org/10.1016/0012-821X\(88\)90013-1](https://doi.org/10.1016/0012-821X(88)90013-1)
- 1194 Goodbred Jr, S.L., Kuehl, S.A., 2000. Enormous Ganges-Brahmaputra sediment discharge  
1195 during strengthened early Holocene monsoon. *Geology* 28, 1083–1086.  
1196 [https://doi.org/10.1130/0091-7613\(2000\)28<1083:EGSDDS>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<1083:EGSDDS>2.0.CO;2)
- 1197 Gorlach, A., Hang, T., Kalm, V., 2017. GIS-based reconstruction of Late Weichselian  
1198 proglacial lakes in northwestern Russia and Belarus. *Boreas* 46, 486–502.  
1199 <https://doi.org/10.1111/bor.12223>

1200 Grosswald, M.G., 1980. Late Weichselian ice sheet of northern Eurasia. *Quaternary Research*  
1201 13, 1–32. [https://doi.org/10.1016/0033-5894\(80\)90080-0](https://doi.org/10.1016/0033-5894(80)90080-0)

1202 Guichard, F., Carey, S., Arthur, M.A., Sigurdsson, H., Arnold, M., 1993. Tephra from the  
1203 Minoan eruption of Santorini in sediments of the Black Sea. *Nature* 363, 610.  
1204 <https://doi.org/10.1038/363610a0>

1205 Habbe, K.A., Rögner, K., 1989. The Pleistocene Iller Glaciers and their outwash fields.  
1206 *Catena Supplement* 15, 311–328.

1207 Haeberli, W., 1983. Frequency and characteristics of glacier floods in the Swiss Alps. *Annals*  
1208 *of Glaciology* 4, 85–90. <https://doi.org/10.3189/S026030550000528>

1209 Hallet, B., Hunter, L., Bogen, J., 1996. Rates of erosion and sediment evacuation by glaciers:  
1210 A review of field data and their implications. *Glob. Planet. Chang.* 12, 213–235.  
1211 [https://doi.org/10.1016/0921-8181\(95\)00021-6](https://doi.org/10.1016/0921-8181(95)00021-6)

1212 Heaton, T. J., Bard, E., & Hughen, K. A., 2013. Elastic tie-pointing-transferring chronologies  
1213 between records via a Gaussian process. *Radiocarbon*, 55(4), 1975–1997.  
1214 [https://doi.org/10.2458/azu\\_js\\_rc.55.17777](https://doi.org/10.2458/azu_js_rc.55.17777)

1215 Heiri, O., Koinig, K.A., Spötl, C., Barrett, S., Brauer, A., Drescher-Schneider, R., Gaar, D.,  
1216 Ivy-Ochs, S., Kerschner, H., Luetscher, M., 2014. Palaeoclimate records 60–8 ka in  
1217 the Austrian and Swiss Alps and their forelands. *Quat. Sci. Rev.* 106, 186–205.  
1218 <https://doi.org/10.1016/j.quascirev.2014.05.021>

1219 Hemming, S.R., 2004. Heinrich events: Massive late Pleistocene detritus layers of the North  
1220 Atlantic and their global climate imprint. *Rev. Geophys.* 42.  
1221 <https://doi.org/10.1029/2003RG000128>

1222 Henry, F., Jeandel, C., Dupre, B., Minster, J.-F., 1994. Particulate and dissolved Nd in the  
1223 western Mediterranean Sea: sources, fate and budget. *Mar. Chem.* 45, 283–305.  
1224 [https://doi.org/10.1016/0304-4203\(94\)90075-2](https://doi.org/10.1016/0304-4203(94)90075-2)

1225 Hinderer, M., 2001. Late Quaternary denudation of the Alps, valley and lake fillings and  
1226 modern river loads. *Geodinamica Acta* 14, 231–263.  
1227 <https://doi.org/10.1080/09853111.2001.11432446>

1228 Hinderer, M., Kastowski, M., Kamelger, A., Bartolini, C., Schlunegger, F., 2013. River loads  
1229 and modern denudation of the Alps: a review. *Earth. Sci. Rev.* 118, 11–44.  
1230 <https://doi.org/10.1016/j.earscirev.2013.01.001>

1231 Hughes, A.L.C., Gyllencreutz, R., Lohne, Ø.S., Mangerud, J., Svendsen, J.I., 2016. The last  
1232 Eurasian ice sheets - a chronological database and time-slice reconstruction, DATED-  
1233 1. *Boreas* 45, 1–45. <https://doi.org/10.1111/bor.12142>

1234 Hughes, P.D., Woodward, J.C., Van Calsteren, P.C., Thomas, L.E., 2011. The glacial history  
1235 of the Dinaric Alps, Montenegro. *Quat. Sci. Rev.* 30, 3393–3412.  
1236 <https://doi.org/10.1016/j.quascirev.2011.08.016>

1237 Huybers, P., 2006. Early Pleistocene glacial cycles and the integrated summer insolation  
1238 forcing. *Science* 313, 508–511. <https://doi.org/10.1126/science.1125249>

1239 Ivy-Ochs, S., 2015. Glacier variations in the European Alps at the end of the last glaciation.  
1240 *Cuadernos de Investigación Geográfica* 41, 295. <https://doi.org/10.18172/cig.2750>

1241 Ivy-Ochs, S., Kerschner, H., Reuther, A., Preusser, F., Heine, K., Maisch, M., Kubik, P.W.,  
1242 Schlüchter, C., 2008. Chronology of the last glacial cycle in the European Alps. *J.*  
1243 *Quat. Sci.* 23, 559–573. <https://doi.org/10.1002/jqs.1202>

1244 Ivy-Ochs, S., Lucchesi, S., Baggio, P., Fioraso, G., Gianotti, F., Monegato, G., Graf, A.A.,  
1245 Akçar, N., Christl, M., Carraro, F., Forno, M.G., Schlüchter, C., 2018. New  
1246 geomorphological and chronological constraints for glacial deposits in the Rivoli-  
1247 Avigliana end-moraine system and the lower Susa Valley (Western Alps, NW Italy). *J.*  
1248 *Quat. Sci.* 33, 550–562. <https://doi.org/10.1002/jqs.3034>

1249 Ivy-Ochs, S., Schäfer, J., Kubik, P.W., Synal, H.-A., Schlüchter, C., 2004. Timing of  
1250 deglaciation on the northern Alpine foreland (Switzerland). *Eclogae Geologicae*  
1251 *Helveticae* 97, 47–55.

1252 Jacobsen, S.B., Wasserburg, G.J., 1980. Sm-Nd isotopic evolution of chondrites. *Earth Planet.*  
1253 *Sci. Lett.* 50, 139–155. [https://doi.org/10.1016/0012-821X\(80\)90125-9](https://doi.org/10.1016/0012-821X(80)90125-9)

1254 Jacq, K., Perrette, Y., Fanget, B., Sabatier, P., Coquin, D., Martinez-Lamas, R., Debret, M.,  
1255 Arnaud, F., 2019. High-resolution prediction of organic matter concentration with  
1256 hyperspectral imaging on a sediment core. *Science of the Total Environment* 663,  
1257 236–244. <https://doi.org/10.1016/j.scitotenv.2019.01.320>

1258 Jaeger, J.M., Koppes, M.N., 2016. The role of the cryosphere in source-to-sink systems. *Earth*  
1259 *Sci. Rev.* 153, 43–76. <https://doi.org/10.1016/j.earscirev.2015.09.011>

1260 Jiao, R., Herman, F., Beyssac, O., Adatte, T., Cox, S.C., Nelson, F.E., Neil, H.L., 2018.  
1261 Erosion of the Southern Alps of New Zealand during the last deglaciation. *Geology*  
1262 46, 975–978. <https://doi.org/10.1130/G45160.1>

1263 Jones, G.A., Gagnon, A.R., 1994. Radiocarbon chronology of Black Sea sediments. *Deep-Sea*  
1264 *Res. Part I: Oceanogr.* 41, 531–557. [https://doi.org/10.1016/0967-0637\(94\)90094-9](https://doi.org/10.1016/0967-0637(94)90094-9)

1265 Juteau, M., Michard, A., Albarede, F., 1986. The Pb-Sr-Nd isotope geochemistry of some  
1266 recent circum-Mediterranean granites. *Contributions to Mineralogy and Petrology* 92,  
1267 331–340.

1268 Karabanov, A.K., Matveyev, A.V., 2011. The Pleistocene glaciations in Belarus. *Quaternary*  
1269 *Sciences*. Elsevier, pp. 29–35. <https://doi.org/10.1016/B978-0-444-53447-7.00003-9>

1270 Kasse, C., Bohncke, S.J.P., Vandenberghe, J., Gábris, G., 2010. Fluvial style changes during  
1271 the last glacial–interglacial transition in the middle Tisza valley (Hungary).  
1272 *Proceedings of the Geologists' Association* 121, 180–194.  
1273 <https://doi.org/10.1016/j.pgeola.2010.02.005>

1274 Ker, S., Riboulot, V., 2015. GHASS cruise, R/V Pourquoi Pas?.  
1275 <https://doi.org/10.17600/15000500>

1276 Kisch, H.J., 1991. Illite crystallinity: recommendations on sample preparation, X-ray  
1277 diffraction settings, and interlaboratory samples. *Journal of Metamorphic Geology* 9,  
1278 665–670. <https://doi.org/10.1111/j.1525-1314.1991.tb00556.x>

1279 Kiss, P., Gméling, K., Molnár, F., Pécskay, Z., 2010. Geochemistry of Sarmatian volcanic  
1280 rocks in the Tokaj Mts (NE Hungary) and their relationship to hydrothermal  
1281 mineralization. *Central European Geology* 53, 377–403.  
1282 <https://doi.org/10.1556/CEuGeol.53.2010.4.3>

1283 Kolla, V., Perlmutter, M.A., 1993. Timing of turbidite sedimentation on the Mississippi Fan.  
1284 *Aapg Bulletin* 77, 1129–1141.

1285 Krinsley, D.H., Donahue, J., 1968. Environmental interpretation of sand grain surface textures  
1286 by electron microscopy. *GSA Bulletin* 79, 743–748. [https://doi.org/10.1130/0016-](https://doi.org/10.1130/0016-7606(1968)79[743:EIOSGS]2.0.CO;2)  
1287 [7606\(1968\)79\[743:EIOSGS\]2.0.CO;2](https://doi.org/10.1130/0016-7606(1968)79[743:EIOSGS]2.0.CO;2)

1288 Krumm, H., Petschick, R., Wolf, M., 1988. From diagenesis to anchimetamorphism, upper  
1289 Austroalpine sedimentary cover in Bavaria and Tyrol. *Geodinamica Acta* 2, 33–47.  
1290 <https://doi.org/10.1080/09853111.1988.11105154>

1291 Kubler, B., 1967. La cristallinité de l'illite et les zones tout à fait supérieures du  
1292 métamorphisme. *Etages tectoniques* 105–121.

1293 Kusch, S., Rethemeyer, J., Schefuß, E., Mollenhauer, G., 2010. Controls on the age of  
1294 vascular plant biomarkers in Black Sea sediments. *Geochimica et Cosmochimica Acta*  
1295 74, 7031–7047. <https://doi.org/10.1016/j.gca.2010.09.005>

1296 Kwiecien, O., Arz, H.W., Lamy, F., Wulf, S., Bahr, A., Röhl, U., Haug, G.H., 2008.  
1297 Estimated reservoir ages of the black sea since the last glacial. *Radiocarbon* 50, 99–  
1298 118. <https://doi.org/10.1017/S0033822200043393>

1299 Larsen, N.K., Knudsen, K.L., Krohn, C.F., Kronborg, C., Murray, A.S., Nielsen, O.B., 2009.  
1300 Late Quaternary ice sheet, lake and sea history of southwest Scandinavia—a synthesis.  
1301 *Boreas* 38, 732–761. <https://doi.org/10.1111/j.1502-3885.2009.00101.x>

1302 Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A.C.M., Levrard, B., 2004. A long-  
1303 term numerical solution for the insolation quantities of the Earth. *Astronomy &*  
1304 *Astrophysics* 428, 261–285. <https://doi.org/10.1051/0004-6361:20041335>

1305 Lehmann, B., Herman, F., Valla, P.G., King, G.E., Biswas, R.H., Ivy-Ochs, S., Steinemann,  
1306 O., Christl, M., 2020. Postglacial erosion of bedrock surfaces and deglaciation timing:  
1307 New insights from the Mont Blanc massif (western Alps). *Geology* 48.  
1308 <https://doi.org/10.1130/G46585.1>

1309 Lericolais, G., 2004. ASSEMBLAGE-1 cruise, R/V Marion Dufresne.  
1310 <https://doi.org/10.17600/4200070>

1311 Lericolais, G., Bourget, J., Popescu, I., Jermannaud, P., Mulder, T., Jorry, S., Panin, N., 2013.  
1312 Late Quaternary deep-sea sedimentation in the western Black Sea: New insights from  
1313 recent coring and seismic data in the deep basin. *Glob. Planet. Chang.* 103, 232–247.  
1314 <https://doi.org/10.1016/j.gloplacha.2012.05.002>

1315 Lericolais, G., Guichard, F., Morigi, C., Popescu, I., Bulois, C., Gillet, H., Ryan, W.B.F.,  
1316 2011. Assessment of Black Sea water-level fluctuations since the Last Glacial  
1317 Maximum. *Geology and Geoarchaeology of the Black Sea Region: Beyond the Flood*  
1318 *Hypothesis* 473, 33.

