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EVIDENCE OF MULTIPLE THERMOKARST EVENTS IN NORTHEASTERN FRANCE AND SOUTHERN BELGIUM DURING THE TWO LAST GLACIATIONS. A DISCUSSION ON 'FEATURES CAUSED BY GROUND ICE GROWTH AND DECAY IN LATE PLEISTOCENE FLUVIAL DEPOSITS, PARIS BASIN, FRANCE' (BERTRAN ET AL., 2018)

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KEYWORDS: Thermokarst ; Frost mounds ; Orbital forcing ; Weichselian ; Saalian ; Dansgaerd-Oeschger events

ABSTRACT

The past thermokarst activities in valleys of northern France and Belgium covered the Upper Weichselian and the Upper Saalian periods. To develop in western Europe, thermokarst first requires an accumulation of ground ice close to the surface progressively stored along the glacial-time permafrost aggradation: it is regionally uncommon during the early glacials, especially on the plateau, but frequent during the Upper Pleniglacials in valleys. These features mostly relate to various frost mounds created by injection and segregated ices. The role of ice wedges is really very limited in this zone of southern extent of the European palaeo-permafrost on plateau and terraces. Thermokarst events are mostly susceptible to occur during the coldest part of the glacial. With a more progressive warming or a retrogressive thermokarst triggered by erosion, as in Arctic today, deformations are more gradual, in direct relation with the rheological properties of the sediments and usually local drainage. They are in concurrence with the vegetation dynamic that will limit its expression. Thermokarst events are in first order orbitally forced under control of a maximum insolation and a minimum in precession, as well as during the Weichselian and the Saalian. They are moreover, associated with abrupt warming transmitted by Dansgaerd Oeschger events. Snowiness and mild winter temperatures are probably the main triggers for thermokarst activity as of today. Other events can be triggered by solar activity as at 20 ka or perhaps enhanced by major ash splay as during the MIS 6b Zeiffen interstadial. Thermokarst events are usually followed on the continent by a reorganization of the rivers from braided to meandering systems.

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1. Introduction

Traces of thermokarst are rarely preserved in geology, moreover if they are older than the last Glaciation. Periglacial features are well preserved on chalky landscapes of Europe, from Britain (Ballantyne and Harris, 1993) to southern Poland (Jahn, 1970). Chalk and soft limestones are porous and highly frost susceptible rocks (Lautridou et al., 1986), indirectly allowing the accumulation of ice masses or of segregated ice in the frozen grounds. As past periglacial activity implies deep freezing and/or permafrost extent, we may expect to find evidence for past thermokarst in most chalky regions. It seems to be the same for such regions located south of the last Glacial limit in northern America (Washburn, 1983; Van Vliet-Lanoe et al., 2018a).

We have synthetized in 2017 the observations in chalky northeastern France in a paper "Quaternary thermokarst and thermal erosion features in northern France: dynamic palaeo-environments" (Boreas). These features were first described by Baize et al, in 2007 following the observations of M. Coulon in 1994. With 'Features caused by ground ice growth and decay in late Pleistocene fluvial deposits, Paris basin, France', the paper of Bertran et al., (2018) reappraises our data at the light of a section in a new quarry. The data presented in this second paper were mostly gathered and provided by a chemist, P. Benoït, in quarries at Sauvage (Aube River), Marcilly (Seine River), and also Gourgançon (Maurienne River) (Fig. 1). Bertran's paper is based on literature and digital data bases (Bertran et al., 2013; Andrieux et al., 2016), created on mostly undated observations for the north of France. It presents three main problems: the morphology and stratigraphical interpretation of the Fy terrace in the eastern basin of Paris, the dating of the units and in fine, the attribution of the deformation to a peculiar type of thermokarst and to other periglacial processes.

In this review, on the base of the paper of Bertran et al. (2018) and our data (2017), we shall discuss first the dating of the deformed host sediment within the regional context. We shall, in a second step, run a discussion about the ground ice and climate conditions responsible for such deformations, on the basis of present-day thermokarst features and from personal observations and published data. The stratigraphic position of the thermokarst events within dated sediments will finally allow, with their environmental context, their interpretation in term of palaeoclimate forcing (such as snowiness or change in precipitation' regime) or other triggers (such as Dansgaerd-Oeschger warming events or solar activity).

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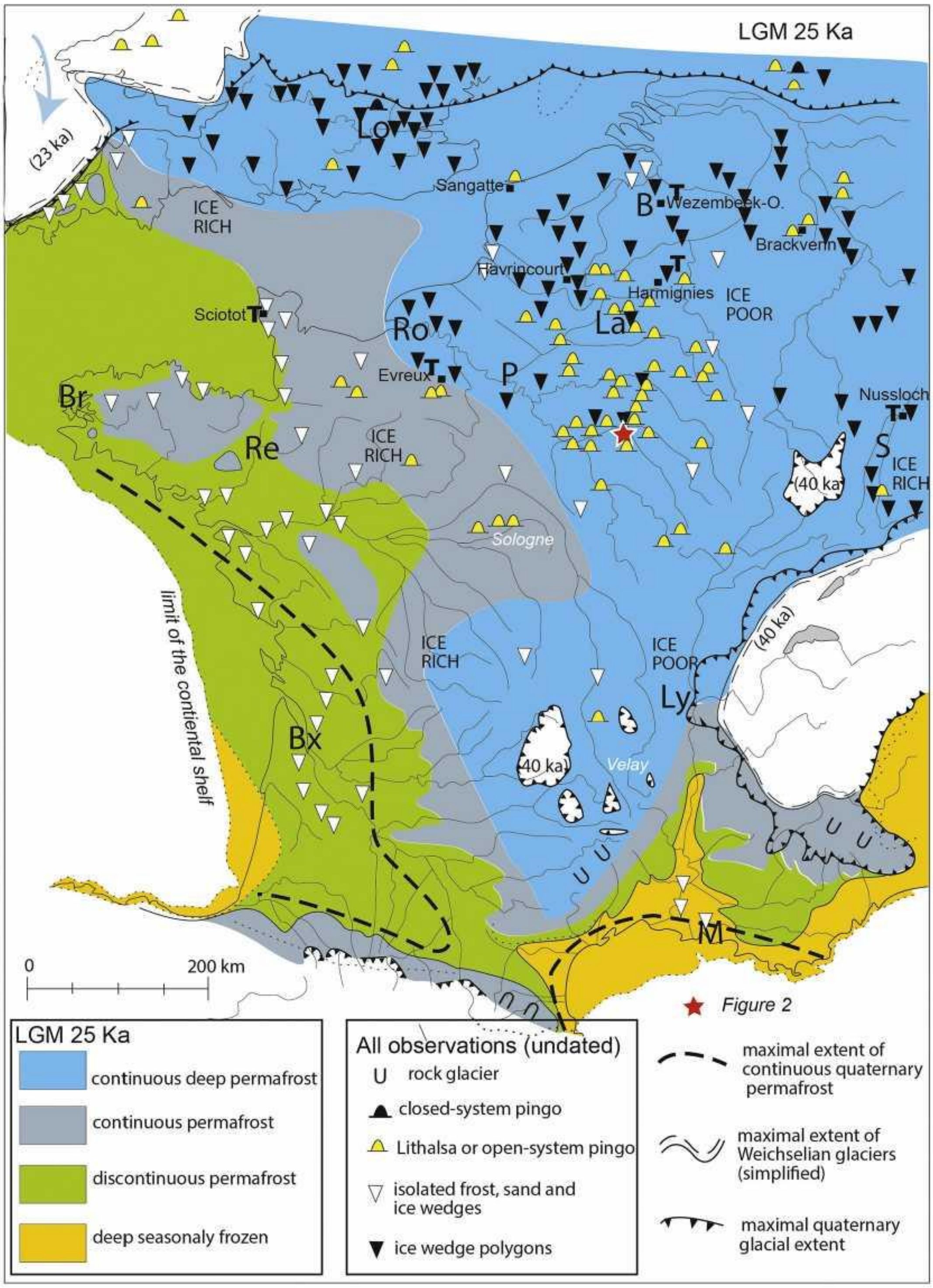
Fig. 1. Permafrost in western Europe, completed from Van Vliet-Lanoe (1989) and Ballantyne and Harris (1993) for UK, with Van Vliet-Lanoe et al., (2018a) and Andrieux et al. (2016). P. Paris; La: Laon, Ro: Rouen; B: Brussel; S: Strasbourg; Lo: London; Ly: Lyon, Br: Brest; Re: Rennes; M: Marseille; Bx: Bordeaux.



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2. Periglacial deformations and permafrost

Frost is an oriented desiccation (Van Vliet-Lanoe, 1985,1988,2014). Cryogenic deformations (cryoturbations) are mostly issued from differential frost heave resulting in differential accumulation of segregated ice in the ground following their grain size composition as initially stressed by Sharp (1942), Dylikowa (1964), and Washburn (1983). It has been demonstrated by experiment (Pissart, 1982) and by micromorphology (Van Vliet-Lanoe, 1985, 1988; Harris and Cook, 1988; Harris, 1990) as observed in the Arctic and Subarctic. It has also been demonstrated by modeling (Rempel et al., 2004; Peterson and Krantz, 2008). Cryoturbation developments exist on permafrost and on seasonal frost domains (Van Vliet-Lanoe, 1985; Harris, 1990). Liquefaction is

exceptional, excepted superficially (b10 cm), often in downslope position. It may occur with vibrations that destroy the cryogenic fabric as in animal trampling or earthquakes (Van Vliet-Lanoe, 1988; Van Vliet- Lanoe et al., 2004) or by mass wasting associated with thermokarst (Murton and French, 1993). Load cast involutions attributed to cryoturbations are always described in zones with active seismicity as the Dutch Rhine grabens (Kasse and Bohncke, 1992; Vandenberghe, 2013) or other actives tectonic zones as the Mackensie delta (Murton and French, 1993), the Baikal graben (Alexeev et al., 2014), and the Gironde estuary located very close to an active shearing zone (Sitzia et al., 2015), with instrumental earthquakes of Mw ≥ 6 (Van Vliet-Lanoe et al., 2018b).

Permafrost is attested in France and Belgium at least from MIS 8 at the level of southern Brittany (Van Vliet-Lanoe, 1989; map in Fig. 1). At the European scale, this map fits the model by Levavasseur et al. (2011), but takes into account relief and some local climate forcing. In northeastern France, permafrost is relatively poor in ice on plateaus, valley bottom excepted (Van Vliet-Lanoe, 1989, 2014; Fig. 1), by the recurrent occurrence of metric pattern grounds and sorted stripes at the surface of the chalk, as already stressed by Tricart and Cailleux (1967) or by Michel (1973) in French Champagne (Fig. 2) and by Ballantyne and Harris (1993) in chalky England.

