Earth-Science Reviews 232 (2022) 104126

Contents lists available at ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

The role of post UK-LGM erosion processes in the long-term storage of buried organic C across Great Britain – A 'first order' assessment

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ARTICLE INFO

Keywords:
Soil
Erosion
Burial
Organic Carbon
Holocene
Quaternary

ABSTRACT

Increasing consideration is being given as to whether and how the subsoil can be utilised as a resource to store greater quantities of organic carbon through a range of 'frontier' technologies. However, recent work suggests 'priming' effects may occur when fresh soil organic carbon (SOC) is mixed with older organic carbon (OC). Combined with increasing intensity of land use and perturbation of the surface environment there is potential for buried organic carbon (OC) to be re-incorporated into the active global C cycle. Therefore, understanding the nature of existing buried organic carbon (OC) within Soil Parent Material (SPM) and landscapes is increasingly important. A major OC burial route within landscapes is via erosion and deposition processes. This paper aims to provide a 'first order' overview of the role erosion processes have made since the UK Last Glacial Maximum (UK-LGM) in the burial of OC in Great Britain. Using collated information, Monte-Carlo simulations were used to produce 'first-order' estimates of the mass of OC buried within three deposit types; Devensian Till, Devensian Glacio-fluvial deposits and Holocene Alluvium. Combined median estimates for these three deposit types alone suggest, that 385 MT of OC has been buried in these deposits across Great Britain, demonstrating the importance of post UK-LGM erosion processes in long-term sequestration of OC. The paper provides a basis of a framework to describe where buried OC may be found within UK SPM and landscapes, whilst identifying gaps in our knowledge base. Whilst focusing on Great Britain, the processes are relevant to many countries, each of which will have experienced erosion processes unique to their own history of geology, geomorphology and climate.

1. Introduction

There is continued interest in using soil and the deeper soil parent material (SPM) resource to increase soil organic carbon (OC) sequestration (Minasny et al., 2017), based partly on effective mechanisms recognized in the long-term burial and sequestration of soil organic carbon (OC) by externally transported material (Chaopricha and Marin-Spiotta, 2014; Hoffmann et al., 2013; D'Elia et al., 2017). Whilst results from long-term experiments have identified possible limitations regarding increasing OC storage in top-soil (Poulton et al., 2018), potential remains for increasing OC storage in the sub- (~1 m) and deeper (> 1 m) parts of the SPM profile (Lorenz and Lal, 2005), where substantial quantities of minerals (e.g. iron oxides, clays) that are important as OC stabilisers exist (Whitmore et al., 2015). Suggested emplacement methods of OC into these subsoil horizons have included utilising natural processes such as DOC leaching (Whitmore et al., 2015), breeding plants with deeper rooting systems (Mikutta et al., 2006) or the injection of pelletized organic matter (Leskiw and Welsh, 2012). Paustian et al. (2019) describe these and other possibilities as 'frontier technologies' for which significant economic and technological barriers still exist, and for which further research and incentivization are required. However, complications exist regarding the possible re-incorporation of buried carbon into the global C cycle. The potential 'priming' of existing buried carbon leading to changes in SOM decomposition, through addition of labile C and nutrients, has been recognized in both terrestrial and aquatic environments (Bianchi and Ward, 2019; Kayler et al., 2019), and with changes in land-use or agricultural practice (Fontaine et al., 2007). In addition, erosion or excavation processes may translocate buried carbon towards the soil surface where oxidation is more likely.

Given the urgency to abate climate change, it is important that research considers how to maximise OC storage and sequestration potential, whilst ensuring that OC storage mechanisms are protected

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https://doi.org/10.1016/j.earscirev.2022.104126
Received 2 December 2021; Received in revised form 7 July 2022; Accepted 11 July 2022
Available online 14 July 2022

against, and resilient to, perturbation. The first stages in this effort are to establish where, and how much, buried OC exists, the processes that led to its burial, and the mechanisms that have hitherto protected it against mineralization. By doing so, effective national strategies for using SPM for OC sequestration may be developed. Such effort has not enjoyed much scholarly attention in the past. This gap is addressed here by reviewing the role of soil/sediment erosion and deposition since the UK Last Glacial Maximum (UK-LGM; MIS 2), which has led to OC burial. The UK-LGM represents the most recent event where a widespread natural perturbation of the landscape occurred.