1319 Ludwig, P., Schaffernicht, E.J., Shao, Y., Pinto, J.G., 2016. Regional atmospheric circulation  
1320 over Europe during the Last Glacial Maximum and its links to precipitation. *Journal of*  
1321 *Geophysical Research: Atmospheres* 121, 2130–2145.  
1322 <https://doi.org/10.1002/2015JD024444>

1323 Luetscher, M., Boch, R., Sodemann, H., Spötl, C., Cheng, H., Edwards, R.L., Frisia, S., Hof,  
1324 F., Müller, W., 2015. North Atlantic storm track changes during the Last Glacial  
1325 Maximum recorded by Alpine speleothems. *Nature Com.* 6.  
1326 <https://doi.org/10.1038/ncomms7344>

- 1327 Macklin, M.G., Benito, G., Gregory, K.J., Johnstone, E., Lewin, J., Michczyńska, D.J., Soja,  
1328 R., Starkel, L., Thorndycraft, V.R., 2006. Past hydrological events reflected in the  
1329 Holocene fluvial record of Europe. *Catena* 66, 145–154.  
1330 <https://doi.org/10.1016/j.catena.2005.07.015>
- 1331 Macklin, M.G., Johnstone, E., Lewin, J., 2005. Pervasive and long-term forcing of Holocene  
1332 river instability and flooding in Great Britain by centennial-scale climate change. *The*  
1333 *Holocene* 15, 937–943. <https://doi.org/10.1191/0959683605hl867ft>
- 1334 Macklin, M.G., Lewin, J., 2003. River sediments, great floods and centennial-scale Holocene  
1335 climate change. *J. Quat. Sci.* 18, 101–105. <https://doi.org/10.1002/jqs.751>
- 1336 Macklin, M.G., Lewin, J., Woodward, J.C., 2012. The fluvial record of climate change.  
1337 *Philosophical Transactions of the Royal Society A: Mathematical, Physical and*  
1338 *Engineering Sciences* 370, 2143–2172. <https://doi.org/10.1098/rsta.2011.0608>
- 1339 Magyari, E. K., Pál, I., Vincze, I., Veres, D., Jakab, G., Braun, M., Szalai, Z., Szabo, Z.,  
1340 Korponai, J., 2019. Warm Younger Dryas summers and early late glacial spread of  
1341 temperate deciduous trees in the Pannonian Basin during the last glacial termination  
1342 (20-9 kyr cal BP). *Quat. Sci. Rev.*, 225, 105980.  
1343 <https://doi.org/10.1016/j.quascirev.2019.105980>
- 1344 Major, C., Ryan, W., Lericolais, G., Hajdas, I., 2002. Constraints on Black Sea outflow to the  
1345 Sea of Marmara during the last glacial–interglacial transition. *Mar. Geol.* 190 (1-2),  
1346 19–34. [https://doi.org/10.1016/S0025-3227\(02\)00340-7](https://doi.org/10.1016/S0025-3227(02)00340-7)
- 1347 Major, C.O., Goldstein, S.L., Ryan, W.B., Lericolais, G., Piotrowski, A.M., Hajdas, I., 2006.  
1348 The co-evolution of Black Sea level and composition through the last deglaciation and  
1349 its paleoclimatic significance. *Quat. Sci. Rev.* 25, 2031–2047.  
1350 <https://doi.org/10.1016/j.quascirev.2006.01.032>
- 1351 Makos, M., Dzierżek, J., Nitychoruk, J., Zreda, M., 2014. Timing of glacier advances and  
1352 climate in the High Tatra Mountains (Western Carpathians) during the Last Glacial  
1353 Maximum. *Quat. Res.* 82, 1–13. <https://doi.org/10.1016/j.yqres.2014.04.001>
- 1354 Makos, M., Nitychoruk, J., Zreda, M., 2013. Deglaciation chronology and paleoclimate of the  
1355 Pięciu Stawów Polskich/Roztoki Valley, high Tatra Mountains, Western Carpathians,  
1356 since the Last Glacial Maximum, inferred from <sup>36</sup>Cl exposure dating and glacier–  
1357 climate modelling. *Quat. Int.* 293, 63–78. <https://doi.org/10.1016/j.quaint.2012.01.016>
- 1358 Makos, M., Rinterknecht, V., Braucher, R., Tołoczko-Pasek, A., Arnold, M., Aumaître, G.,  
1359 Bourlès, D., Keddadouche, K., 2018. Last Glacial Maximum and Lateglacial in the  
1360 Polish High Tatra Mountains - Revised deglaciation chronology based on the 10 Be  
1361 exposure age dating. *Quat. Sci. Rev.* 187, 130–156.  
1362 <https://doi.org/10.1016/j.quascirev.2018.03.006>
- 1363 Makshaev, R.R., Svitoch, A.A., 2016. Chocolate clays of the Northern Caspian Sea Region:  
1364 Distribution, structure, and origin. *Quat. Int.* 409, 44–49.  
1365 <https://doi.org/10.1016/j.quaint.2015.07.018>
- 1366 Mangerud, J., Jakobsson, M., Alexanderson, H., Astakhov, V., Clarke, G.K., Henriksen, M.,  
1367 Hjort, C., Krinner, G., Lunkka, J.-P., Möller, P., 2004. Ice-dammed lakes and  
1368 rerouting of the drainage of northern Eurasia during the Last Glaciation. *Quat. Sci.*  
1369 *Rev.* 23, 1313–1332. <https://doi.org/10.1016/j.quascirev.2003.12.009>

- 1370 Marković, S.B., Stevens, T., Kukla, G.J., Hambach, U., Fitzsimmons, K.E., Gibbard, P.,  
1371 Buggle, B., Zech, M., Guo, Z., Hao, Q., Wu, H., O'Hara Dhand, K., Smalley, I.J.,  
1372 Újvári, G., Sümegi, P., Timar-Gabor, A., Veres, D., Sirocko, F., Vasiljević, D.A., Jary,  
1373 Z., Svensson, A., Jović, V., Lehmkuhl, F., Kovács, J., Svirčev, Z., 2015. Danube loess  
1374 stratigraphy - Towards a pan-European loess stratigraphic model. *Earth Sci. Rev.* 148,  
1375 228–258. <https://doi.org/10.1016/j.earscirev.2015.06.005>
- 1376 Marret, F., Mudie, P., Aksu, A., Hiscott, R. N., 2009. A Holocene dinocyst record of a two-  
1377 step transformation of the Neoeuxinian brackish water lake into the Black Sea. *Quat.*  
1378 *Int.* 197(1-2), 72-86. <https://doi.org/10.1016/j.quaint.2007.01.010>
- 1379 Matenco, L., Andriessen, P., 2013. Quantifying the mass transfer from mountain ranges to  
1380 deposition in sedimentary basins: Source to sink studies in the Danube Basin–Black  
1381 Sea system. *Glob. Planet. Chang.* 103, 1–18.  
1382 <https://doi.org/10.1016/j.gloplacha.2013.01.003>
- 1383 Mayr, C., Stojakowits, P., Lempe, B., Blaauw, M., Diersche, V., Grohgan, M., Correa, M.L.,  
1384 Ohlendorf, C., Reimer, P., Zolitschka, B., 2019. High-resolution geochemical record  
1385 of environmental changes during MIS 3 from the northern Alps (Nesseltalgraben,  
1386 Germany). *Quat. Sci. Rev.* 218, 122–136.  
1387 <https://doi.org/10.1016/j.quascirev.2019.06.013>
- 1388 Ménot, G., Bard, E., Rostek, F., Weijers, J.W., Hopmans, E.C., Schouten, S., Damsté, J.S.S.,  
1389 2006. Early reactivation of European rivers during the last deglaciation. *Science* 313,  
1390 1623–1625. <https://doi.org/10.1126/science.1130511>
- 1391 Milliman, J.D., Farnsworth, K.L., 2013. River discharge to the coastal ocean: a global  
1392 synthesis. Cambridge University Press.
- 1393 Milliman, J.D., Syvitski, J.P., 1992. Geomorphic/tectonic control of sediment discharge to the  
1394 ocean: the importance of small mountainous rivers. *J. Geol.* 100, 525–544.  
1395 <https://doi.org/10.1086/629606>
- 1396 Misch, D., Wegerer, E., Gross, D., Sachsenhofer, R.F., Rachetti, A., Gratzner, R., 2018.  
1397 Mineralogy and facies variations of Devonian and Carboniferous shales in the  
1398 Ukrainian Dniepr-Donets Basin. *Aust. J. Earth. Sci.* 111, 15–25.  
1399 <https://doi.org/10.17738/ajes.2018.0002>
- 1400 Mol, J., Vandenberghe, J., Kasse, C., 2000. River response to variations of periglacial climate  
1401 in mid-latitude Europe. *Geomorphology* 33, 131–148. [https://doi.org/10.1016/S0169-555X\(99\)00126-9](https://doi.org/10.1016/S0169-555X(99)00126-9)
- 1403 Monegato, G., Ravazzi, C., Donegana, M., Pini, R., Calderoni, G., Wick, L., 2007. Evidence  
1404 of a two-fold glacial advance during the last glacial maximum in the Tagliamento end  
1405 moraine system (eastern Alps). *Quat. Res.* 68, 284–302.  
1406 <https://doi.org/10.1016/j.yqres.2007.07.002>
- 1407 Monegato, G., Scardia, G., Hajdas, I., Rizzini, F., Piccin, A., 2017. The Alpine LGM in the  
1408 boreal ice-sheets game. *Sci. Rep.* 7, 2078. [https://doi.org/10.1038/s41598-017-02148-](https://doi.org/10.1038/s41598-017-02148-7)  
1409 7
- 1410 Mulder, T., Alexander, J., 2001. The physical character of subaqueous sedimentary density  
1411 flows and their deposits. *Sedimentology* 48, 269–299. [https://doi.org/10.1046/j.1365-](https://doi.org/10.1046/j.1365-3091.2001.00360.x)  
1412 3091.2001.00360.x

- 1413 Mulder, T., Syvitski, J.P., 1995. Turbidity currents generated at river mouths during  
1414 exceptional discharges to the world oceans. *J. Geol.* 103, 285–299.  
1415 <https://doi.org/10.1086/629747>
- 1416 Mulder, T., Syvitski, J.P., Migeon, S., Faugeres, J.-C., Savoye, B., 2003. Marine hyperpycnal  
1417 flows: initiation, behavior and related deposits. A review. *Mar. Pet. Geol.* 20, 861–  
1418 882. <https://doi.org/10.1016/j.marpetgeo.2003.01.003>
- 1419 Mulder, T., Syvitski, J.P., Skene, K.I., 1998. Modeling of erosion and deposition by turbidity  
1420 currents generated at river mouths. *J. Sedim. Res.* 68, 124–137.  
1421 <https://doi.org/10.2110/jsr.68.124>
- 1422 Müller, G., Quakernaat, J., 1969. Diffractometric clay mineral analysis of recent sediments of  
1423 Lake Constance (Central Europe). *Contributions to Mineralogy and Petrology* 22,  
1424 268–275.
- 1425 Muttoni, G., Carcano, C., Garzanti, E., Ghielmi, M., Piccin, A., Pini, R., Rogledi, S.,  
1426 Sciunnach, D., 2003. Onset of major Pleistocene glaciations in the Alps. *Geology* 31,  
1427 989–992. <https://doi.org/10.1130/G19445.1>
- 1428 NGRIP-members, 2004. High resolution record of Northern Hemisphere climate extending  
1429 into the last interglacial period. *Nature* 431, 147–151.  
1430 <https://doi.org/10.1038/nature02805>
- 1431 Nievergelt, P., Liniger, M., Froitzheim, N., Mählmann, R.F., 1996. Early to mid-Tertiary  
1432 crustal extension in the Central Alps: The Turba mylonite zone (eastern Switzerland).  
1433 *Tectonics* 15, 329–340. <https://doi.org/10.1029/93TC02312>
- 1434 Nikishin, A.M., Korotaev, M.V., Ershov, A.V., Brunet, M.-F., 2003. The Black Sea basin:  
1435 tectonic history and Neogene–Quaternary rapid subsidence modelling.  
1436 *Sedim. Geol.* 156, 149–168. [https://doi.org/10.1016/S0037-0738\(02\)00286-5](https://doi.org/10.1016/S0037-0738(02)00286-5)
- 1437 Normark, W.R., Piper, D.J., 1991. Initiation processes and flow evolution of turbidity  
1438 currents: implications for the depositional record. *SEPM Special Publications*, 46,  
1439 207–230
- 1440 Oerlemans, J., Greuell, W., Denby, B., van de Wal, R., 2001. 10 years of mass-balance  
1441 measurements along a transect near Kangerlussuaq, central West Greenland. *Journal*  
1442 *of Glaciology* 47, 157–158. <https://doi.org/10.3189/172756501781832458>
- 1443 Panin, N., Jipa, D., 2002. Danube River sediment input and its interaction with the north-  
1444 western Black Sea. *Estuar. Coast. Shelf Sci.* 54, 551–562.  
1445 <https://doi.org/10.1006/ecss.2000.0664>
- 1446 Panin, N., Panin, S., Herz, N., Noakes, J.E., 1983. Radiocarbon dating of Danube delta  
1447 deposits. *Quat. Res.* 19, 249–255. [https://doi.org/10.1016/0033-5894\(83\)90008-X](https://doi.org/10.1016/0033-5894(83)90008-X)
- 1448 Parajka, J., Kohnová, S., Bálint, G., Barbuc, M., Borga, M., Claps, P., Cheval, S., Dumitrescu,  
1449 A., Gaume, E., Hlavčová, K., 2010. Seasonal characteristics of flood regimes across  
1450 the Alpine–Carpathian range. *J. Hydrol.* 394, 78–89.  
1451 <https://doi.org/10.1016/j.jhydrol.2010.05.015>
- 1452 Pastre, J.-F., Limondin-Lozouet, N., Leroyer, C., Ponel, P., Fontugne, M., 2003. River system  
1453 evolution and environmental changes during the Lateglacial in the Paris Basin  
1454 (France). *Quat. Sci. Rev.* 22, 2177–2188. [https://doi.org/10.1016/S0277-3791\(03\)00147-1](https://doi.org/10.1016/S0277-3791(03)00147-1)  
1455