Palaeo-permafrost is also recorded by several generations of ice wedge casts, usually common north of the Seine River (Michel, 1973; Haesaerts and Van Vliet-Lanoe, 1981; Lautridou and Somme, 1981; Antoine et al., 2005, Fig. 1). Regional Weichselian permafrost has been the object of four industrial reports for the Andra and published modellings (Lebret et al., 1994, 1996; Andrieu-Ponel et al., 2007; Holmen et al., 2011; Grenier et al., 2013).

Ground palaeo-thermokarst has been rarely observed in sections, often described as subsiding depression filled by younger sediments, set under milder climate conditions (e.g., Van Huissteden, 1990).

To develop in western Europe, palaeo-thermokarst requires an accumulation of ground ice close to the surface, uncommon during the early glacial (especially on plateaus) but frequent in valleys during the Upper Pleniglacial, progressively accumulated along the Weichselian permafrost aggradation.

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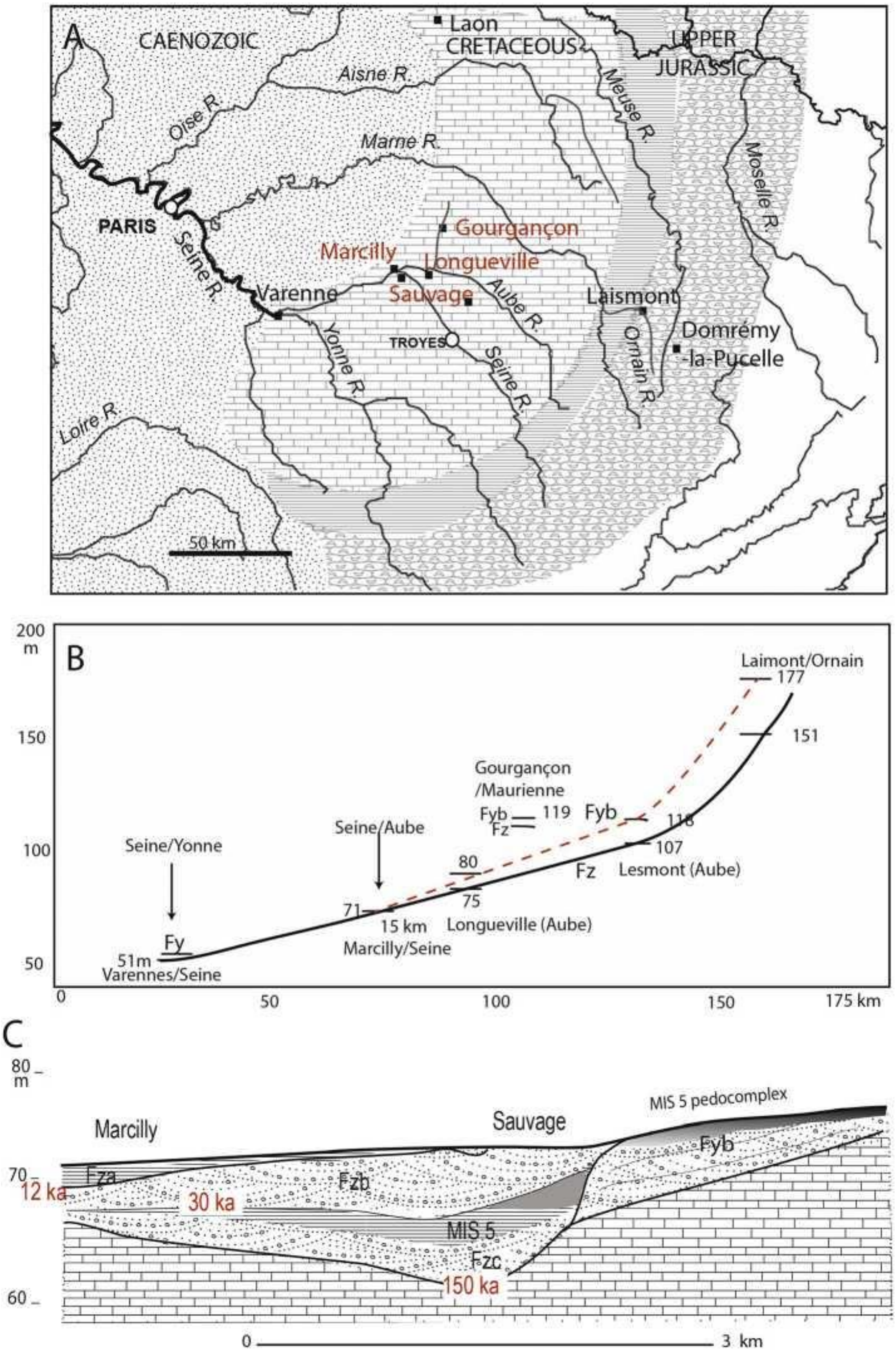
Fig. 2



A) Location of the sites ojn a simplified geological map. B) Geometry of the Fyb terrace related to the recent on (Fx) in the Ornain-Aube-Seine Rivers. B) Sketch of the valley infilling at the Seine-Aube Rivers confluence, based on BRGM corings and field data.

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3. Regional setting

To understand the thermokarst records in the alluvia of the eastern Paris basin, we needed to take into account the geomorphological setting of the deposits within their geological context. Eastern Paris basin is a monoclinal geological structure, somewhat deformed by Paleogene and Cretaceous tectonics (e.g., Guillocheau et al., 2000; Pisapia et al., 2018), but currently, aseismic from earthquake catalogs (ReNaSS: http://renass.u-strasbg.fr/, SISFRANCE: http://www.sisfrance.net/ and SiHex: Cara et al., 2015) even some very low magnitude events can be triggered by pumping that provides the city of Paris in drinking water. These rocks are associated with highly frost susceptible Cretaceous chalks, with upstream, Jurassic limestones and sandstones (Lautridou et al., 1986). Karst in the regional chalk is rather diffuse and very limited in depth (Jaillet et al., 2004).

4. Data

We have synthesized the data from quarries in the eastern Paris basin that provide very nice records of thermokarst collapses in alluvial calcareous and chalky sands and gravels (Van Vliet-Lanoe et al., 2018a). In the upper Seine River watershed, the Longueville/Aube section, the Gourgançon/Maurienne section, the Laimont and Domrémy-la-Pucelle quarries (Fig. 1), as they revealed similar patterns in sections. Sauvage quarries were less interesting because of the shallowness of the sections and the end of activity in the quarry. These forms, mostly observed on the Fyb terrace have a circular shape viewed on aerial images (Michel, 1967; Van Vliet-Lanoe et al., 2018a; Bertran et al., 2018). They partly resemble the pitted morphology of the loessic Russian plain (Velichko et al., 1984) related to thermokarst, especially on large ice wedge casts in silty material. They are often associated with normal and reverse fractures (Baize et al., 2007) confused with tectonics (Benoit et al., 2013). These fractures now attributed to thermokarst collapses (Van Vliet- Lanoe et al., 2018a) by comparison with active features, various collapse formations and their peculiar reverse or "bell" shaped faulting (pingo, frost blister, lithalsa, passive glacitectonism, iceberg thermokarst, volcanic crater, karst) and geotechnical experiments (Roche et al. 2011; Poppe et al. 2015). The interpretation of Bertran et al., 2018 (Section 4.3) thus is not new. (Ice) wedge casts features occur, mostly close but slightly above to the valley bottom, although thermal cracking may appears even in the river bed. Bertran et al. (2018) restricts the round thermokarst pounding to ice wedge degradation, but not to the melting of other plausible types of ground ice in valley environments such as open-system hydrolaccolithes, frost blisters, ice sills and lake ice (Van Vliet-Lanoe et al., 2018a). Lithalsa degradation is restricted by Bertran et al. (2018) to a thin "lacustrine" silt deposit about 2 m thick with evidence of ice lensing. These authors believe that these events are limited to the Upper Pleistocene (GI 1 of Rasmussen et al., 2014), although they should not be expressed at the present-days field surface. We have stressed (2017) that they occurred during MIS 6b on the Fyb terrace but that some of the forms were younger or reactivated during the late Weichselian.

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5. Discussion

5.1. DATING AND MEANING OF THE FY TERRACES AND FZ UNITS

One problem in the paper of Bertran et al. (2018) is the dating at Marcilly and Gourgançon and the interpretation of the Fy terrace, considered by these authors as Last Glacial terrace. For the French geological survey (BRGM), the Fy labell for a terrace corresponds to the penultimate interglacial/glacial transition or MIS 6/5 or the lower terrace, Fz correspond to the present valley alluvia, including the MIS-5 to MIS2 sediments. The valleys (Seine, Aube and Maurienne Rivers) are disconnected from the glacio-eustatic signal and have been dated upstream by ESR on quartz (Cojan et al., 2007; Voinchet et al., 2015). The outcropping calcareous rocks are responsible regionally for the production by frost shattering of large splay of stratified slopes and alluvial grèzes (Michel, 1973), forming the Fz and Fy alluvial sheets of the French geological map, forming a paleo-fan with a slope much steeper than the Holocene alluvial plain (Fig. 2A; Van Vliet-Lanoe et al., 2018a). This fan (Fy) is subdivided into two steps from which the lower one get an average age of 164 ± 23 ka (five dates between 178 and 153 ka; Voinchet et al., 2015). The Fyb terrace on the Marne River was incised by the Fza alluvium after the capture of the upper Meuse River ca 150 ka (Ornain/Sault Rivers) and the increase in river discharge (Cojan et al., 2007), modifying the regional base level. Loesses are present on the interfluves but commonly integrated into multiphased sorted stripes, including deformed palaeosols. These loesses are reworked during warming events and integrated into valley alluvial deposits, usually calcareous and sandy to gravelly (Haesaerts and van Vliet-Lanoe, 1981; Jamagne et al., 1981). Interglacial sedimentation is usually silty, often with peat layers. The early glacial reworks the palaeosoils formed at the surface of the terraces and the slopes.