2. Materials and methods

This review aims to use spatial knowledge of sediment deposited since the UK-LGM along with knowledge of OC concentrations to produce an overview of buried OC within the UK. The periods of interest are the Chronostratigraphic Stages (Global Quaternary timescale) of Marine Isotope Stages (MIS) 2 and 1. MIS 2, often referred to as the Late Devensian in the UK is the period between ~29 and 11.7 K yr BP, whilst MIS 1 is known as the Holocene and covers the period ~11.7 K yr BP to present. Within MIS 2 is the UK-LGM which has been estimated at ~27 K yr BP, and represents the period of maximum ice sheet expansion in Britain and Ireland (Clark et al., 2012). The biostratigraphic terms for Interstadials (warm) and Stadials (cold) during MIS 2 are also used. These include the Oldest (~18.5–15 K yr BP, Older (~14 K yr BP) and Younger Dryas (~12.9–11.7 K yr BP) stadials and the Bølling and Allerød interstadials. For the purposes of published geological maps, the British Geological Survey has historically used the term Devensian to cover the time interval that incorporates the Late Devensian. Information comes from geological and soil surveying, geomorphological interpretation associated with landscape development, and archaeological studies where dating (^{14}C or OSL) of deposited sediments has taken place. Collating this information from spatially local studies, a chronology can be established of the interactions between climate events and human impacts that produce both SPM deposits and buried OC. Typically, the magnitude of these erosion and depositional events are largely reported as thicknesses in borehole logs, and these can be unrepresentative of wider areas, as they often represent only a specific site of interest. The use of 3D geological models allow estimations of deposit thickness over wider areas (Brown, 2009; Hoffmann et al., 2007). Thus, a novel aspect of this review is the use of regional 3D geological models (1:50000) to estimate mean deposit thickness within landscapes by combining estimates of area and volume (see Supplementary Information A). Data from regional 3D models are complemented with outputs from two high-resolution 3D models of the near surface environment. These models were produced for Talla, Scotland; a landscape within the areal extent of the UK-LGM and typical of Southern Uplands terrain (Scheib et al., 2008), and at Shelford, England; a landscape positioned south of this extent (Tye et al., 2011). The locations of these 3D models are shown in Supplementary Information B. In addition, published data from the Soil Survey of England and Wales are used, along with field data collected during the production of the high-resolution 3D models, to identify concentrations of OC in SPM, namely the 'C' soil horizon.

To demonstrate the importance of erosion processes and the potential extent of OC burial that may result, data collated from the review (e.g. deposit thickness, deposit area, Total OC concentrations) were used to produce 'first order' estimates of the mass of organic carbon stored within three of the deposit types examined as sufficient data was available. Monte-Carlo simulations were undertaken in the statistical package 'R' to produce cumulative distribution functions and results presented as box plots. The 'runif' command was used to select values from within a defined range, since the 'rnorm' function requires a meaningful standard deviation which the limited data collated could not provide. Selected distributions of data were tested for normality after the simulation was run 100,000 times. Results are presented in MT organic carbon.

3. Conceptual approach

Sediment deposition is identified in soil development models, particularly those involving slopes, such as *catena* classification (Sommer and Schlichting, 1997), Butler's K cycle concept (Butler, 1982) or soil-landscape chronograms (Vreeken, 1984). The resulting spatial distribution of SPM, along with properties such as thickness and texture, are fundamental in defining how pre-existing OC, along with that present in the transported material, is buried and undergoes long term sequestration. Within the UK, the redistribution of sediment to form SPM and buried OC is linked to a wide range of climate related geomorphological processes operating over different temporal and spatial scales, in addition to later anthropogenic interventions. However, the redistribution events following the UK-LGM must also be placed in the context of a much longer time series of re-distributional events, particularly previous glaciations (Clayton, 2000; Bridgland et al., 2004). Geomorphological controls which result in depositional SPM generation and OC burial include (i) Quaternary controls associated with the retreat of the British and Irish Ice Sheet (BIIS) and the associated Glacial Isostatic Adjustment (GIA) and Relative Sea Level (RSL) change which determines slope adjustment (Shennan and Horton, 2002); (ii) the distribution of geology within the UK including its strength, and; (iii) sediment availability.

The main conceptual approach of this synthesis is based on the Climate-Vegetation-Erosion (C-V-E) hypothesis, which is used as a framework to estimate the age of buried OC within the landscape. Vandenberghe (2002) suggested a coupled role between vegetation and erosion, driven by climate, whilst Dosseto et al. (2010) expanded this hypothesis, suggesting that the climatic influence on erosion over glacial cycles is indirect, and instead operates largely through amplifying erosion when vegetation cover diminishes in periods of climate deterioration. Thus, vegetation acts to bind the soil/sediment surface, reduce the impact of rainfall intensity, provides an evapotranspiration pathway that reduces soil moisture and pore water pressure, and increases the infiltration of precipitation. In addition, Net Primary Productivity (NPP) is likely to decrease during long climate cooling phases, thus impacting the role OC has on soil structure. Therefore, the C-V-E hypothesis suggests that erosion will correspond to both regular interglacial and glacial cycles (Milankovitch-scale), and shorter-term, often more abrupt (sub-Milankovitch), cycles that impact on a regional basis.