Pellegrini, C., Asioli, A., Bohacs, K.M., Drexler, T.M., Feldman, H.R., Sweet, M.L., Maselli, V., Rovere, M., Gamberi, F., Dalla Valle, G., 2018. The late Pleistocene Po River lowstand wedge in the Adriatic Sea: Controls on architecture variability and sediment partitioning. *Mar. Pet. Geol.* 96, 16–50. <https://doi.org/10.1016/j.marpetgeo.2018.03.002>

Pellegrini, C., Maselli, V., Gamberi, F., Asioli, A., Bohacs, K.M., Drexler, T.M., Trincardi, F., 2017. How to make a 350-m-thick lowstand systems tract in 17,000 years: The Late Pleistocene Po River (Italy) lowstand wedge. *Geology* 45, 327–330. <https://doi.org/10.1130/G38848.1>

Piper, D.J., Normark, W.R., 2009. Processes that initiate turbidity currents and their influence on turbidites: a marine geology perspective. *J. Sedim. Res.* 79, 347–362. <https://doi.org/10.2110/jsr.2009.046>

Popescu, I., Lericolais, G., Panin, N., Normand, A., Dinu, C., Le Drezen, E., 2004. The Danube submarine canyon (Black Sea): Morphology and sedimentary processes. *Mar. Geol.* 206 (1-4), 249–265. <https://doi.org/10.1016/j.margeo.2004.03.003>

Popescu, I., Lericolais, G., Panin, N., Wong, H.K., Droz, L., 2001. Late Quaternary channel avulsions on the Danube deep-sea fan, Black Sea. *Mar. Geol.* 179, 25–37. [https://doi.org/10.1016/S0025-3227\(01\)00197-9](https://doi.org/10.1016/S0025-3227(01)00197-9)

Preusser, F., 2004. Towards a chronology of the Late Pleistocene in the northern Alpine Foreland. *Boreas* 33, 195–210. <https://doi.org/10.1111/j.1502-3885.2004.tb01141.x>

Preusser, F., Graf, H.R., Keller, O., Krayss, E., Schlüchter, C., 2011. Quaternary glaciation history of northern Switzerland. *Quat. Sci. J.* 60, 282–305. <https://doi.org/10.3285/eg.60.2-3.06>

Rainer, T., Herlec, U., Rantitsch, G., Sachsenhofer, R.F., Vrabec, M., 2002. Organic matter maturation vs clay mineralogy: A comparison for Carboniferous to Eocene sediments from the Alpine-Dinaride junction (Slovenia, Austria). *Geologija* 45, 513–518.

Rantitsch, G., 1997. Thermal history of the Carnic Alps (Southern Alps, Austria) and its palaeogeographic implications. *Tectonophysics* 272, 213–232. [https://doi.org/10.1016/S0040-1951\(96\)00259-4](https://doi.org/10.1016/S0040-1951(96)00259-4)

Rasmussen, S. O., Bigler, M., Blockley, S. P., Blunier, T., Buchardt, S. L., Clausen, H., Cvijanovic, I., Dahl-Jensen, D., Johnsen, S. J., Fisher, H., Gkinis, V., Guillevic, M., Hoek, W. Z., Lowe, J. J., Pedro, J. B., Poop, T., Walker, I. K., Wheatley, J. J., Winstrup, M. 2014. A stratigraphic framework for abrupt climatic changes during the Last Glacial period based on three synchronized Greenland ice-core records: refining and extending the INTIMATE event stratigraphy. *Quat. Sci. Rev.*, 106, 14–28. <https://doi.org/10.1016/j.quascirev.2014.09.007>

Ravazzi, C., Pini, R., Badino, F., De Amicis, M., Londeix, L., Reimer, P.J., 2014. The latest LGM culmination of the Garda Glacier (Italian Alps) and the onset of glacial termination. Age of glacial collapse and vegetation chronosequence. *Quat. Sci. Rev.* 105, 26–47. <https://doi.org/10.1016/j.quascirev.2014.09.014>

Reber, R., Akcar, N., Ivy-Ochs, S., Tikhomirov, D., Burkhalter, R., Zahno, C., Lüthold, A., Kubik, P., Vockenhuber, C., Schlüchter, C., 2014. Deglaciation of the Reuss Glacier in the Alps at the end of the Last Glacial Maximum. *Swiss Journal of Geosciences* 107, 293–307. <https://doi.org/10.1007/s00015-014-0169-5>

1500 Reimer, P.J., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Ramsey, C.B., Buck, C.E.,  
 1501 Cheng, H., Edwards, R.L., Friedrich, M., 2013. IntCal13 and Marine13 radiocarbon  
 1502 age calibration curves 0–50,000 years cal BP. *Radiocarbon* 55, 1869–1887.  
 1503 [https://doi.org/10.2458/azu\\_js\\_rc.55.16947](https://doi.org/10.2458/azu_js_rc.55.16947)

1504 Reitner, J.M., 2012. Challenges in deciphering the Last Glacial Cycle in Eastern Alps:  
 1505 Examples from the Drau glacier system. *Quat. Int.* 279, 398.  
 1506 <https://doi.org/10.1016/j.quaint.2012.08.1255>

1507 Reitner, J.M., 2007. Glacial dynamics at the beginning of Termination I in the Eastern Alps  
 1508 and their stratigraphic implications. *Quat. Int.* 164, 64–84.  
 1509 <https://doi.org/10.1016/j.quaint.2006.12.016>

1510 Reuther, A.U., Urdea, P., Geiger, C., Ivy-Ochs, S., Niller, H.-P., Kubik, P.W., Heine, K.,  
 1511 2007. Late Pleistocene glacial chronology of the Pietrele Valley, Retezat Mountains,  
 1512 Southern Carpathians constrained by <sup>10</sup>Be exposure ages and pedological  
 1513 investigations. *Quat. Int.* 164, 151–169. <https://doi.org/10.1016/j.quaint.2006.10.011>

1514 Revel, M., Ducassou, E., Grousset, F.E., Bernasconi, S.M., Migeon, S., Révillon, S., Mascle,  
 1515 J., Murat, A., Zaragosi, S., Bosch, D., 2010. 100,000 years of African monsoon  
 1516 variability recorded in sediments of the Nile margin. *Quat. Sci. Rev.* 29, 1342–1362.  
 1517 <https://doi.org/10.1016/j.quascirev.2010.02.006>

1518 Riboulot, V., Cattaneo, A., Scalabrin, C., Gaillot, A., Jouet, G., Ballas, G., Marsset, T.,  
 1519 Garziglia, S., Ker, S., 2017. Control of the geomorphology and gas hydrate extent on  
 1520 widespread gas emissions offshore Romania. *Bulletin De La Societe Geologique De*  
 1521 *France* 188 (4), 26. <https://doi.org/10.1051/bsgf/2017182>

1522 Ross, D.A., Degens, E.T., MacIlvaine, J., 1970. Black Sea: recent sedimentary history.  
 1523 *Science* 170, 163–165. <https://doi.org/10.1126/science.170.3954.163>

1524 Ross, D.A., Degens, E.T., 1974. Recent sediments of Black Sea. In: Degens, E.T., Ross, D.A.  
 1525 (Eds.), *The Black Sea: Geology, Chemistry, and Biology*. American Association of  
 1526 *Petroleum Geologists*, Tulsa, 183–199.

1527 Rossato, S., Mozzi, P., 2016. Inferring LGM sedimentary and climatic changes in the  
 1528 southern Eastern Alps foreland through the analysis of a <sup>14</sup>C ages database (Brenta  
 1529 megafan, Italy). *Quat. Sci. Rev.* 148, 115–127.  
 1530 <https://doi.org/10.1016/j.quascirev.2016.07.013>

1531 Ryan, W.B., Major, C.O., Lericolais, G., Goldstein, S.L., 2003. Catastrophic flooding of the  
 1532 Black Sea. *Annual Rev. Earth. Planet. Sci.* 31, 525–554.  
 1533 <https://doi.org/10.1146/annurev.earth.31.100901.141249>

1534 Ryan, W.B., Pitman III, W.C., Major, C.O., Shimkus, K., Moskalenko, V., Jones, G.A.,  
 1535 Dimitrov, P., Gorür, N., Sakinç, M., Yüce, H., 1997. An abrupt drowning of the Black  
 1536 Sea shelf. *Mar. Geol.* 138, 119–126. [https://doi.org/10.1016/S0025-3227\(97\)00007-8](https://doi.org/10.1016/S0025-3227(97)00007-8)

1537 Sanchez Goñi, M.F., Harrison, S.P., 2010. Millennial-scale climate variability and vegetation  
 1538 changes during the Last Glacial: Concepts and terminology. *Quat. Sci. Rev.* 29, 2823–  
 1539 2827. <https://doi.org/10.1016/j.quascirev.2009.11.014>

1540 Schenk, F., Välranta, M., Muschitiello, F., Tarasov, L., Heikkilä, M., Björck, S., Brandefelt,  
 1541 J., Johansson, A.V., Näslund, J.-O., Wohlfarth, B., 2018. Warm summers during the  
 1542 Younger Dryas cold reversal. *Nature Com.* 9. [https://doi.org/10.1038/s41467-018-](https://doi.org/10.1038/s41467-018-04071-5)  
 1543 [04071-5](https://doi.org/10.1038/s41467-018-04071-5)

- 1544 Sanchi, L., Ménot, G., Bard, E., 2014. Insights into continental temperatures in the  
1545 northwestern Black Sea area during the Last Glacial period using branched tetraether  
1546 lipids. *Quat. Sci. Rev.*, 84, 98-108. <https://doi.org/10.1016/j.quascirev.2013.11.013>
- 1547 Schmieder, K., Schünemann, B., Schröder, H.G., 2004. Spatial patterns of surface sediment  
1548 variables in the littoral zone of Lake Constance (Germany). *Archiv für Hydrobiologie*  
1549 161, 455–468. <https://doi.org/10.1127/0003-9136/2004/0161-0455>
- 1550 Schramm, J.-M., 1991. The Permian–Triassic of the Gartnerkofel-1 Core (Carnic Alps,  
1551 Austria): Illite crystallinity in shaly sediments and its comparison with pre-Variscan  
1552 sequences. *Abh Geol BA* 45, 69–77.
- 1553 Schumilovskikh, L. S., Fleitmann, D., Nowaczyk, N. R., Behling, H., Marret, F., Wegwerth,  
1554 A., Arz, H. W., 2014. Orbital and millennial-scale environmental changes between 64  
1555 and 25 ka BP recorded in Black Sea sediments. *Clim. Past*, 10, 939-945.  
1556 <https://doi.org/10.5194/cp-10-939-2014>
- 1557 Seguinot, J., Jouvet, G., Huss, M., Funk, M., Ivy-Ochs, S., Preusser, F., 2018. Modelling last  
1558 glacial cycle ice dynamics in the Alps. *The Cryosphere Discussions* 1–30.  
1559 <https://doi.org/10.5194/tc-2018-8>
- 1560 Sidorchuk, A.Y., Panin, A.V., Borisova, O.K., 2011. Surface runoff to the Black Sea from the  
1561 East European Plain during Last Glacial Maximum–Late Glacial time. *Spec Pap Geol*  
1562 *Soc Am* 473, 1–25.
- 1563 Sidorchuk, A.Y., Panin, A.V., Borisova, O.K., 2008. Climate-induced changes in surface  
1564 runoff on the North-Eurasian plains during the late glacial and Holocene. *Water*  
1565 *Resources* 35, 386–396. <https://doi.org/10.1134/S0097807808040027>
- 1566 Simpson, G., Castelltort, S., 2012. Model shows that rivers transmit high-frequency climate  
1567 cycles to the sedimentary record. *Geology* 40, 1131–1134.  
1568 <https://doi.org/10.1130/G33451.1>
- 1569 Sionneau, T., Bout-Roumazeilles, V., Biscaye, P.E., Van Vliet-Lanoe, B., Bory, A., 2008.  
1570 Clay mineral distributions in and around the Mississippi River watershed and Northern  
1571 Gulf of Mexico: sources and transport patterns. *Quat. Sci. Rev.* 27, 1740–1751.  
1572 <https://doi.org/10.1016/j.quascirev.2008.07.001>
- 1573 Skinner, L. C., Fallon, S., Waelbroeck, C., Michel, E., & Barker, S., 2010. Ventilation of the  
1574 deep Southern Ocean and deglacial CO<sub>2</sub> rise. *Science*, 328 (5982), 1147-1151.  
1575 <https://doi.org/10.1126/science.1183627>
- 1576 Smalley, I.J., Leach, J.A., 1978. The origin and distribution of the loess in the Danube Basin  
1577 and associated regions of east-central Europe: a review. *Sed. Geol.* 21, 1–26.  
1578 [https://doi.org/10.1016/0037-0738\(78\)90031-3](https://doi.org/10.1016/0037-0738(78)90031-3)
- 1579 Sommerwerk, N., Hein, T., Schneider-Jajoby, M., Baumgartner, C., Ostojic, M., Paunovic,  
1580 M., Bloesch, J., Siber, R., Tockner, K., 2009. The Danube River Basin. In: K. Tockner,  
1581 C.T. Robinson, U. Uehlinger (Eds.), *Rivers of Europe*, Elsevier Ltd., London, UK  
1582 (2009), 59-113
- 1583 Soulet, G., Delaygue, G., Vallet-Coulomb, C., Böttcher, M.E., Sonzogni, C., Lericolais, G.,  
1584 Bard, E., 2010. Glacial hydrologic conditions in the Black Sea reconstructed using  
1585 geochemical pore water profiles. *Earth. Planet. Sci. Lett.* 296, 57–66.  
1586 <https://doi.org/10.1016/j.epsl.2010.04.045>