Dating obtained at Marcilly by Bertran et al. (2018) are fair and allow a better understanding of the alluviation topographically down of the Fyb terrace at Sauvage, with complements observed in the Aube valley by Pastre et al. (2000). Crossing the BRGM soil data banks (http://infoterre.brgm.fr/page/banque-sol-bss) and the field data, it is possible to reconstruct a section of the alluvial plain from Sauvage to Marcilly quarries (Fig. 2B). The section at Sauvage covers thus partly at depth the Fyb terrace, but is lapped on by several units with at least the late glacial with rare, clean gravel with small frost wedge or ice vein (Fig. 3) and the Fzb corresponding the upper Weichselian with a net of narrow (b1 m wide) ice wedge cast, on older oxidized sands and gravels material corresponding to the Fxa terrace with late Eemian/ lower Weichselian units, with older ice wedges casts and large, ductile deformations.

Both sections of Marcilly and Sauvage yield a condensed stratigraphy, between 5 and 10 m on the chalky basement, attesting mostly to sediment transit, as expected with pleniglacial braided rivers: Units are commonly b2 m thick. It is a little simplistic to attribute Units 2.1 and 2.2 to thermokarst lakes on ice wedge net. The top of Gourgancon section is marked by a net of frost wedge and cryoturbations cannot be Bölling in age (Greenland GI 1, Rasmussen et al., 2014), nor even Weichselian as stressed by Bertran et al. (2018).

Fig. 3

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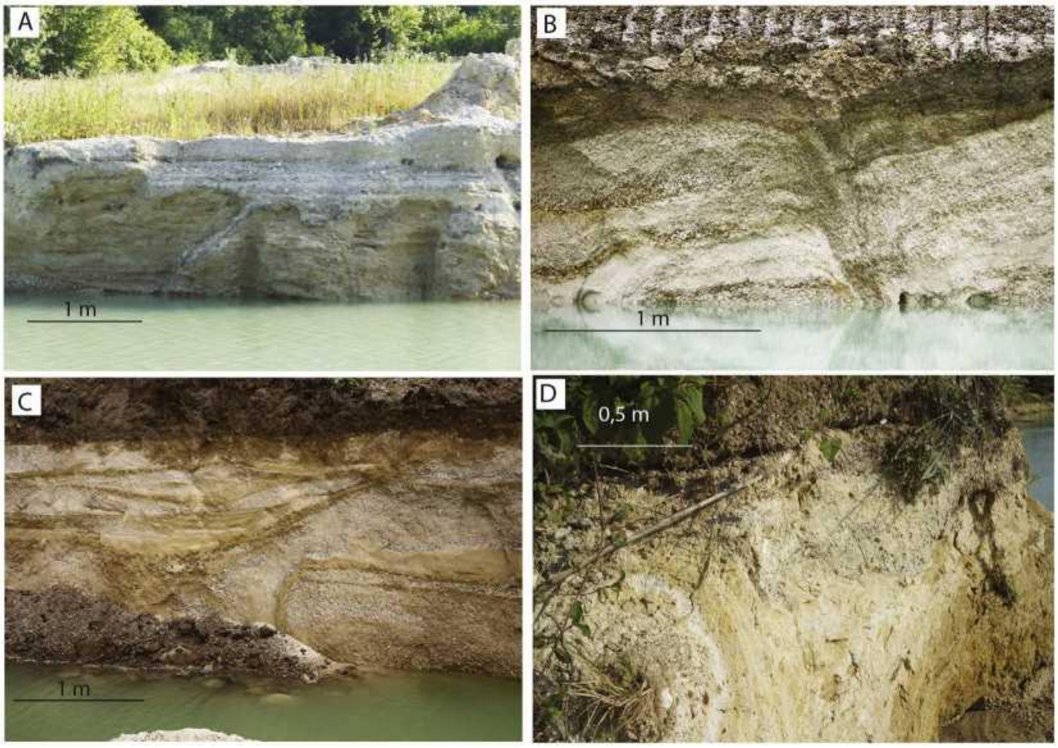

A) Sauvage quarry: Bolling meandering river truncated by reduced Younger Dryas braided river (Fzc) with frost wedges, incised on the oxidized alluvial loam with rafted gravels, B) Sauvage quarry: frost wedge in sandy gravel (photo P. Benoit) C) Sauvage quarry: composite wedge cast, unit 1.6 (photo P. Benoit); D) Sauvage: oldest generation of composite wedge cast with siliceous gravel: post 150 ka: Fza terrace, unit 2.1.

5.1.1. STRATIGRAPHY:

Fzc: Holocene alluvium with peats b10 ka.

Fzb: 1) the Upper Weichselian alluvial unit (30-12 ka)

1.1 A bleached (redox) sandy gravel, calcareous (Pastre et al., 2000), Younger Dryas in age with small frost wedges (Fig. 3) at Sauvage. Sometimes faulted very locally (reverse faults).

1.2 Prograding meander clinoforms (meandering river system, Fig. 4), younger than the main degradation/faulting at Sauvage, presumed Bolling in age, locally truncated by unit 1.1.

1.3 A main deformation phase associated with listric but ductile faulting, prior to the chenalized deposit and collapse of the section analyzed and dated by Bertran et al., 2018). Some reverse faults are clearly younger than 1.3 "cover sands" (Fig. 4)

1.4 Oxidized stratified sand with trough cross stratifications yielding by OSL 16.6 ± 0.9 ka (Shfd 17,101) associated with a wedge cast, deformed by folding ("recumbent fold) (Bertran et al. 2018)

1.5 Oxidized lake silts at Marcilly yields 24,645-24,120 a. cal BP (Beta- 470,451; Bertran et al., 2018), with evidence of segregated ice, deformed by reverse faults

1.6 Oxidized stratified sand and gravels (braided river system)

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Fza: 2) oxidized calcareous gravels and sands including evidence of composite wedges, reworking sandstone gravels at Sauvage quarry.

2.1 Brown clays or alluvium reworking the MIS 5 pedocomplex (BRGM drilling)

2.2 Grey silty loam from the MIS 5 isotopic stages (probably MIS 5e and MIS 5d) (BRGM drilling)

2.3 Alluvial gravels and sands (braided river): End of MIS6 (BRGM drilling)

It also means that the events at Marcilly and Sauvage quarries are much younger than those observed at Longueville/Aube R. and Gourgançon/Maurienne R., as they rework some siliceous (sandstone) gravels from the Jurassic Lorraine. The polygonal net of the Varenne/ Seine R.(Google Earth 2011) is perched 3 to 5 m higher than the valley traces of thermokarst lakes. It represents thus a Fyb (MIS6) terrace of the Seine River, further incised. It cannot be responsible of the thermokarst at the base of Marcilly section, considering the current conditions for active thermal cracking (see Section 5.3). The onset of the next alluvial body, (Fza), formed after the capture of the Sault River by the Ornain River has been dated c 150 ka on the Marne River (Cojan et al., 2007). The dated Yonne River records a similar story (Chaussé et al., 2004).

It means that the Fyb terrace has been formed during the MIS6b part of the Saalian, in a context slightly less cold that the Weichselian MIS2, but probably colder than the permafrost developed in Brittany on north-facing slopes after 150 ka (Van Vliet-Lanoë, 1988, Seidenkrantz et al., 1996) although it developed much earlier in this eastern region.

Seeing the limited loess deposition during the Weichselian in the sector, the preservation of some microrelief of the former thaw lakes suggest for most of them a rather young age, prior to the last main vegetation event and landscape stabilization, the Bolling (Van Vliet-Lanoe et al., 1992; Pastre et al., 2000). The forms described at Marcilly by Bertran et al. (2018) are probably sub-synchronic with the last event at Sauvage quarry: a late reactivation of the thermokarst in the lower Fyb terrace of Clesle/Aube River (Van Vliet-Lanoe et al., 2018a) as at Laismont/Aube River and Domremy-la-Pucelle quarries. The fresh forms in the alluvial plain, to the North of Laon city are probably also of the same generation (Fig. 1; Van Vliet-Lanoe et al., 2018a).

5.2. ICE SEGREGATION IN FROZEN MOUNDS AND RELATED THERMOKARST

Ice segregation in lenses is usually the commonest feature observed in fine-grained soils and sediments subject to freezing (Shumskii, 1964; Arakawa, 1966; Van Vliet-Lanoe, 1985). It results of the migration of pore water from the unfrozen substratum under thermal gradient. It is thus the signature of frost penetration as in the "lacustrine" silty loam alluvium of Marcilly section (Bertran et al., 2018) but, with, at that time, a drained substratum, despite its position in valley bottom. It signs for us cold and dry conditions with very limited snow cover, favourable for thermal cracking.

A peculiar type of ice lenses occurs in unconsolidated, water saturated fine sediments: reticulate ice veins (Mackay, 1998; Kudryavtsev, 1978). This type also exists in complex palsa formed on unconsolidated sediments (Svensson, 1964), in lithalsa in Altaï mountains (Iwahana et al., 2012), in the infill of thermokarst depressions as in the permafrost tunnel of Fairbanks (Alaska). The rate of freezing is rather rapid as observed by experiment (Pissart, 1970).

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The traces of this type of ice have been observed in Gourgançon section (Van Vliet-Lanoe et al., 2018a). It occurred apparently seasonally with the colluvial infill of a thermokarst depression (Figs. 4 and 5; unit Gou 2). The large brittle fracturation occurred after a first melting of the reticulate ice (bacterial iron staining) signing a late age for the main collapse event, independently of the reticulate ice traces. This fracturation affects the onset of units Gou 3 (channelized river), mostly unit Gou 2 and also the basal unit. It is thus a very late event, sub-synchronic with the development ofa meandering river (interstadial). Unit Gou 1 is characterized by ductile deformation, the probable signature of a melting in subaquatic conditions with evidence (faulted sides) of a temporary but collapsed small frost blister, the 'liquefaction' of

Bertran et al. (2018). The formation of lithalsa is probably the main source of thermokarst deformation in the silty calcareous mud, derived from a marly chalk into the Gourgançon section. We observed four successive events of thermokarst, from accumulation of segregated ice and the resulting sedimentary record (Van Vliet-Lanoe et al., 2018a).

Itis not the case for Marcilly, Sauvage and Longueville sections, dominated by sandy-gravelly facies, with interstitial ice and most probably injected ice, in form of sill or of hydrolaccolithes (frost blister or opensystem pingo; Pollard, 1991; Pollard and Everdingen, 1992; Van Everdingen, 1982; Van Vliet-Lanoe et al., 2018a). The major brittle faulting affects the section from the basement close to the top: the major event is thus late in both Weichselian and Saalian gravelly deposits (Figs. 4 and 5).