In this review, the C-V-E hypothesis is extended to assess the impact of Holocene climate events, and inevitably, how interactions with anthropogenic activity increase the frequency and magnitude of these events, leaving identifiable signatures of erosion and OC burial. In recent years, climate cycles have been elucidated using the Greenland Ice Core Project (GRIP) (Johnsen et al., 2001; Rasmussen et al., 2006). These reveal a complex pattern of cooling and warming phases within broader Milankovitch glacial cycles. Thus, sub-Milankovitch cycles such as cooling (Heinrich) events (stadials) (Oldest Dryas (GS-2a), Older Dryas and Younger Dryas (GS-1),) and warming interstadials including the Bølling (GIS-1e) and Allerød (GIS1abcd) events have been identified (see Table 1). In the Holocene, smaller cooling (Bond) events have been identified, occurring at ~11.1, 9.4, 8.2, 5.9, 4.2, 2.8, 1.4, 0.5 K yr BP. These are millennial scale cycles in the north Atlantic typified by cooling temperatures (Bond et al., 1999). There is considerable variability in the strength of these climate fluctuations and the subsequent responses. For example, most Bond events do not have a clear climate signal (only the 8.2 k yr event has a clear cooling signal in the GRIP analysis). However, the cooling of sea surface temperatures in connection with these ice rafting events are thought to impact on global climate patterns due to disruptions in the North Atlantic thermohaline circulation (Bond et al., 1999). These events vary globally, some corresponding to cooling periods, others coincident with aridification particularly at lower latitudes (Smith et al., 2016). A recent study by Dreibodt et al. (2020) has

Table 1
Sub-Milankovitch Climatic periods of relevance to this study.

Event	Date	GRIP Code*	Evident in UK	Climate	Refs
Oldest Dryas	approx. 18 –15 K yr BP	GS-2a	Yes	Polar climate Mean July temp: <10°C Mean Jan temps –20°C	Coope (1977)
Bølling Interstadial	13 –12 K yr BP. Not usually separated from Allerod in UK	GIS-1e	?	Mean July temp: 18°C Mean Dec temp: 2°C	Coope (1977)
Older Dryas	13.9 – 13.6 K yr BP		?	A brief cool interval within the Bølling-Allerød.	
Allerød interstadial	14.7–12.7 K yr BP	GIS-1abcd	Yes	If older Dryas not found than Bølling-Allerød is considered all Allerød. Characterized by peat formation across Uk and NW Europe. Mean July temp: 18°C Mean Dec temp: 2°C	Coope (1977)
Younger Dryas	12.9 –11.7 K yr BP	GS-1	Yes	Polar Climate Mean July temp: <10°C Mean Jan temps –20°C Latter period more arid than beginning	Walker et al., (1994) Coope (1977)
Bond event 8	11.1 K yr BP		?		
Bond event 7	10.3 K yr BP Bond event		?	Cooling detected from chromonid analysis of Hawes water	Lang et al., 2010
Bond event 6	9.4 K yr BP Bond event		?	Cooling detected from chromonid analysis of Hawes water	Lang et al., 2010
Bond event 5	8.2 K yr BP Bond event		Only Bond event that has a noticeable drop in temperature according to GRIP	Drop in temperature of between 1 and 5 degrees. Total duration 150 yr. Coldest period lasted about 60 yr	
Bond event 4	5.2 K yr BP Bond event		Transition in global climate between 6 and 5 k yr BP. Identified in peat bog records	In Ireland increase in wetness between 5525 BP and 5125 yr BP suggesting wet period before returning to dry values at 4950 a^{-1} BP. Cooler conditions also likely	Roland et al., 2015
Bond event 3	4.2 K yr BP Bond event		Evidence in northern Europe is ambiguous, suggesting the origin and effect of this event is spatially complex	Potentially wetter in Northern Britain in period 4.25–3.75 K yr BP but no regionally coherent phase of wetter / colder climatic conditions identified	Roland et al., 2017
Wet period	3.6 K yr BP (cal)		Evidence from bog records	Shift to wetter conditions	Charman et al., 2006; Charman 2010
Bond event 2	2.8 K yr BP Bond event		Detected in analysis of bogs	Climate instability event between 3.25 and 2.6 k yr BP. Initially a dryer period and then a shift to wetter/cooler conditions	Roland et al., 2017
Wet period	Centred around 2.76 K yr BP (cal) but likely 2.88–2.50 K yr BP (cal)		Evidence from bog records Considered distinct from Bond event 2	Maybe part of Bond 2 climatic instability event. Initially a dryer period and then a shift to wetter/cooler conditions at 2.7 K yr BP	Charman et al., 2006; Charman 2010 Plunkett & Swindles 2008
Wet period	1.6 K yr BP (cal)		Evidence from bog records	Shift to wetter conditions	Charman et al., 2006; Charman 2010
Bond event 1	1.4 k yr BP Bond event			Minor shift to wetter conditions	Charman et al., 2006; Charman 2010
Bond event 0	0.5 K yr BP event Approx. 1500-1800 AD		Often referred to as the Little ice age	Period of cooling after Medieval warm period. Wetter conditions identified in peat bogs in northern England	Mauquoy et al., 2002 Chambers et al., 1997