1587 Soulet, G., Giosan, L., Flaux, C., Galy, V., 2019. Using Stable Carbon Isotopes to Quantify  
1588 Radiocarbon Reservoir Age Offsets in the Coastal Black Sea. *Radiocarbon* 61, 309–  
1589 318. <https://doi.org/10.1017/RDC.2018.61>

1590 Soulet, G., Ménot, G., Garreta, V., Rostek, F., Zaragosi, S., Lericolais, G., Bard, E. 2011a.  
1591 Black Sea “Lake” reservoir age evolution since the Last Glacial-hydrologic and  
1592 climatic implications. *Earth Planet. Sci.Lett.*, 308(1-2), 245-258.  
1593 <https://doi.org/10.1016/j.epsl.2011.06.002>

1594 Soulet, G., Ménot, G., Lericolais, G., Bard, E., 2011b. A revised calendar age for the last  
1595 reconnection of the Black Sea to the global ocean. *Quat. Sci. Rev.* 30, 1019–1026.  
1596 <https://doi.org/10.1016/j.quascirev.2011.03.001>

1597 Soulet, G., Ménot, G., Bayon, G., Rostek, F., Ponzevera, E., Toucanne, S., Lericolais, G.,  
1598 Bard, E., 2013. Abrupt drainage cycles of the Fennoscandian Ice Sheet. *Proc. Natl.*  
1599 *Acad. Sci. USA* 110, 6682–6687. <https://doi.org/10.1073/pnas.1214676110>

1600 Spötl, C., Reimer, P.J., Starnberger, R., Reimer, R.W., 2013. A new radiocarbon chronology  
1601 of Baumkirchen, stratotype for the onset of the Upper Würmian in the Alps: a new  
1602 radiocarbon chronology of baumkirchen. *J. Quat. Sci.* 28, 552–558.  
1603 <https://doi.org/10.1002/jqs.2645>

1604 Starkel, L., Michczyńska, D.J., Gębica, P., Kiss, T., Panin, A., Perşoiu, I., 2015. Climatic  
1605 fluctuations reflected in the evolution of fluvial systems of Central-Eastern Europe  
1606 (60–8 ka cal BP). *Quat. Int.* 388, 97–118. <https://doi.org/10.1016/j.quaint.2015.04.017>

1607 Starnberger, R., Drescher-Schneider, R., Reitner, J.M., Rodnight, H., Reimer, P.J., Spötl, C.,  
1608 2013. Late Pleistocene climate change and landscape dynamics in the Eastern Alps:  
1609 the inner-alpine Unterangerberg record (Austria). *Quat. Sci. Rev.* 68, 17–42.  
1610 <https://doi.org/10.1016/j.quascirev.2013.02.008>

1611 Starnberger, R., Rodnight, H., Spötl, C., 2011. Chronology of the last glacial maximum in the  
1612 Salzach Palaeoglacier area (Eastern Alps). *J. Quat. Sci.* 26, 502–510.  
1613 <https://doi.org/10.1002/jqs.1477>

1614 Starnberger, R., Terhorst, B., Rähle, W., Peticzka, R., Haas, J.N., 2009. Palaeoecology of  
1615 Quaternary periglacial environments during OIS-2 in the forefields of the Salzach  
1616 Glacier (Upper Austria). *Quat. Int.* 198, 51–61.  
1617 <https://doi.org/10.1016/j.quaint.2008.06.011>

1618 Stoffers, P., Müller, G., 1978. Mineralogy and lithofacies of Black Sea sediments, Leg 42B  
1619 deep sea drilling project. Initial Reports of the Deep Sea Drilling Project 42, 373–411.

1620 Stroeven, A.P., Hättestrand, C., Kleman, J., Heyman, J., Fabel, D., Fredin, O., Goodfellow,  
1621 B.W., Harbor, J.M., Jansen, J.D., Olsen, L., 2016. Deglaciation of fennoscandia. *Quat.*  
1622 *Sci. Rev.* 147, 91–121. <https://doi.org/10.1016/j.quascirev.2015.09.016>

1623 Swierczynski, T., Lauterbach, S., Dulski, P., Delgado, J., Merz, B., Brauer, A., 2013. Mid-to  
1624 late Holocene flood frequency changes in the northeastern Alps as recorded in varved  
1625 sediments of Lake Mondsee (Upper Austria). *Quat. Sci. Rev.* 80, 78–90.  
1626 <https://doi.org/10.1016/j.quascirev.2013.08.018>

1627 Tanaka, T., Togashi, S., Kamioka, H., Amakawa, H., Kagami, H., Hamamoto, T., Yuhara, M.,  
1628 Orihashi, Y., Yoneda, S., Shimizu, H., 2000. JNdi-1: a neodymium isotopic reference  
1629 in consistency with LaJolla neodymium. *Chem. Geol.* 168, 279–281.  
1630 [https://doi.org/10.1016/S0009-2541\(00\)00198-4](https://doi.org/10.1016/S0009-2541(00)00198-4)

1631 Toucanne, S., Soulet, G., Freslon, N., Jacinto, R.S., Dennielou, B., Zaragosi, S., Eynaud, F.,  
1632 Bourillet, J.-F., Bayon, G., 2015. Millennial-scale fluctuations of the European Ice  
1633 Sheet at the end of the last glacial, and their potential impact on global climate. *Quat.*  
1634 *Sci. Rev.* 123, 113–133. <https://doi.org/10.1016/j.quascirev.2015.06.010>

1635 Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B.,  
1636 Pénaud, A., Fontanier, C., Turon, J.L., Cortijo, E., 2009. Timing of massive ‘Fleuve  
1637 Manche’ discharges over the last 350 kyr: insights into the European ice-sheet  
1638 oscillations and the European drainage network from MIS 10 to 2. *Quat. Sci. Rev.* 28,  
1639 1238–1256. <https://doi.org/10.1016/j.quascirev.2009.01.006>

1640 Toucanne, S., Zaragosi, S., Bourillet, J.-F., Marieu, V., Cremer, M., Kageyama, M., Van  
1641 Vliet-Lanoë, B., Eynaud, F., Turon, J.-L., Gibbard, P.L., 2010. The first estimation of  
1642 Fleuve Manche palaeoriver discharge during the last deglaciation: Evidence for  
1643 Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18 ka ago.  
1644 *Earth Planet. Sci. Lett.* 290, 459–473. <https://doi.org/10.1016/j.epsl.2009.12.050>

1645 Tudryn, A., Leroy, S.A., Toucanne, S., Gibert-Brunet, E., Tucholka, P., Lavrushin, Y.A.,  
1646 Dufaure, O., Miska, S., Bayon, G., 2016. The Ponto-Caspian basin as a final trap for  
1647 southeastern Scandinavian Ice-Sheet meltwater. *Quat. Sci. Rev.* 148, 29–43.  
1648 <https://doi.org/10.1016/j.quascirev.2016.06.019>

1649 Újvári, G., Kovács, J., Varga, G., Raucsik, B., Marković, S.B., 2010. Dust flux estimates for  
1650 the Last Glacial Period in East Central Europe based on terrestrial records of loess  
1651 deposits: a review. *Quat. sci. Rev.* 29, 3157–3166.  
1652 <https://doi.org/10.1016/j.quascirev.2010.07.005>

1653 Újvári, G., Varga, A., Ramos, F.C., Kovács, J., Németh, T., Stevens, T., 2012. Evaluating the  
1654 use of clay mineralogy, Sr–Nd isotopes and zircon U–Pb ages in tracking dust  
1655 provenance: An example from loess of the Carpathian Basin. *Chem. Geol.* 304, 83–96.  
1656 <https://doi.org/10.1016/j.chemgeo.2012.02.007>

1657 Urdea, P., 2004. The Pleistocene glaciation of the Romanian Carpathians, in: *Developments*  
1658 *in Quaternary Sciences*. Elsevier, pp. 301–308.

1659 Van Exem, A., Debret, M., Copard, Y., Verpoorter, C., De Wet, G., Lecoq, N., Sorrel, P.,  
1660 Werner, A., Roof, S., Laignel, B., 2019. New source-to-sink approach in an arctic  
1661 catchment based on hyperspectral core-logging (Lake Linné, Svalbard). *Quat. Sci.*  
1662 *Rev.* 203, 128–140. <https://doi.org/10.1016/j.quascirev.2018.10.038>

1663 Van Husen, D., 2004. Quaternary glaciations in Austria, in: *Developments in Quaternary*  
1664 *Sciences*. Elsevier, pp. 1–13.

1665 van Husen, D., 2000. Geological processes during the Quaternary. *Mitteilungen der*  
1666 *Österreichischen Geologischen Gesellschaft* 92, 135–156.

1667 Vandenberghe, J., 2003. Climate forcing of fluvial system development: an evolution of ideas.  
1668 *Quat. Sci. Rev.* 22, 2053–2060. [https://doi.org/10.1016/S0277-3791\(03\)00213-0](https://doi.org/10.1016/S0277-3791(03)00213-0)

1669 Vandenberghe, J., 2002. The relation between climate and river processes, landforms and  
1670 deposits during the Quaternary. *Quat. Int.* 91, 17–23. [https://doi.org/10.1016/S1040-6182\(01\)00098-2](https://doi.org/10.1016/S1040-6182(01)00098-2)

1671

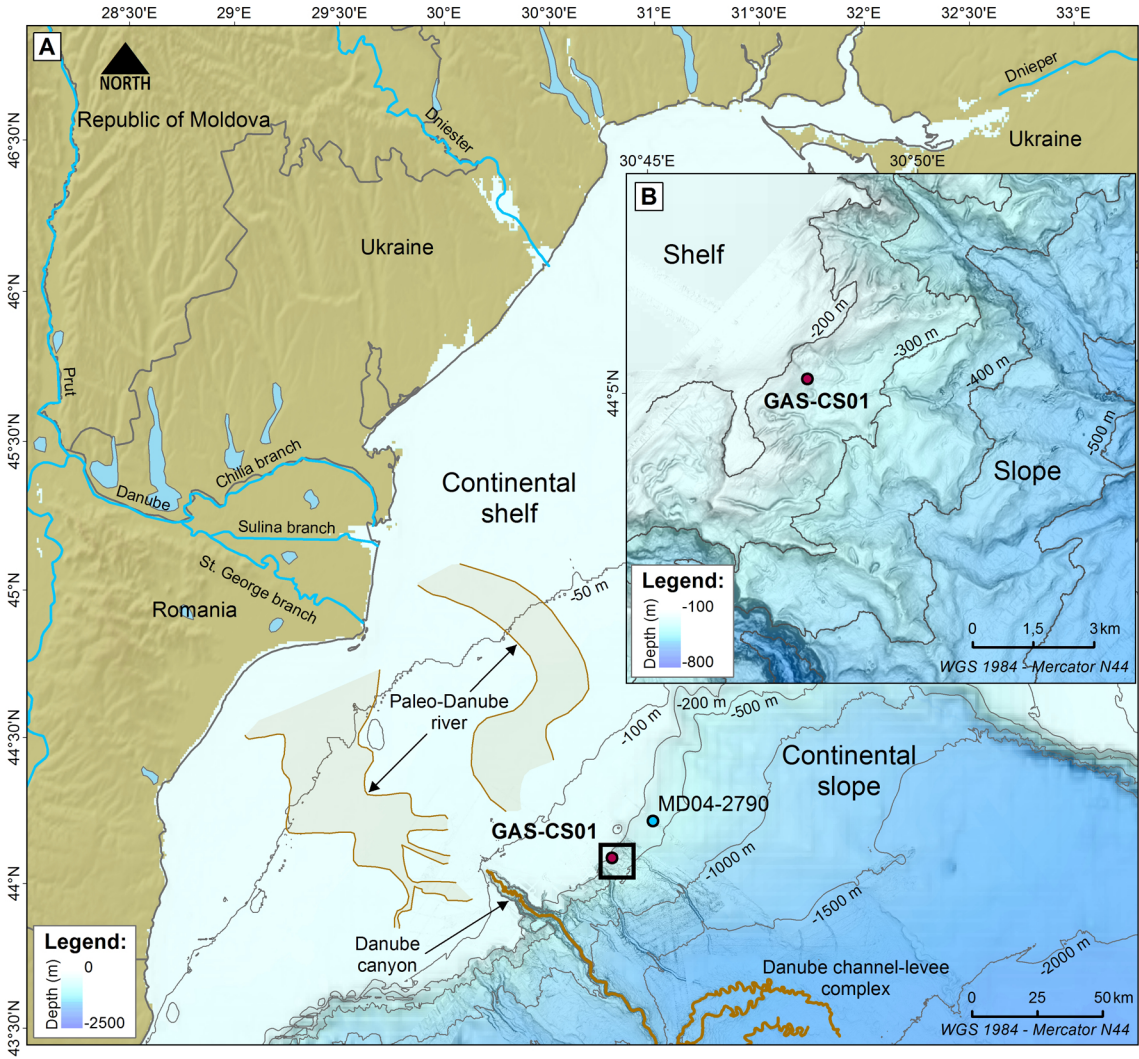
1672 Vidal, L., Menot, G., Joly, C., Bruneton, H., Rostek, F., Çağatay, M.N., Major, C., Bard, E.,  
1673 2010. Hydrology in the Sea of Marmara during the last 23 ka: Implications for timing  
1674 of Black Sea connections and sapropel deposition. *Paleoceanography* 25.  
1675 <https://doi.org/10.1029/2009PA001735>