The clustered but collapsed Holocene open-system pingos of Trail Island, Greenland (Worsley and Gurney, 1996; inherited from the thermal Optimum, 8-6 ka BP) yield today about -10 °C MAT, but their size fit our circular traces (50 to 100 m in diameter). From our observations in central Iceland, Holocene lithalsa in loess derived alluvium require ca -2 °C MAT (600 m in elevation, discontinuous permafrost, high snowiness) and in the same region, open-system pingo on pumice ca -4° MAT (800 m in elevation; Fig. 6). All these forms are degrading due to the recent warming. In the Advent valley (Svalbard), the open system pingo's were destabilized by thaw with the recent shift in MAT from -6 °C in 1984 to -4 °C in 2004 in relation to the recent warming. (http://www. climate4you.com/SvalbardTemperatureSince1912.htm).

Nevertheless, into the Saalian, at Longueville, smaller palaeostructures were associated with pounding and formed at least 4 times as at Gourgançon (Van Vliet-Lanoe et al., 2017). It is not possible to say if they are exactly synchronic. From this, we may suppose that the four events within MIS 6 represent consecutive cooling steps within the same warming event. It seems also the case for both Marcilly and Sauvage sections, with evidence of synsedimentary complex Weichselian events, in association with local wedges casts.

***Fig.4**. Sections of Gourgançon, completed from Van Vliet-Lanoë et al. (2017) with a picture from P. Benoit for Thermokarst I (frost blister at Gourgançon), Longueville and Sauvage West. Longueville and Gourgançon at the same vertical and horizontal scale.*

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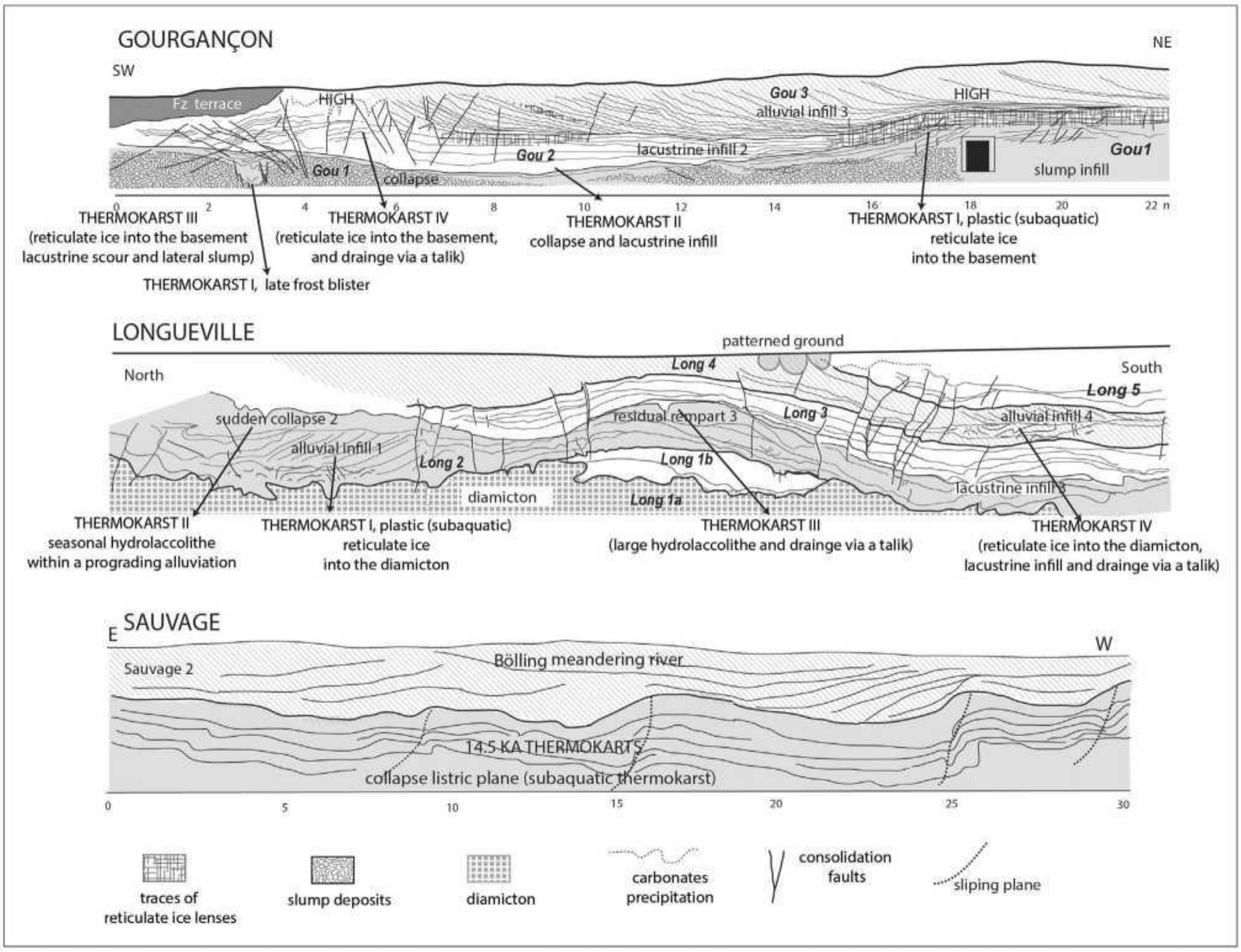

5.3. THERMOKARST ON ICE WEDGES

Nowaday, thermal contraction also needs abrupt cooling larger than 10 °C within a few hours to develop (Lachenbruch, 1966). It is not specific of permafrost. This means that the thermal contraction will only develop in peculiar conditions: in a temporarily ice rich (Svensson, 1978) or permanently frozen ground (Burn, 1990). Thermokarst lake formation generally starts with the coalescence of polygonal and (or) ice-wedge elongated pounds overlying melting large ice-wedge (N6m wide) networks (Czudek and Demek, 1970). In ice dominant material like the Yedoma complex (N60% ice) in Russia (Solovyev, 1973; Murton, 1996; Morgenstern et al., 2013; Schirrmeister et al., 2017), circular lake develop as alas on ≥6 m wide ice wedges. This cannot be the case for small complex wedges developed in gravels and sands (1020% of ice) or for the small true ice wedges in the frozen sitly loams (max 30% of ice), as at Marcilly section. Palaeo-ice wedging occurs in western Europe usually in depressions in cold and dry environment (Van Vliet-Lanoë, 1989, 2005; Kasse and Vandenberghe, 1998).

Moisture is stored today at depth in the permafrost, commonly within the upper 15 m, but mostly close to the permafrost table, in equilibrium with the winter snow depth and the freezing day index. In the valley, as the water is supplied by superficial melt, it is provided by river run off and by slope external and internal drainage. During abrupt warming, as in Late Glacial or early

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Holocene, the thaw of the ice wedge is very fast, induces collapse features (Solovyev, 1973; Hyatt, 1990; Michel and Van Everdingen, 1994) and piping at the level of the permafrost table (Seppälä, 1997). Lateral slides of the wedge rims develop in association with narrow zigzagging gullies, known as "badjaraks" from Yakoutia (Solovyev, 1973). It is probably the case for the Marcilly section. This collapse process normally induces upwards injections of plastic thawed material along tension gashes controlled by the subsidence, further evolving in mud boils and stretched towards the furrow by frost creep (Fig. 9), leading to sometimes very complicated deformations (Van Vliet-Lanoe, 2005). Present ice wedge and ice vein degradation has been analyzed in Arctic and Subarctic (Murton and French, 1993; Van Vliet-Lanoe, 2005).

Upper Pleniglacials in western Europe are commonly arid and the ice content in the ground limited. Regional precipitation are around 200 mm/y, although early glacial are closer to 400 mm/y (based on palynology and soils; Lebret et al., 1996). It means that the snow cover was too thin to prevent thermal cracking and the possible formation of sand wedges, complex wedges or true ice wedges.

If the soil cover is vegetated, from our experience in Arctic (Van Vliet-Lanoe, 2005), the ice of the wedge is usually rather clean. Seeing the limited content in organics in the sediments of Sauvage and Marcilly quarries (barren grounds), most of the wedges are probably complex or sandy, not pure ice. The morphology of the epigenetic palaeo-wedges is probably shallow (2 m), like the Holocene ones in the Advent valley (central Svalbard) or in the sector of Salluit (Nunavut), but rather wide close to the surface to at maximum a meter wide (Van Vliet-Lanoe, 2005). The widening, even not constant in time, is about 2 mm/year (Burn, 1990; Mackay, 1992).

From the observation of the Little Ice Age in Svalbard, we may estimate its maximum wide to 60 cm at the permafrost table. Compared to the section of triangular and shallow palaeo-sands wedges of the same age in Flanders (Buylaert et al., 2009), the "ice" wedge at Marcilly section was 1.25 m in depth, with an active layer of about 1 m, for about

3 ka of cracking activity. Such small ice wedges are insufficient to promote alas formation. Recumbent folds forms (Allard and Kasper, 2001; Van Vliet-Lanoe, 2005) occur today with the aeolian aggradation and the ice wedge development, but may be accentuated by a limited thermokarst related to change in precipitation, before being stabilized by a late permafrost aggradation. A final gravitational collapse occurred to the East at Marcilly section, related to the major Bölling phase of thermokarst (See Section 5.4.1) perhaps related to a river channel incision and migration.

5.4. DATING THE DEPOSITS AND INDIRECTLY THE THERMOKARST EVENTS

Several upper Weichselian events of thermokarst have been recognized in the loesses of western Europe, mostly in association with ice wedge casting. The site of Harmignies, located in Southern Belgium, at 20 km from the French boarder and at 165 km North of Marcilly quarry and the nearby Maisières-Canal provided the longest and clearest regional record on chalk (Haesaerts and Van Vliet-Lanoe, 1981; Van Vliet-Lanoe, 1992; Van Vliet-Lanoe, 1989; Haesaerts et al., 2016). It is thus a very good analogue for the situation in low valley at the confluence of Aube and Seine Rivers. At Maisieres-Canal, three main events were recorded at 28,2 ka (14 C), 14.50 ka (onset of the Bolling; valley record) and ca 12 ka (14 C; end of the Younger Dryas). At

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Harmignies (plateaus record; Figs. 8 and 10) older events were first attributed to 23.6 ka and 16-17 ka. The 23.6 ka event (ex-Nagelbeek horizon) occurred here just after a tundra gley dated by IRSL at 24 ± 4 ka (Frechen et al., 2001).

The major complex thermokarst exists immediately after the Sol des Vaux (Harmignies HB4), developed in sediments dated at 37,5 ka by OSL at Harvrincourt (Antoine et al., 2014) (Fig. 1) and Nussloch (Germany; Antoine et al., 2009) (Fig. 1).

Other type of thermokarst are much more difficult to recognize and to date, but were observed in the Brackveen (Bastin et al., 1974; Pissart, 2003) and in Wales (Watson, 1971; Watson, 1996).

5.4.1. LAST DEFORMATIONS

Seeing the limited loess deposition during the Weichselian in the sector of the Seine and Aube Rivers (Fig. 1), the preservation of some microrelief (DEM) and on the impact on vegetation of the former thaw lakes suggest a rather young age for most of them, after 16.6 ka (Marcilly section). It represents a permafrost aggradation followed by a thermokarst prior to the last main vegetation event and landscape stabilization, the onset of the Bolling Interstadial (Van Vliet-Lanoe et al., 1992; Pastre et al., 2000). This should fitGS 1 at NGRIP (15.12-12.9 b2k age; Rasmussen et al., 2014). All this should emplace at Sauvage (Fig. 10) and Marcilly the upper major thermokarst at the onset of the Bolling, during NGRIP interstadial GI 1e (14.7 ± 0.1 ka; Rasmussen et al., 2014). The superficial deformations described close to Marcilly section by Bertran et al. (2018) are probably all formed at the same age as the last large event at Sauvage and the latest reactivation in the lower Fyb terrace of Clesle/Aube River (Van Vliet-Lanoe et al., 2018a) as at Laismont and Domrémy-la-Pucelle quarries. The fresh forms on alluvial grezes on chalk, North of Laon city, are probably also of the same generation (Fig. 1): some of them are even ramparded (40 m in diameter, 3 m high) like in Wales (Gurney, 1995), and very similar to the ramparded depressions on Askja volcano (Fig. 7), issued from opensystem pingos as stressed by the Watson's (1971, 1996), Ballantyne and Harris (1993) and Gurney (2001).

5.4.2. THE 16.6 KA EVENT

The small stabilization event at 16.6 ka at Marcilly section yields a very weak warming prior to NGRIP stadial GS 1 at the onset of GI 1 (Andersen et al., 2006). It is thus not a climatic thermokarst, but deformations are more probably controlled by ground water conditions like in the Sauvage quarry and possibly the rise in precipitations. The OSL dating of the cover sands at Marcilly (16.6 ± 0.9 ka, Shfd 17,101) fits particularly the late Weichselian with moderate and slow thermokarst associated with recurrent aeolian aggradation (Van Vliet-Lanoe, 1992; Kasse et al., 2007).

Fig. 5. *Various forms of collapse related with their thermokarst signature. Lake ice structure is from Lake Boniface terrace, Nunavut (Canada). The collapse forms are rather similar, adapted to rheological properties of the sediment.*

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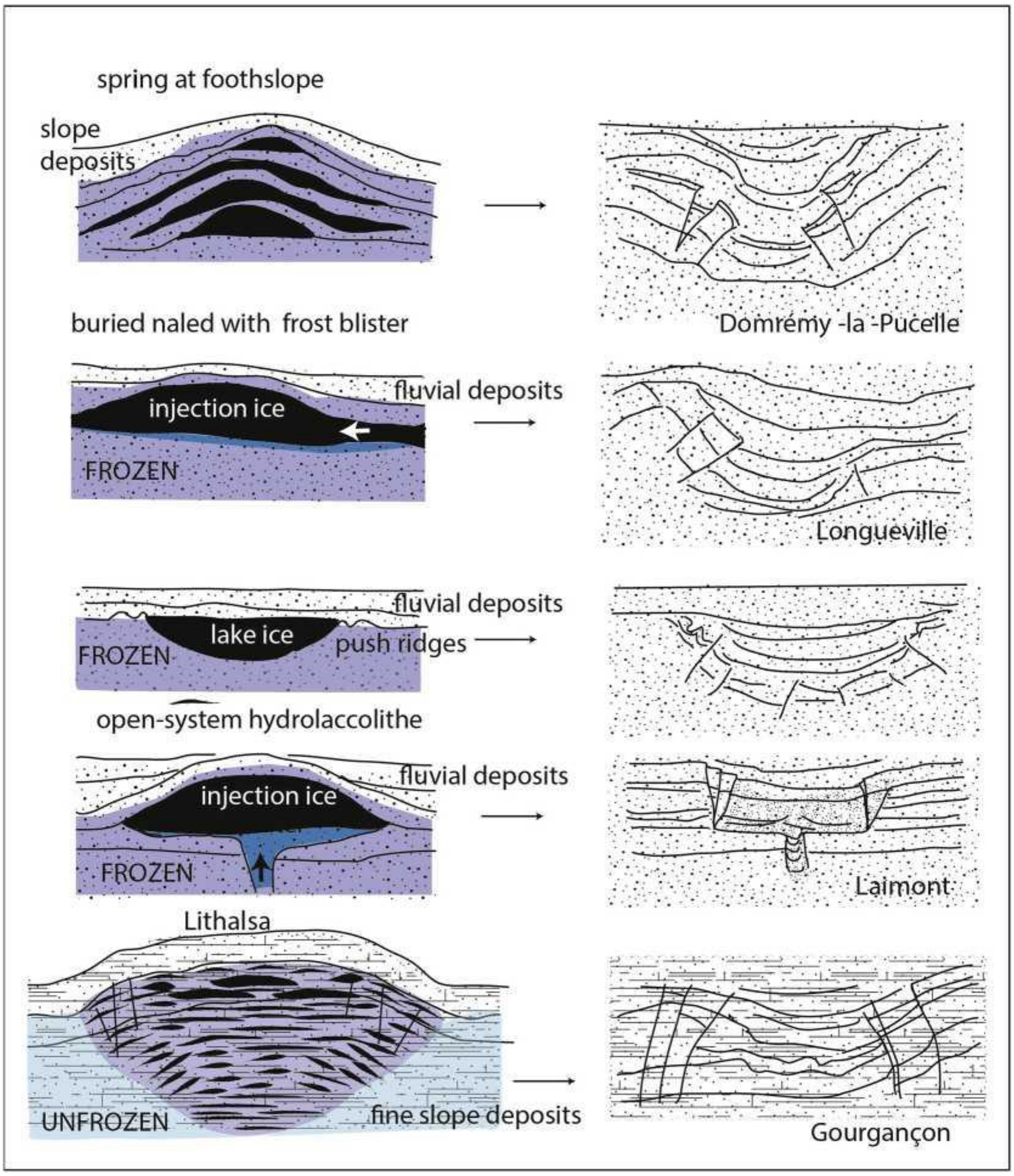

5.4.3. THE 23 KA EVENT

The radiocarbon dating at Marcilly section (24.6-24.1 cal ka) corresponds at Harmignies quarry to thermokarst gullies on the largest net of ice wedge polygons (unit HC6, IRSL dating at 24 ± 0.4 ka; Fig. 4). The post-HC6 degradation at Harmignies event yields the NGRIP inter- stadials GI 2.1 and GI 2.2, 23.0 and 23.3 ka 2bk (Rasmussen et al., 2014) as also stressed by Bertran et al. for the Marcilly lower event. The "lacustrine" stratified silts correspond thus well to a thermokarst, but seeing the size of the ice wedge cast, it is too small to induce large thermokarst lake. It is more probably issued from the melting of ice bodies within the gravels and sands of Fx b unit.

5.4.4. THE OLDER EVENTS

The Maisières-Canal event at 28,2 ka (organic soil) yields the ±GI-3 and GI-4, respectively at 27,540 and 28,900 a 2bk (Rasmussen et al., 2014).

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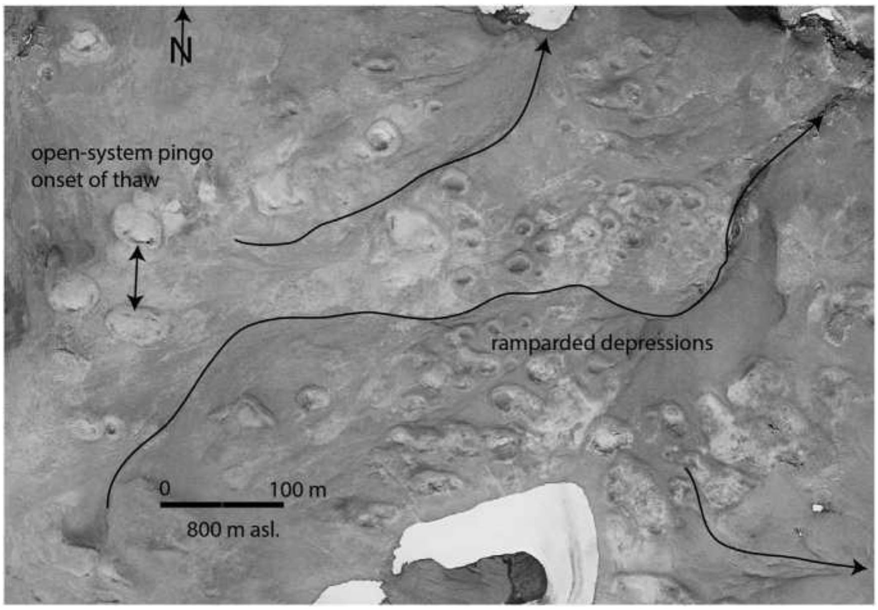
The HB4 soil and the post-HB4 thermokarst complex at Harmignies yield NGRIP interstadials GI 5, 6 and 7.1 and 7.2, respectively at 30,840, 33,740 and 34,880 & 35,020 a 2bk (Rasmussen et al., 2014), in conformity with the OSL age of the host sediment (37.5 ka, followed by soil formation during GI 8 and by permafrost installation). Fig. 4 reveals at least two consecutive events, the first being the major one.

No major thermokarst have been observed earlier, GI 18 and 19 warmings being of lesser temperature rise. The only event susceptible to induce such an abrupt warming is the onset of GI 21.1 at 84,760 a 2bk (Rasmussen et al., 2014).

The other features observed in quarries, are usually observed below a major precipitation of carbonates issued from a slightly rubified MIS 5 pedocomplex, observed in Longueville, Gourgançon and Domremy-la- Pucelle quarries. They are pre-Eemian, mostly connected to a major interstadial, ca 157-147 kyr BP (Van Vliet-Lanoe et al., 2018a), orbital forced and commonly associated with a major glacial retreat intra MIS 6 in northern Europe (Toucanne et al., 2009; Margari et al., 2014), between the Drenthe and Warthe glacial Advances.