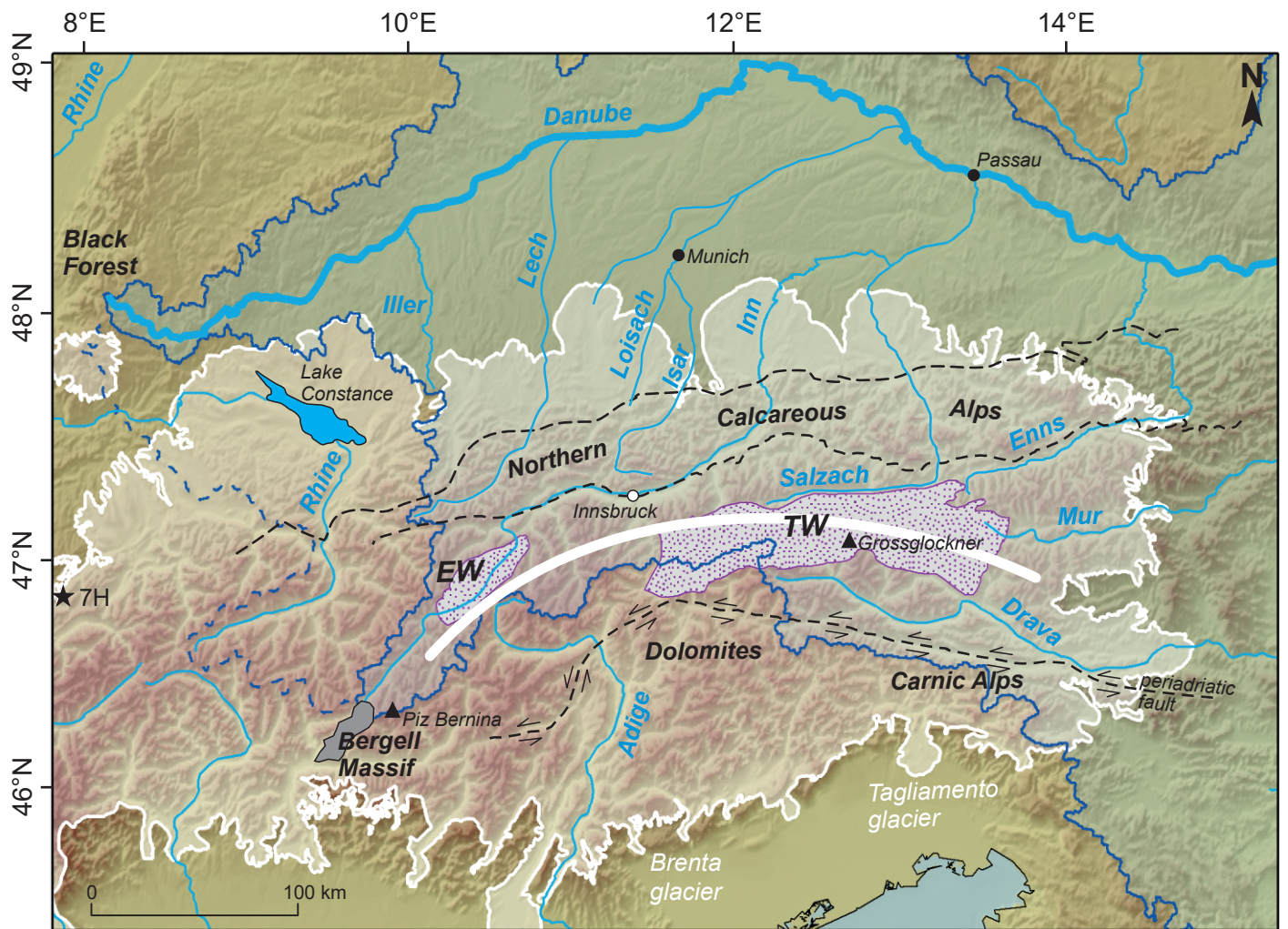
- 1676 Vishnevskaya, V.S., Sedaeva, K.M., 2000. Specific features of Early-Middle Carboniferous  
1677 sedimentation in the southern part of the East European Platform. *Lithology and*  
1678 *Mineral Resources* 35, 455–465.
- 1679 Wang, Y. J., Cheng, H., Edwards, R. L., An, Z. S., Wu, J. Y., Shen, C. C., Dorale, J. A., 2001.  
1680 A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave,  
1681 China. *Science*, 294 (5550), 2345–2348. <https://doi.org/10.1126/science.1064618>
- 1682 Wegwerth, A., Kaiser, J., Dellwig, O., Shumilovskikh, L. S., Nowaczyk, N. R., Arz, H. W.,  
1683 2016. Northern hemisphere climate control on the environmental dynamics in the  
1684 glacial Black Sea 'Lake'. *Quat. Sci.Rev.* 135, 41–53.  
1685 <https://doi.org/10.1016/j.quascirev.2016.01.016>
- 1686 Wirsig, C., Zasadni, J., Ivy-Ochs, S., Christl, M., Kober, F., Schlüchter, C., 2016. A  
1687 deglaciation model of the Oberhasli, Switzerland. *J. Quat. Sci.* 31, 46–59.  
1688 <https://doi.org/10.1002/jqs.2831>
- 1689 Wirth, S.B., Glur, L., Gilli, A., Anselmetti, F.S., 2013. Holocene flood frequency across the  
1690 Central Alps–solar forcing and evidence for variations in North Atlantic atmospheric  
1691 circulation. *Quat. Sci. Rev.* 80, 112–128.  
1692 <https://doi.org/10.1016/j.quascirev.2013.09.002>
- 1693 Wong, H., Panin, N., Dinu, C., Georgescu, P., Rahn, C., 1994. Morphology and post-  
1694 Chaudian (Late Pleistocene) evolution of the submarine Danube fan complex. *Terra*  
1695 *Nova* 6, 502–511. <https://doi.org/10.1111/j.1365-3121.1994.tb00894.x>
- 1696 Wu, H., Guiot, J., Brewer, S., Guo, Z., 2007. Climatic changes in Eurasia and Africa at the  
1697 last glacial maximum and mid-Holocene: reconstruction from pollen data using  
1698 inverse vegetation modelling. *Climate Dynamics* 29, 211–229.  
1699 <https://doi.org/10.1007/s00382-007-0231-3>
- 1700 Yanchilina, A.G., Ryan, W.B.F., McManus, J.F., Dimitrov, P., Dimitrov, D., Slavova, K.,  
1701 Filipova-Marinova, M., 2017. Compilation of geophysical, geochronological, and  
1702 geochemical evidence indicates a rapid Mediterranean-derived submergence of the  
1703 Black Sea's shelf and subsequent substantial salinification in the early Holocene. *Mar.*  
1704 *Geol.* 383, 14–34. <https://doi.org/10.1016/j.margeo.2016.11.001>
- 1705 Zaragosi, S., Bourillet, J.-F., Eynaud, F., Toucanne, S., Denhard, B., Van Toer, A., Lanfumey,  
1706 V., 2006. The impact of the last European deglaciation on the deep-sea turbidite  
1707 systems of the Celtic-Armorican margin (Bay of Biscay). *Geo-Mar. Lett.* 26, 317–329.  
1708 <https://doi.org/10.1007/s00367-006-0048-9>
- 1709 Zaragosi, S., Eynaud, F., Pujol, C., Auffret, G.A., Turon, J.-L., Garlan, T., 2001. Initiation of  
1710 the European deglaciation as recorded in the northwestern Bay of Biscay slope  
1711 environments (Meriadzek Terrace and Trevelyan Escarpment): a multi-proxy  
1712 approach. *Earth. Planet. Sci. Lett.* 188, 493–507. [https://doi.org/10.1016/S0012-](https://doi.org/10.1016/S0012-821X(01)00332-6)  
1713 [821X\(01\)00332-6](https://doi.org/10.1016/S0012-821X(01)00332-6)



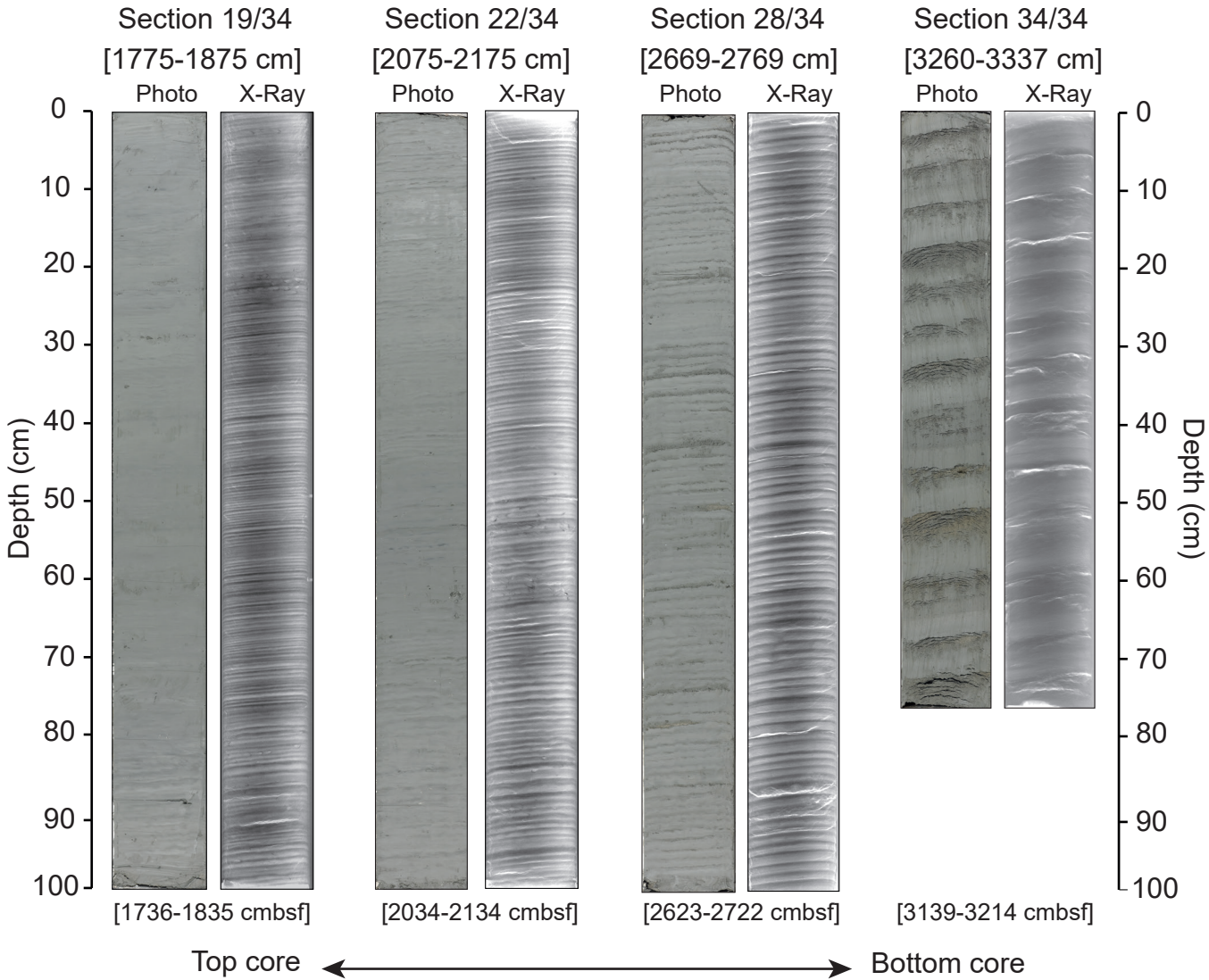


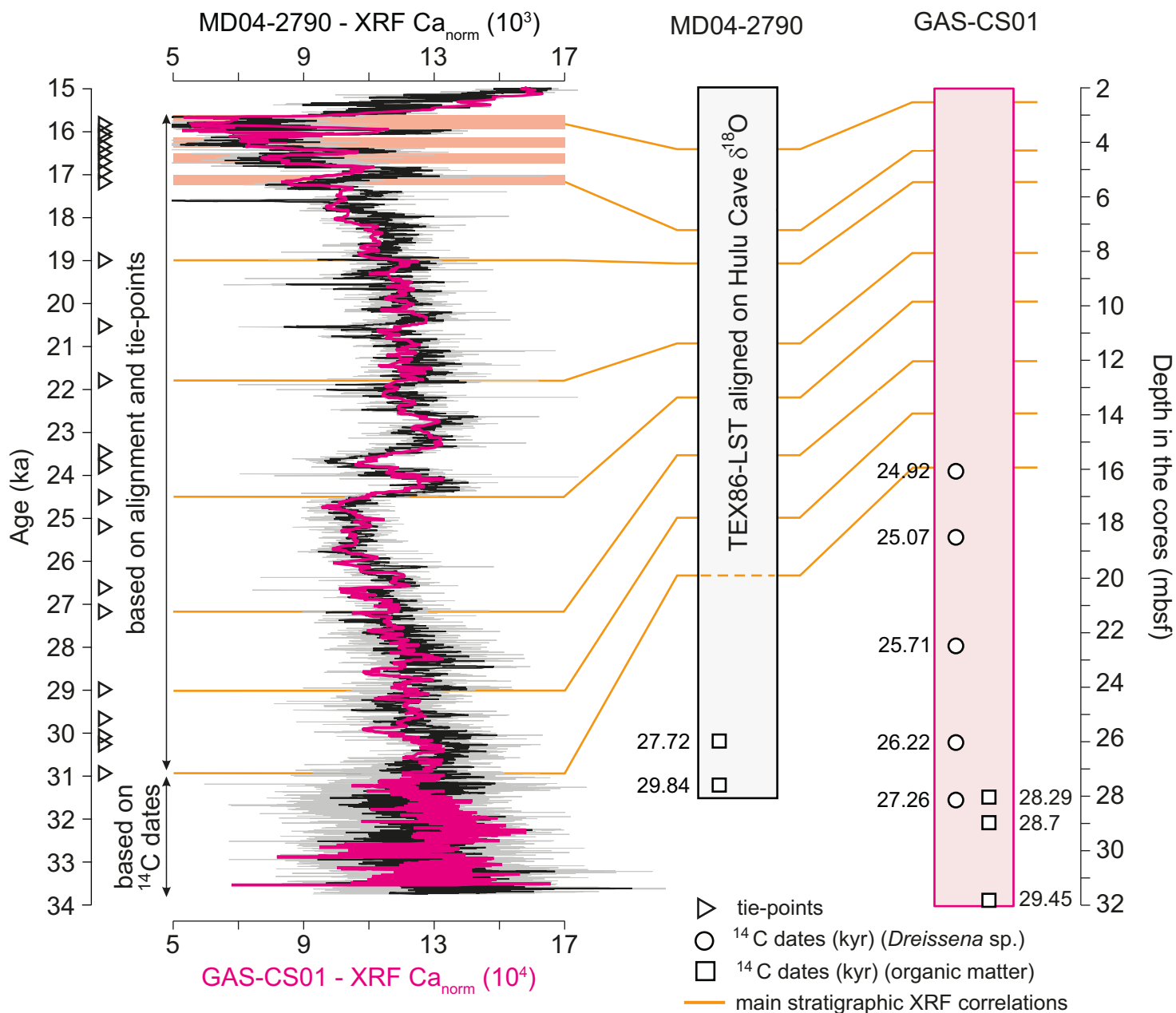


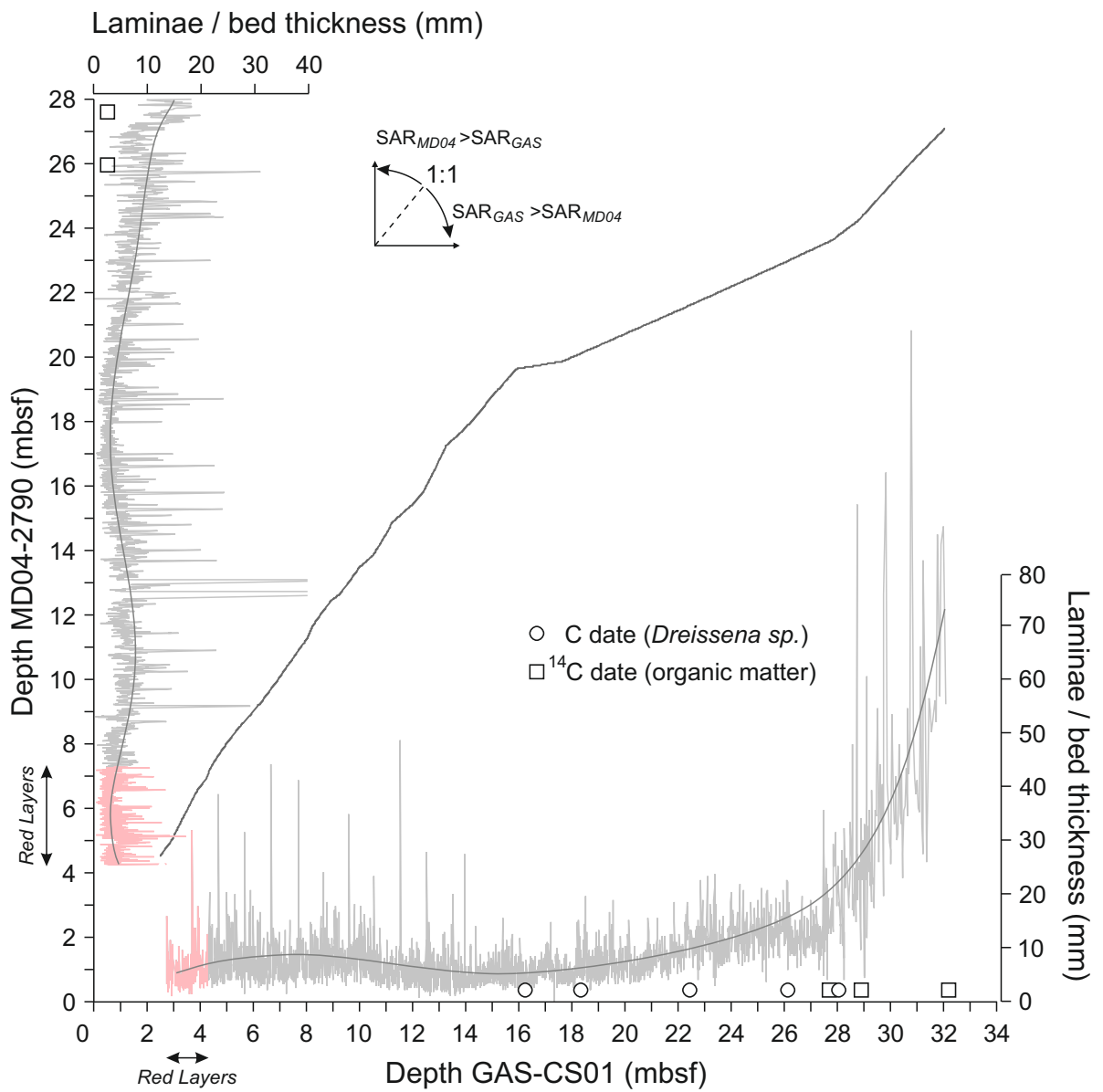