Fig. 6. *Present-day open-system hydrolaccolithes at onset of degradation (arrows for first ring fault), evolving to ramparded depressions (rhyolitic pumices), on gentle slope, West of Askja volcano, Iceland. Some overlapping of the rampart suggests a polyphased system.*



5.5. POTENTIAL EVOLUTION OF LITHALSA AND HYDROLACCOLITHES DURING A PERIGLACIAL CYCLE

Trenches and geophysical surveys crossing circular frost mound or circular ramparded depressions revealed complicated organization. Ambiguous structures exist suggesting a mixed functioning as lithalsa but with a supplied subjacent contribution of water (Akerman and Malmström, 1986) or open-system pingo surrounded by lithalsa plateau, as the complex open-system pingo of Innerhytta in Advent valley in Spitzberg (Ross et al., 2007).

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The traces of lithalses of Brackvenn (Belgium) and some circular ramparded depressions of Britain (Bastin et al., 1974; Gurney, 1995 and 2001) were developed especially in silto-argillaceous materials. These materials yield a high retention of water and highly susceptible to segregation of ice, especially reticulate in not-consolidated materials (marine, alluvial or lake sediments; Svensson, 1964; Mac Roberts and Nixon, 1975; Ahman, 1976; Iwahana et al., 2012) also observe at Gourgançon. These structures form after melt metric faulted depressions (3-4 m of depth) for initial structures of limited diameter (50-70 m; Fig. 4C; Van Vliet-Lanoe et al., 2018a).

On the other hand, morphologies of ramparts developed in sandy or gravelly materials correspond frequently to seasonal hydrolaccolites (410 m in diameter; Van Everdingen, 1982) or perennial hydrolaccolites (open system pingo; 15-60 m in diameter), with massive lenses of injected ice supplied with a lateral drainage. This supports an abrupt thermokarst as with Longueville section (Van Vliet-Lanoe et al., 2018a), with internal walls of stiff ramparts and a more variable diameter size.

The lithalses, all like the "open system" pingos, are formed in a grouped way and attest of an episode of significant climatic cooling (Washburn, 1983; Van Vliet-Lanoe, 2014). The forms of the Brackvenn (Fig. 1) have a relatively simple morphology, sometimes faulted, and were formed during the Younger Dryas (Pissart and Juvigne, 1980; Pissart, 2003), like part of those of Wales (Watson, 1971; Watson, 1996; Gurney, 2001). The majority of active structures in Subarctic were formed in Neoglacial (Subboreal) or more recently, during the coldest phases of Little Ice Age, as in Norway (Ahman, 1976) or in Iceland (Friedman et al., 1971; Sæmundsson et al., 2012). Those of Champagne evolved especially at the end of the Saalian and Weichelian, after a cold maxima (Van Vliet-Lanoe et al., 2018a).

Lithalsas and "open" pingos could exist in the same area, as in Central Iceland, with a slight cooler context for open-system pingo (continuous versus discontinuous permafrost). When the cold intensifies, the lithalses can coalesce by lateral development (Pissart et al., 2011), forming a plateau as in Altai' (Blyakharchuk et al., 2008), building by lateral extent a confining barrier for the water-table in valley (Fig. 9). This barrier probably allowed the setting under pressure ofa trapped aquifer resulting into the evolution into open-system pingo of former lithalsa plateau. The seasonal hydrolaccolites of foot of slope may also evolve with time in opensystem pingos.

It is thus not astonishing that with climatic variability during a glacial episode, it is possible to pass by relatively simple forms, as for a short episode such as Younger Dryas, to more complex hybride forms. Certain isolated forms of big size (300-400 m of diameter near Gourgançon quarry) could correspond to coldest of the last glaciations to true closed-system pingos with confined water table (modelled permafrost: 110 m of depth for Champagne with the Weichsel LGM; Lebret et al., 1996; Holmes et al., 2011). Thermal cracking may occur in surface of the frost mound if the ice content is high (cf.Inner Hytta pingo, Advent valley, Fig. 7). The recurrence of episodes of growth and degradation is common during the Holocene Arctic (Lenz et al., 2016; Matthews et al., 1997) and this situation had to prevail during alternations stadials/interstadials during both the Weichselian and Saalian glacials.

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5.6 PERMAFROST THERMOKARST ACTIVITIES IN THE SAALIAN AND THE WEICHSELIAN: A RECURRENT STORY

As the sequence covers as well the Upper Weichselian at Marcilly and Sauvage, and the MIS6c-MIS6c at Gourgançon, Longueville and Domremy-la-Pucelle, the content is ice of the permafrost ought to change with time. Permafrost melt is controlled temperature rise, snow cover and change in hydraulic conditions (e.g. Romanovsky et al., 2003). In Arctic, thermokarst proceeds commonly with higher summer temperature, but more frequently with mild winter temperature and more extended snow cover (Osterkamp, 2005). Winter warm temperatures were recorded in the northern Atlantic during MIS 6b and MIS 3 (Wainer et al., 2011; Wary et al., 2016).

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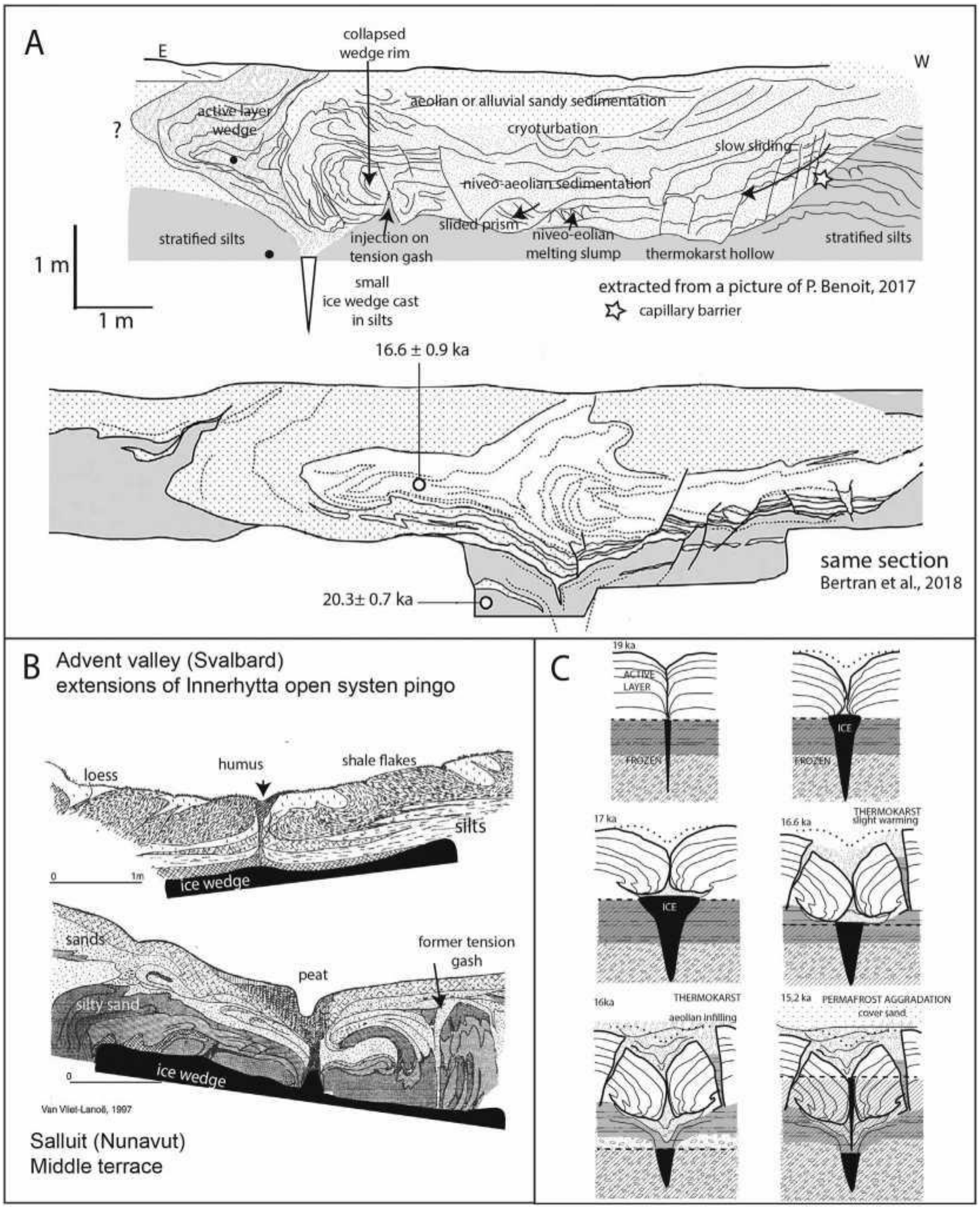
Fig. 7. *Evolution of the ice wedge at Marcilly section*



A) Section interpretation by us at the light of early degradation of ice wedge furrows and by Bertran et al. (2018). B) Section in early degradation of "small" ice wedges in Arctic (from Van Vliet-Lanoe, 2005). C) Reconstitution of the evolution of Marcilly very small ice wedge.