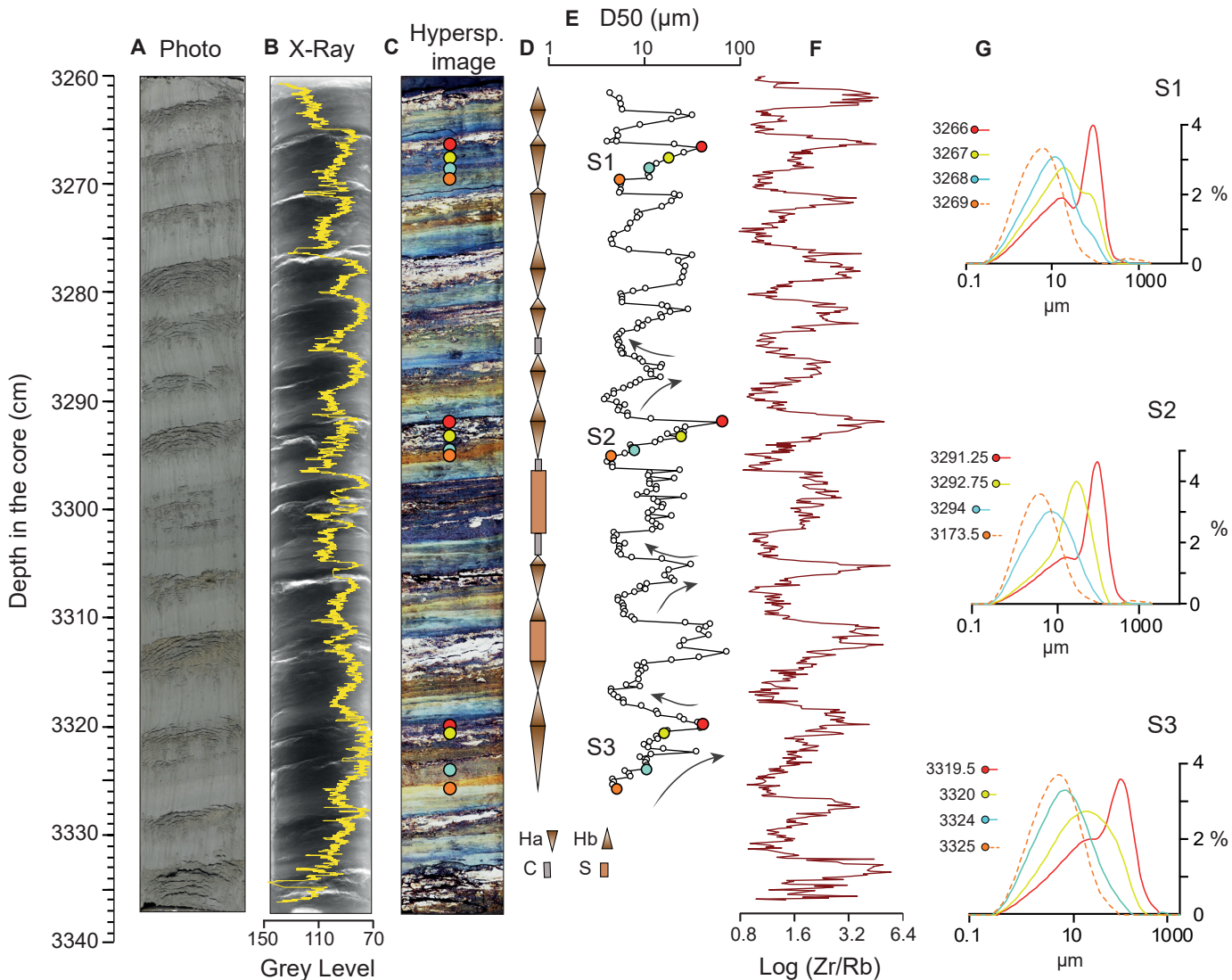


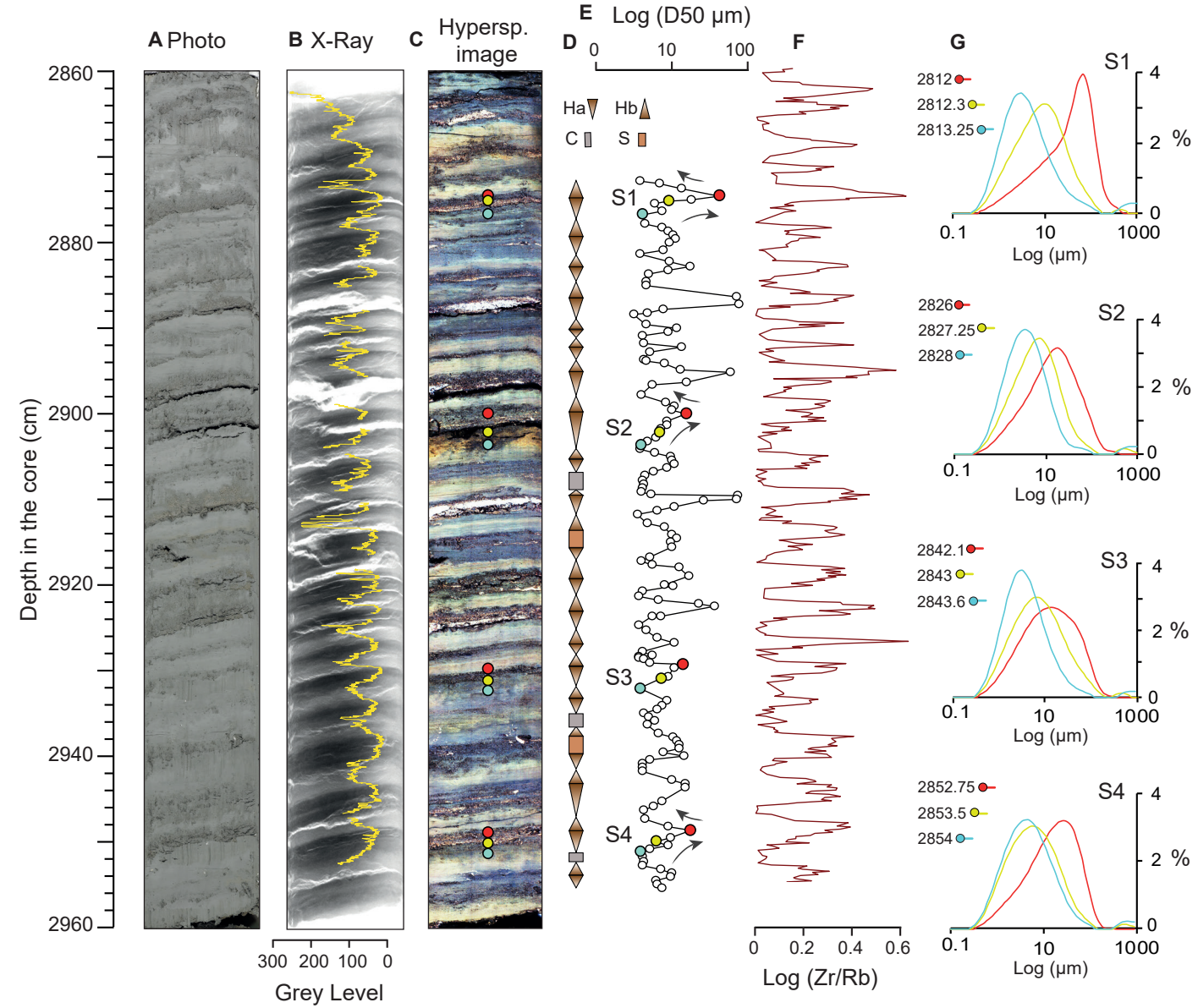




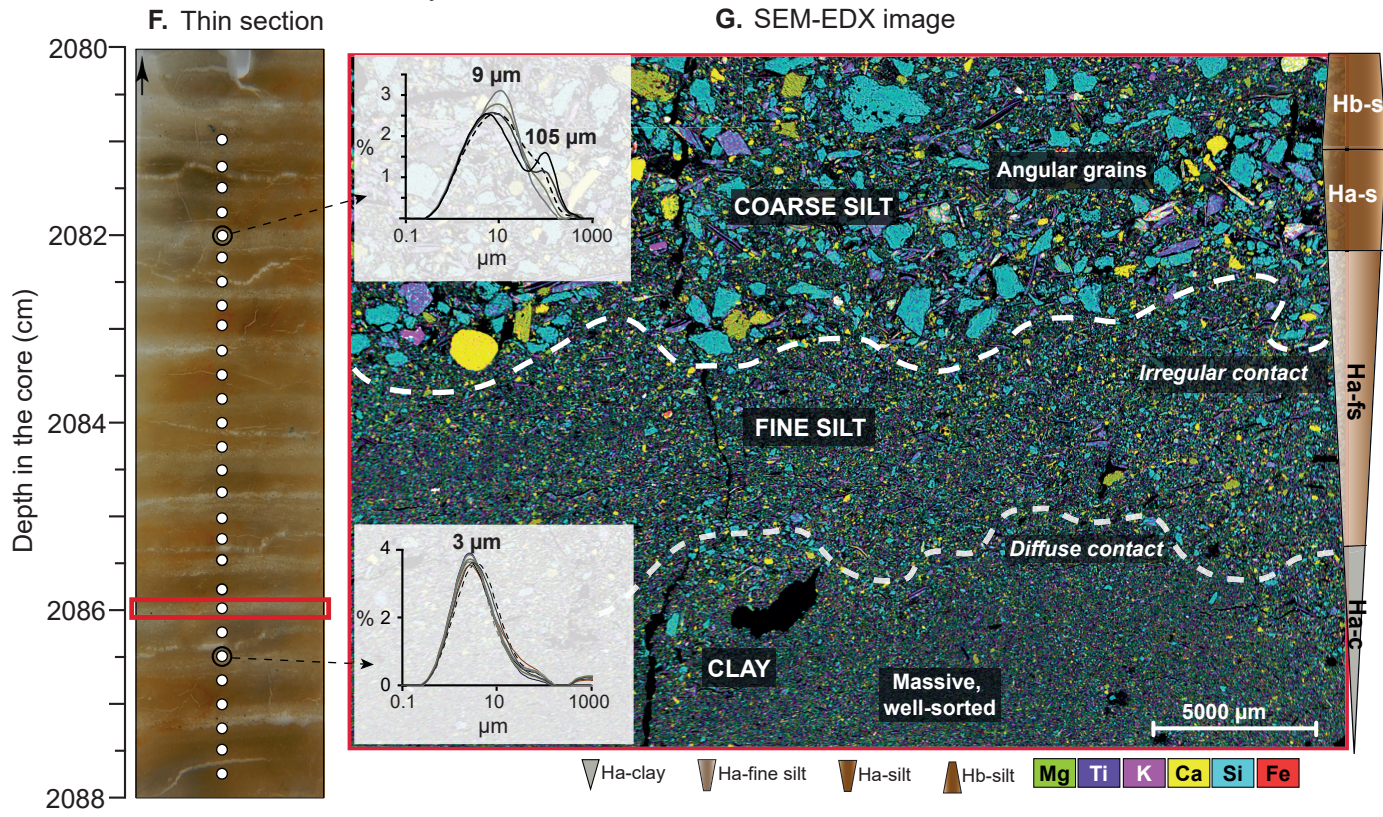
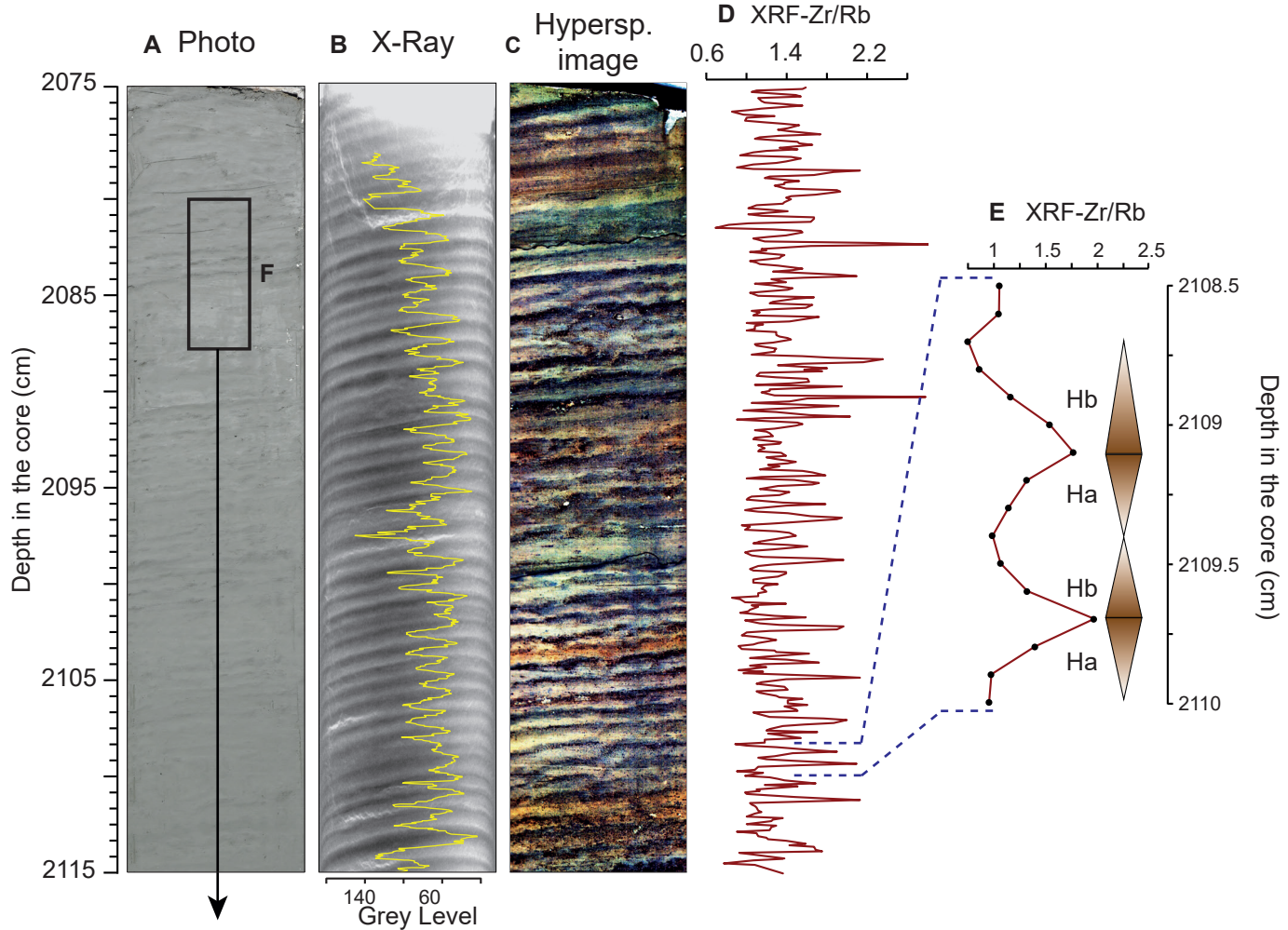


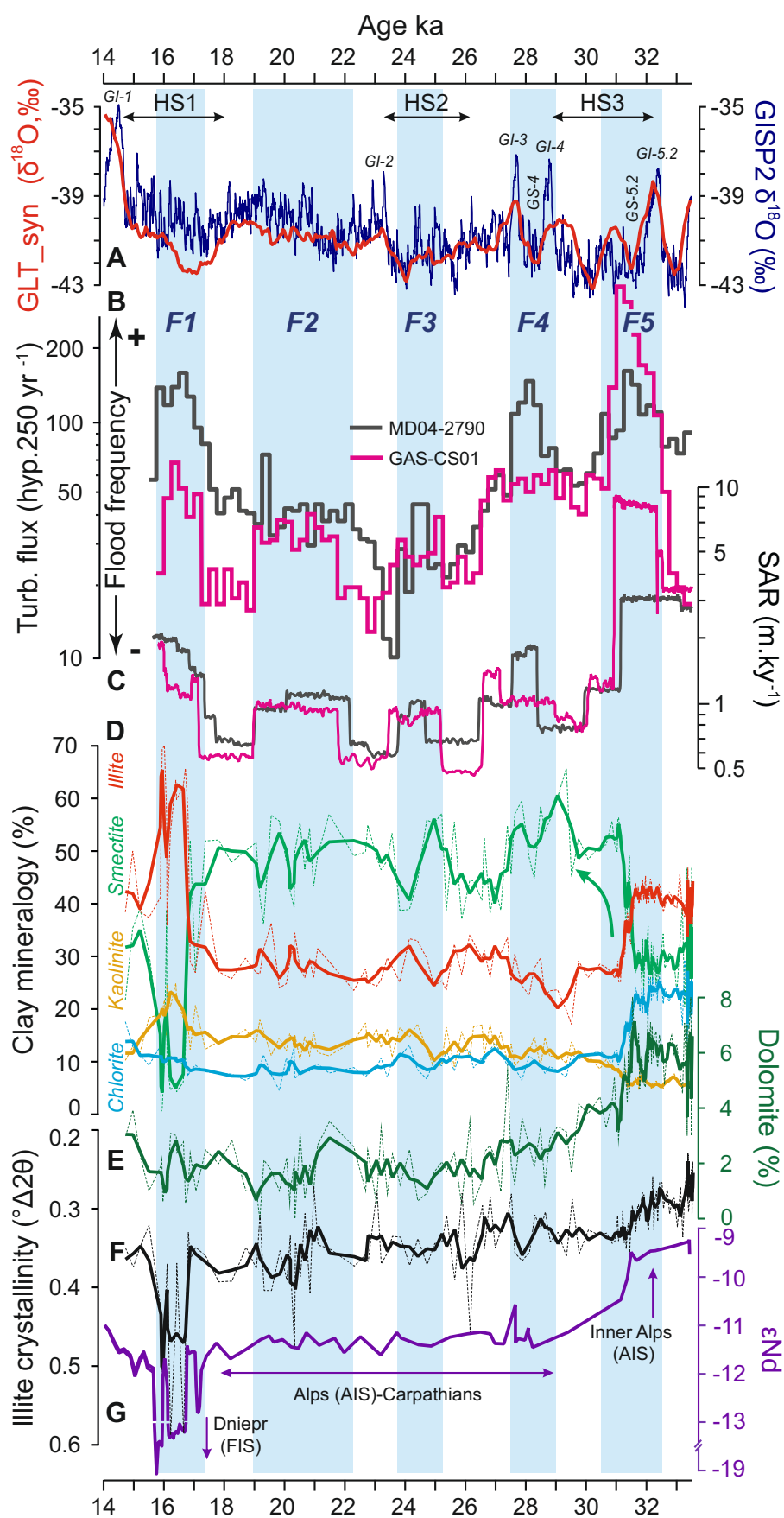




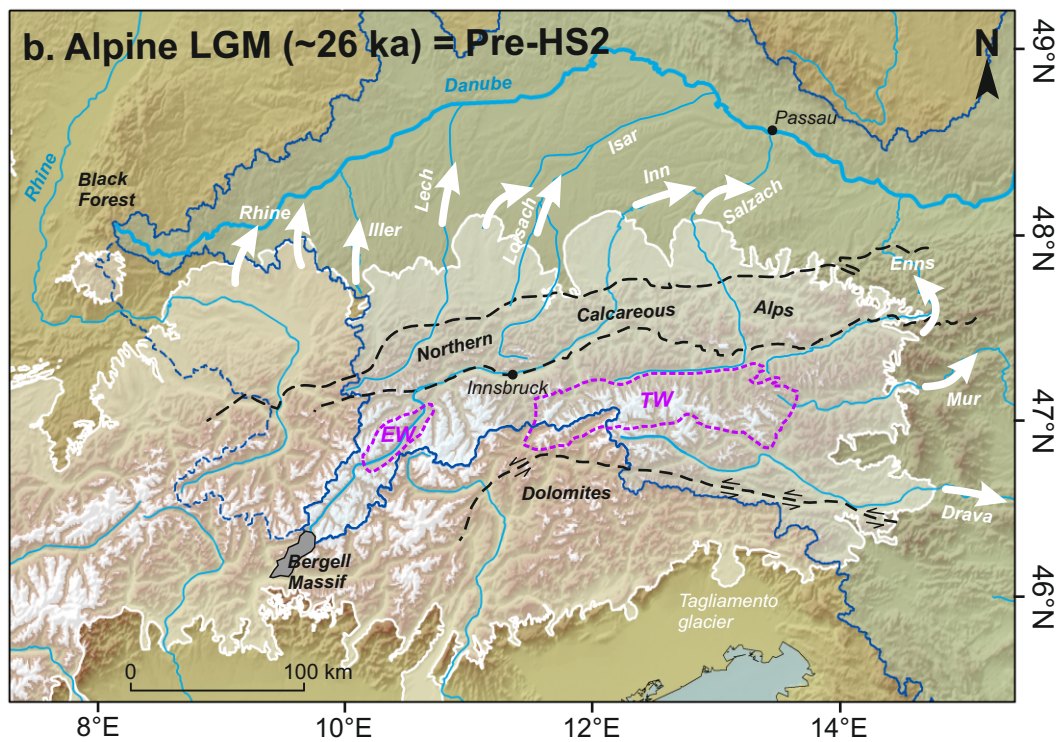
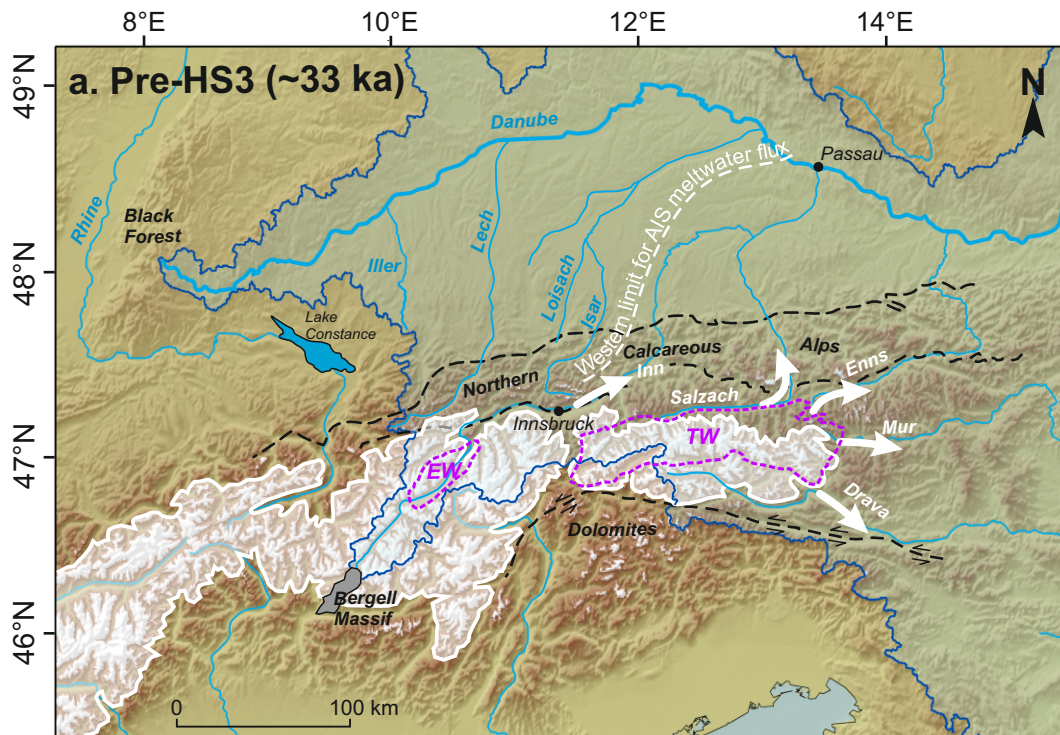


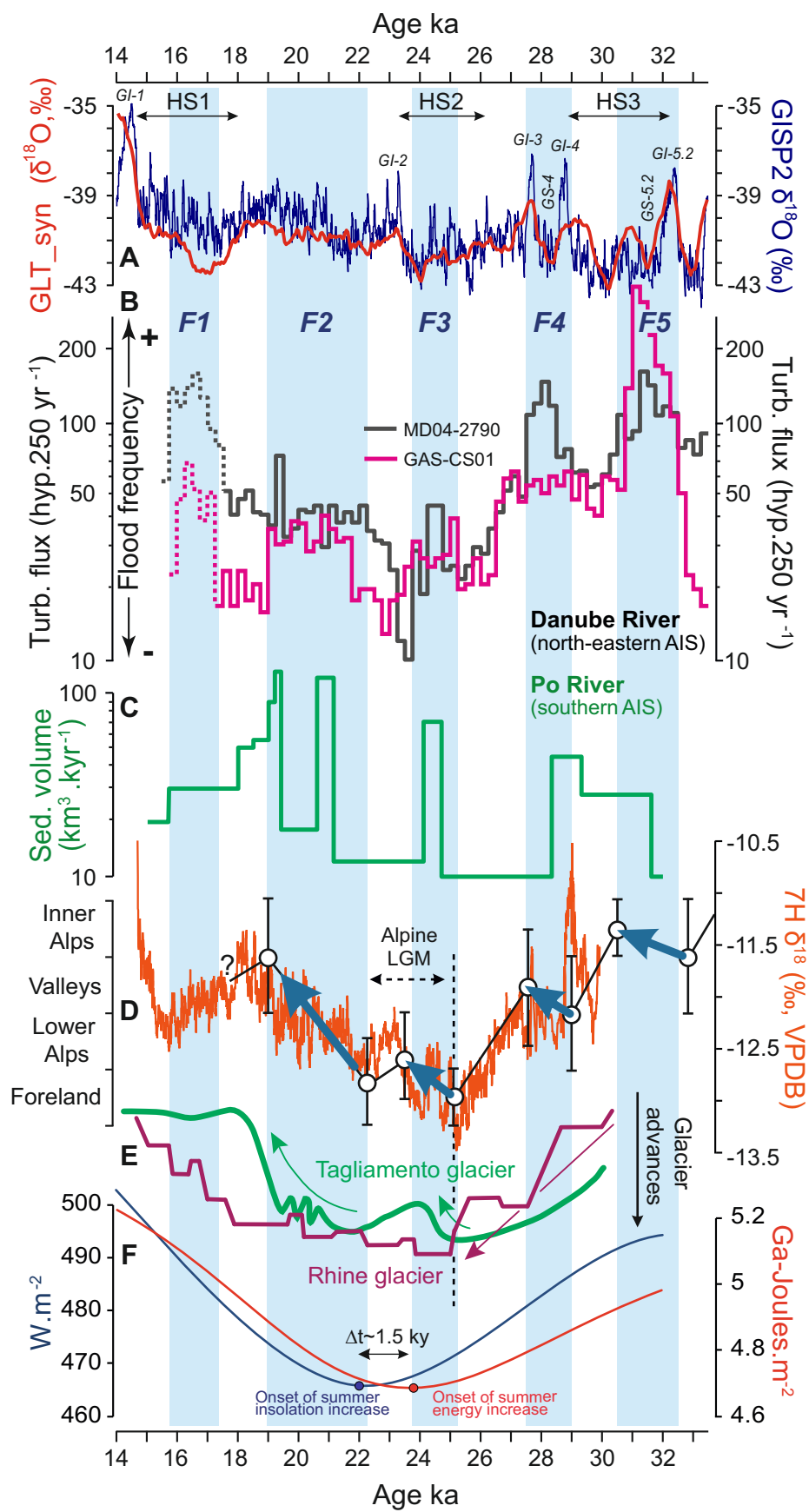












Core Label	Lat. N	Long. E	Water depth (m)	Length (m)	Year	Cruise	IGSN
<b>GAS-CS01</b>	44,0865	30,7993	240	33,4	2015	GHASS	<a href="http://igsn.org/BFBGX-127384">http://igsn.org/BFBGX-127384</a>
<b>MD04-2790</b>	44,2219	30,9935	362	30,3	2004	ASSEMBLAGE-1	<a href="http://igsn.org/BFBGX-88381">http://igsn.org/BFBGX-88381</a>

Core ID	Depth (cm) <sup>a</sup>	Depth (cmbsf) <sup>b</sup>	Lab. Number	Sample type	<sup>14</sup> C age (yr BP)	error (1σ)
<u>GAS-CS01</u>						
GAS-CS01	1651	1614	Beta-440157	<i>Dreissena</i> sp.	24,920	120
GAS-CS01	1870	1831	Beta-440158	<i>Dreissena</i> sp.	25,070	120
GAS-CS01	2280	2238	Beta-463971	<i>Dreissena</i> sp.	25,710	110
GAS-CS01	2650	2604	Beta-463972	<i>Dreissena</i> sp.	26,220	110
GAS-CS01	2851	2796	Beta-479030	Organic matter	28,290	120
GAS-CS01	2853	2798	Beta-440160	<i>Dreissena</i> sp.	27,260	110
GAS-CS01	2955	2882	Beta-463973	Organic matter	28,700	140
GAS-CS01	3336	3214	Beta-479031	Organic matter	29,450	140
<u>MD04-2790</u>						
MD04-2790	2835	2600	Beta-477223	Organic matter	27,720	130
MD04-2790	3001	2765	Beta-463974	Organic matter	29,840	150

<sup>a</sup> Raw (*i.e.* uncorrected) depth in cores

<sup>b</sup> Corrected depth in cores (*i.e.* after removal of gaps), in cm below sea floor (cmbsf)

River	Drainage area (km <sup>2</sup> )	Lat. N	Long. E	ICr (°Δ2θ)	Smectite (%)	Illite 2M1 (%)	Kaolinite (%)	Chlorite (%)	Dolomite (%)
<b>Danube Tributaires (n=12)</b>									
<i>Alps (n=5)</i>				† 0.26 ± 0.04	9.2 ± 15.4	51.2 ± 13.3	4.4 ± 1.9	28.4 ± 3.8	11.2 ± 4.9
				‡ 0.28 ± 0.04	12.2 ± 16.2	44.7 ± 12.6	5.6 ± 2.2	27.8 ± 3.9	9.6 ± 5.8
Iller	26,130	48,36232	10,00110	0,34	40	34	5	21	17
Inn	26,128	47,78181	12,12609	0,28	0	38	3	29	6
Enns	6,185	48,21270	14,48660	0,23	3	65	3	30	17
upper Drava	40,087	46,60046	13,89860	0,2	0	66	3	31	10
lower Drava	40,087	45,69050	18,41658	0,26	3	53	8	31	6
<i>Dinarides (n=2)</i>				† 0.3 ± 0.04	25.5 ± 14.5	44.5 ± 9.5	17 ± 0	14 ± 5	1 ± 1
				‡ 0.28 ± 0.04	19.2 ± 15.8	48.7 ± 10.4	17 ± 0	16.2 ± 5.5	1.4 ± 1.1
Sava	95,793	44,72358	20,30960	0,26	11	54	17	19	2
Velika-Morava	37,571	44,60580	21,08620	0,34	40	35	17	9	0
<i>Carpathians (n=5)</i>				† 0.3 ± 0	44.2 ± 25.5	34.4 ± 15	11.8 ± 7.2	9.4 ± 4.5	0.8 ± 0.8
				‡ 0.3 ± 0	55.6 ± 28	27.6 ± 16.4	8.6 ± 7.9	7.7 ± 4.8	0.3 ± 0.9
Morava	27,267	48,19118	16,97450	0,3	32	38	21	9	0
Tisza	15,609	46,14687	20,06457	0,3	55	28	8	8	0
Siret	46,289	45,39911	28,01163	0,3	66	22	5	7	1
Prut	28,568	45,47172	28,19500	0,3	68	22	5	5	1
Vah	1,966	47,91403	18,01211	0,3	0	62	20	18	2
<i>Alps + Dinarides + Carpathians (351,593km<sup>2</sup>)</i>				‡ 0.28 ± 0.01	29.6 ± 19.1	40.4 ± 9.2	10.9 ± 4.9	16.6 ± 8.3	3.3 ± 4.2
<b>Danube River (n=6)</b>				† 0.3 ± 0.05	29.8 ± 17.5	39.6 ± 7.8	11 ± 5.5	19.5 ± 7.1	5.2 ± 1
Passau	801,093	48,59673	13,38283	0,39	3	41	23	33	0
Upstream Tisza	801,093	45,16503	20,10953	0,28	24	46	8	22	6
Upstream Sava	801,093	44,92347	20,32111	0,33	41	35	9	16	6
Downstream Sava	801,093	44,66575	20,78006	0,27	15	52	11	21	4
Moldava-Noua	801,093	44,69287	21,66701	0,29	42	37	7	14	4
Braila	801,093	45,32294	28,00246	0,25	54	27	8	11	6