A) WEICHSELIAN

The best and dated regional record is located on ice wedges at Harmignies and Maisières-Canal in Southern Belgium. The oldest event occurred at Harmignies at the time of the Rocourt tephra (Haesaerts and Van Vliet-Lanoe, 1981; Haesaerts et al., 2016), just after a brief but brutal cooling associated with the first isolated ice wedge at the onset of MIS 5b, dated ca 105 ka (Frechen et al., 2001), by insolation (precession minimum at mid latitude: Van Vliet- Lanoe, 2014) and

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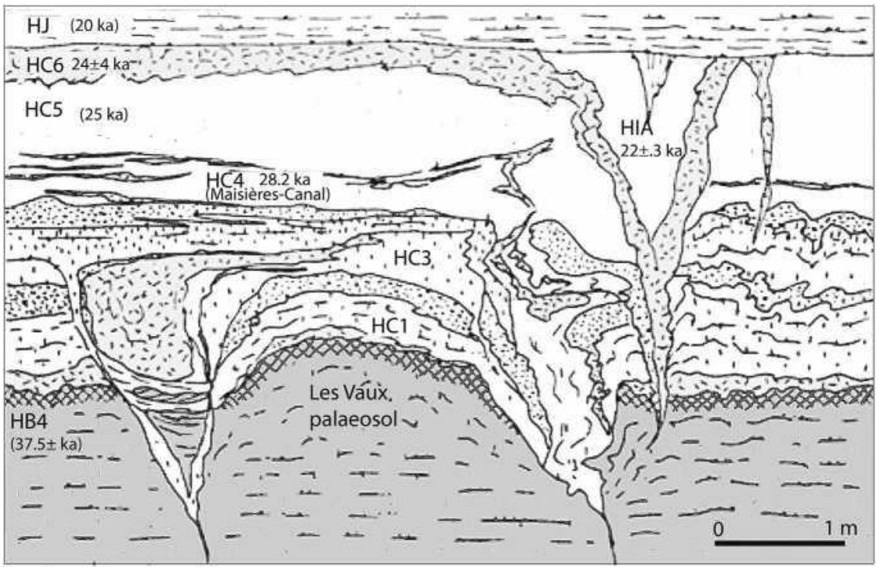
precipitation change related to Dansgaerd-Oeschger events (DO) with the restoral of the thermohaline circulation (Rasmussen et al., 2016; Wary et al., 2016) and thus higher precipitation. Most of the DO (ca 500 yrs) or GI events at NGRIP show a one-step, rapid temperature increase, although DO events 2, 7, 11 and 18 are characterized by a two-step temperature rise with a small interruption of generally b100 years between stadial and interstadial conditions (Kindler et al., 2014).

Permafrost aggraded in Belgium during most MIS 4 and 3 with several periods of thermokarst; between 32 and 28 ka, deforming the "Sol des Vaux". This palaeosoil (ex-Kesselt palaeosoil; Haesaerts and Van Vliet-Lanoe, 1981; Van Vliet-Lanoe, 1991) was associated with Pinus charcoal, preserved in thermokarst in northern France, 40 km south of Harmignies quarry, Fig. 1), meaning that some slow thermokarst occurred progressively during this boreal pedogenesis on a 37;5 ka sediment. The pedogenesis developed synchronously with GI 7, ca 35-34 ka (2bk), the longest and warmest of these interstadials. Permafrost remained as relict at depth (ca 3-5 m).

Fig. 8. *Evidence of middle Weichselian multiphased thermokarst at Harmignie (adapted from Haesaerts and Van Vliet-Lanoë, 1981. IRSL dating: Frechen et al., 2001).*



A second but major thermokarst complex occurred in Harmignies quarry just after HB4 (Fig. 4), between 30 and 33 ka (see higher), in relation with Greenland interstadial GI 5-6 (Rasmussen et al., 2014). It is associated with runoff and gully incision. Each GI event interrupted the permafrost re-aggradation from the Les Vaux interstadial. It corresponds probably to the oldest generation of wedge casts at Sauvage quarry (Fig. 4).

This thermokarst complex is further truncated by a thermokarst event recorded in HC4 at Harmignies and at Maizières-canal in association with humic soils, mammoths and a Gravetian industry dated at 28.2 ka followed by a thermokarst on ice wedges (Haesaerts and Van Vliet-Lanoe, 1981; Haesaerts et al., 2016) fitting the 28,9-27.5 ka (2bk) of GI 4 and GI 3 (Rasmussen et al., 2014).

The next large event occurred after HC6 large ice wedge net and the coldest stadial of the Weichselian (January temperature: -15 °C; 300 mm/y precipitation in Velay; Wainer et al.,

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2011), with a deepening of ca 1 m of the permafrost table, shrub development in the thermokarst and solifluction development (HC6, Van Vliet-Lanoe, 1991) from ca 23 and 23.3 ka. It occurred just after the Marcilly radiocarbon dating (24.624.1 cal ka) in "thermokarst lacustrine silts" (Bertran et al., 2018). At that time, precipitations rise again regionally (Andrieu-Ponel et al., 2007; Wainer et al., 2011). By comparison, the present-day warming in Arctic represents b50 cm of permafrost table deepening since 1990 (Romanovsky et al., 2003; Etzelmuller et al., 2011; Way and Lewkowicz, 2016). Another large event occurred at 20 ± 3 ka in association with large erosion gullies at Harmignies (IRSL; HJ unit), within a cold event GS 2.1b, suggesting another type of rapid forcing than DO events, as shown bya peak in 10Be (Yiou et al., 1997) and potential high solar activity. It postdates an ice wedge net as well at Harmignies as in Flanders around 21 ± 1 ka (Buylaert et al., 2009).

The period between 20 ka and 14.5 ka is marked aeolian aggradation as well at 50°N (loess) as 51° N (cover sands; Kasse et al., 2007), in relation with rising precipitation, and activity in the rivers net to a permafrost aggradation related with GS1. The "recumbent fold" of Marcilly section corresponds to the active layer evidence ca 1 m of depth (limited moisture/ice content) at 16.6 ka in niveo-aeolian sands dated, on a small ice wedge developed after 24 ka. It is stabilized with cover sands. The cover sands equate GS 1, ca 15.2 ka, known as the Pomeranian stadial or Older Dryas (Stroeven et al., 2016) and the 'Beuningen gravel beds' in Netherlands and northern Belgium (Kasse et al., 2007; Buylaert et al., 2009).

The major thermokarst event took place later in eastern France and Belgium at low altitude, just prior the development of the Bolling birch forest (Van Vliet-Lanoe et al., 1992) and the evolution of rivers into a meandering system (Pastre et al., 2000). At the end of the Younger Dryas, the thermokarst is limited by present at low altitude in northern France and Belgium, more marked in altitude (700 m asl.) in Ardennes and Wales.

B) THE SAALIAN

The Fyb terrace is formed from 180 kyr during the Early Saalian (Voinchet et al., 2015). The interaction between orbital forcing and millennial variations recorded by isostopes allows to subdivise the MIS6 into an early Glacial, MIS 6c, an interstadial MIS6b: the Zeifen interstadial and a final stadial MIS 6a or Kattegat Oscillation (Seidenkrantz et al., 1996; Toucanne et al., 2009; Margari et al., 2014). This second cold maximum yield 140 and 136 ka (Margari et al., 2014).

MIS6c is relatively moist and boreal with high-amplitude periodic fluctuations of the European climate (Margari et al., 2014), dominated by orbital forcing. Based on oxygen isotope, an important cooling is reached ca. 166 and 156 kyr with a drop of the sea-level (Elderfield et al., 2012). It is associated with the Drenthe glacial advance, with higher summer Sea-Surface Temperatures in Arctic (Eynaud et al., 2007), and with a global aridity favoring the first pick of dust in the MIS 6 at Vostok (Petit et al., 1999) and EPICA (2004). Precipitation is at 160 ka evaluated to 400 mm/y in Massif Central (Wainer et al., 2011).

During MIS6c, from 180 ka BP, the climate in eastern basin of Paris is oceanic subarctic, with a shallow, rich in ice, discontinuous to continuous permafrost, snowy as prove by the absence of ice wedging. It confined seasonally the diffuse karst water-table and allowed possibly thick perennial snow patches on the North or East-facing slopes, favoring frost shattering. This

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situation is responsible for the formation of seasonal hydrolaccolithes, frost blisters or icings in the Aube River valley, especially in sand and gravels. In fine grained sediments, lithalsa developed in the alluvial plain. The vegetation is probably prostrated on the plateau where huge pattern ground continued to be formed. The North-facing slopes are pasted by stratified solifluction deposits ('grezes litées'), including some ice bodies (Fig. 9). Between ca. 166 and 156 kyr, the climate cooled down (Margari et al., 2014) with a marked aridity fitting the first and highest pick of dust at Vostok (Petit et al., 1999) and EPICA (2004). It is probably the same for the drift sands in the karst of upper Aube. Some seasonal hydrolaccolithes evolve into open-system pingos. The lithalsa are fossilized by increasing cold, even they can expand laterally.

From 156 ka BP, climate improved somewhat. **MIS 6b** is a transitory period, orbitally forced, with a maximum of insolation centered on 151 ka BP (obliquity, minimum of precession, Martinson et al., 1987; Laskar et al., 2004) and is known as the Zeiffen interstadial (Seidenkrantz et al., 1996). It onset is probably after 155 kyr, a period of renewed volcanic activity linked to deglaciation in Iceland (Van Vliet-Lanoë et al., 2018a). Most ice sheets in the northern Hemisphere attest an important retreat (Ehlers et al., 2011; Eynaud et al., 2007), in connection with the summer restoral of the thermohaline circulation to the Arctic and higher precipitation: it is estimated at 600 mm/y in Massif Central (Wainer et al., 2011). It is also highly probable that increasing precipitation, orbitally forced and perhaps solar activity favored a retreat of the southern part of the NH ice sheets (Colleon, 2009) and a major meltwater discharge in the English Channel atca 155 ka (Toucanne et al., 2009). An important thermokarst event occurred as nowadays observed in the Lena River watershed (Iijima et al., 2010). This warming was most probably pulsed by DO events and resulted in a progressive melts of the superficial part of the permafrost, creating locally taliks and favoring icing or frost blister development. Related to this warming, a large unfrozen near-surface aquifer developed, with perennial groundwater flow. Recharge and discharge became more active and increased groundwater discharge certainly affected the base flow of many rivers as today (Michel & Van Everdingen, 1994). This warming with raised precipitation allowed the reworking of sediments from the slopes and loesses from the plateau, that splayed on the Fyb terrace of the Aube River that evolved steeper as also the Ornain River terrace. This situation favored during brief cooling episodes the renewed formation of lithalsa or seasonal frost blisters (discontinuous permafrost). Thermokarst on shallow epigenetic ground ice reached a steady-state depth, achieved within a century only, as today (West and Plug, 2008). By opposition, general deep thermokarst was progressive, at least spread on 5-6000 yrs., but relict permafrost persisted at depth as the result of its ice richness (ca 4 m for the Aube River watershed). Around 150 kyr, in connection with the thermal and orbital warming, thermokarst activity is mostly achieved, vegetation expanded, the slopes were stabilized and the Aube River began to incise the Fyb terrace, although relict permafrost persisted at depth. It allowed the capture of the Ornain River by the Marne River, close to Laimont (Cojan et al., 2007).

MIS6a is a final stadial marked by conditions nearly as cold as the MIS2, responsible for the Wartha advance in northern Europe: the Kattegat Stadial (Seidenkrantz et al., 1996). Winter sea-ice is prograding far south, explaining the aridity with a huge ice-shelf on the Arctic (Jakobsson et al., 2016). Between 145 and 132 kyr (onset of the MIS5e), the region evolved under a cold and dry periglacial climate in similar conditions to the coldest period of the Weichselian (MIS2); precipitation is evaluated to 500 mm/y in Massif Central (Wainer et al., 2011). Relict permafrost

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evolved to continuous by a renewed aggradation, but poorer in ice due to the aridity. In turn, this aridity was favourable to a major Aeolian activity and to thermal contraction. The Fyb terrace at la Varenne/Seine River corresponds to this situation. The regional frost wedges and ice wedge casts occurred in the coldest but driest context of the MIS 6, ca 140 ka, together with the second peak of dust at Vostok (Petit et al., 1999) and EPICA (2004).

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Fig. 9. Evolution of the permafrost on chalk in North-Eastern France during one periglacial cycle. Yellow: alluvial grèzes and grazes litées; orange: Holocene alluvium.Open-system pingos are fed both from the water table or lateraly from slope drainage.

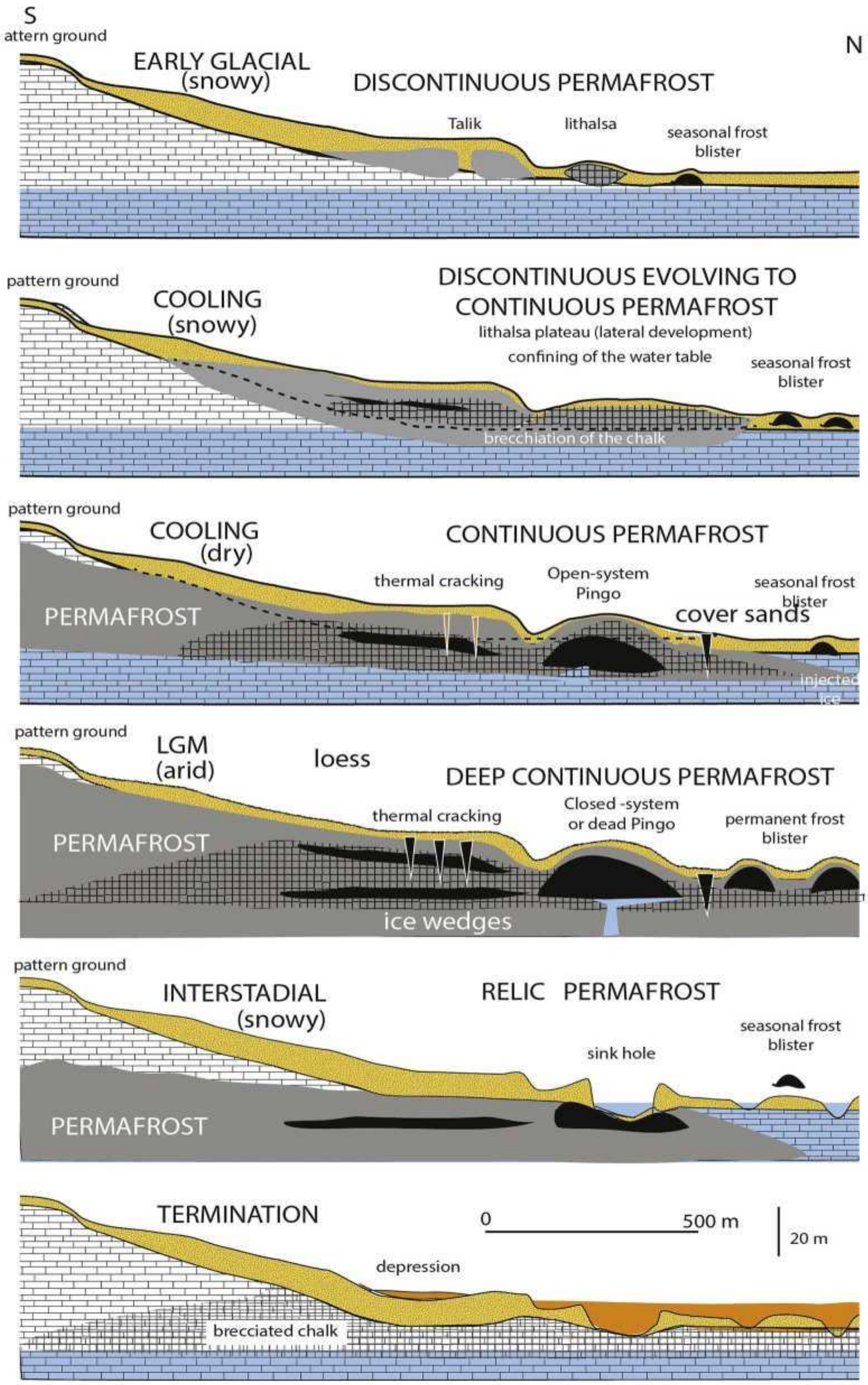

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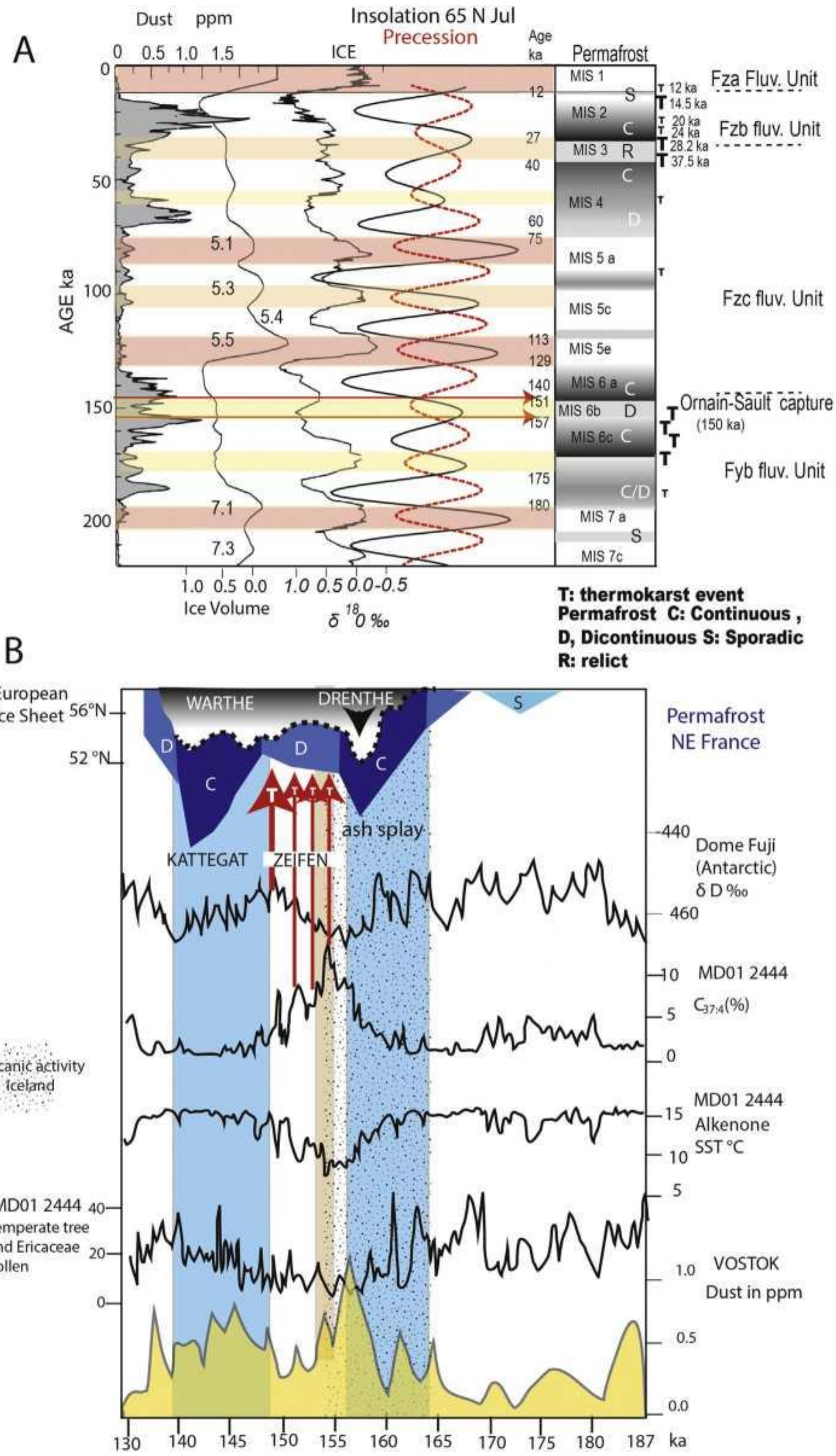
Fig. 10.



A) Positions in time of the different river terraces in relation with the climate forcing, ice volume, permafrost, thermokarst and aeolian activity (insolation and precession: Laskar et al., 2004); ice volume, dust and δ18 O: Vostok (Petit et al., 1999). B) Forcing of the thermokarst in NE France, MDO1 2444: Margari et al. 2014; Vostok: Petit et al. (1999), Dome Fuji: Watanabe et al. (2003); volcanisme: Van Vliet-Lanoë et al. (2018a).

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6. Conclusion

The thermokarst activity in northern France and Belgium is much more variated that in the paper ofP. Bertran et al. (2018). Thermokarst events mostly relate to various frost mounds created by injection ice and segregated ice. The role of ice wedges is really very limited in this zone of southern extent of the European palaeo-permafrost. Thermokarst are in first order orbitally forced as well during the Weichselian and the Saalian. They are associated with abrupt warming by Dansgaerd Oeschger events transmitted regionally by the thermohaline circulation, under control ofa maximum insolation and a minimum in precession. They are for these last orbital reasons followed on the continent by a reorganization of the rivers from braided to meandering systems. Snowiness and mild winter temperatures recorded in long sequences (Velay, South-Eastern France) are probably the main triggers for thermokarst activity as occurring today. Other events can be triggered by solar activity as at 20 ka or perhaps enhanced by large ash splay as during the MIS6b Zeiffen interstadial. Thermokarst events are mostly susceptible to occur during the coldest/latest part of the glacial due to the richness in accumulated ground ice.

With a milder warming or a retrogressive thermokarst, deformations are more progressive, in direct relation with the rheological properties of the sediments, the segregated or injected ice content or the initial size of the ice wedge and in fine the ground drainage. They are in concurrence with the vegetation that will limit its expression.

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