Links to references: Chambers et al., 1997; Coope, 1977; Lang et al., 2010; Mauquoy et al., 2002; Roland et al., 2015; Walker et al., 1994.

demonstrated how increased soil erosion may be coincident with some of these climate deteriorations in Eastern Europe. There are also centennial and decadal processes, driven by solar or volcanic forcing, which interact with oceanic and atmospheric processes to induce changes in climate. Evidence of solar forcing events in the mid- to late Holocene is spatially variable. In the UK, these effects appear less pronounced when compared to those at higher latitudes, where insolation forced change is weaker (Charman, 2010).

Further centennial climate events identified that may influence erosion and buried OC include a series of what were initially termed 'fluvial events' (e.g. Tipping, 1995a, 1995b). Information regarding these changes in the UK and Irish climate during the Holocene were obtained from reconstructions of surface wetness in Ombrotrophic mires (Plunkett and Swindles, 2008). These studies provide evidence of a series of changes in the precipitation-evaporation balances in northern UK (Charman et al., 2006; Charman, 2010). The number of these events identified, and which in this paper are referred to as 'flooding events', has increased. Benito et al. (2015) carried out a global meta-analysis of >2000 radio-metrically dated flood units, and results indicated that the UK may have experienced up to 16 episodes of increased flooding, since 10,700 cal. yr BP and these are outlined in Table 5. After 5000 cal. yr BP a general increase in flood frequency was found, likely caused by a weakening in zonal circulation and an increase in winter insolation. In the UK (and central Europe) 'flooding events' were multi-centennial and some corresponded with periods of minimum solar irradiance. These meta-analysis investigations are based on cumulative probability functions using ^{14}C dates and so, in the context of this study, provide an estimate of periods when there was potential for increased soil erosion and OC burial.

With respect to anthropogenic influence on erosion, the UK has a reasonably well documented narrative, particularly linked to land-use change (e.g. deforestation cycles). These started at the Mesolithic (hunter-gatherer)-Neolithic (cereal-livestock agriculture) transition (Whittle et al., 2011) and were identified through the decline of Lime (Tilia) pollen (Grant et al., 2011). Deforestation spread from southern England northwards (Woodbridge et al., 2014; Coombes et al., 2009; Mackay and Tallis, 1994) often in cycles of deforestation and abandonment as populations grew and declined (Dumayne, 1993) or land became exhausted (Tye et al., 2013). This is consistent with observations made across north western Europe (Fuchs et al., 2004; Starkel, 2005a; Starkel, 2005b). Importantly, Bocquet-Appel et al. (2012) suggested that the major driver on the rate of expansion of agriculture in Europe was soil fertility. Less fertile land, which is often lighter textured and more erodible, was exploited and vacated rapidly, whereas fertile land was able to support a population for longer before exhaustion.

Defining the exact nature and concentrations of OC in deposits formed from erosion and deposition processes is difficult because of the potential leakage and mixing of OC of different ages during later development of soil profiles in the deposited material. However, on a 'first order' basis the following definitions are used. Two classes of pre-Holocene buried OC are of interest in the described deposits. These are defined, for the purposes of this review as: (i) 'ancient' OC released through the weathering of sedimentary rocks during soil formation, physical weathering by ice during glacial times and from old soil horizons mixed with material during transport processes; and (ii) buried layers of OC rich materials such as or peat or palesols where the OC concentrations will likely reflect the soil characteristics at the time of burial. The term 'ancient' used in this context of buried OC refers to the concentration found in the soil 'C' horizon, which should largely be free from inputs of more recent OC. In soil pedological theory the 'C' horizon should exhibit the characteristics of the material that the soil is formed from (the SPM), and where leakage from more recent OC will be minimal as it is below the solum. The exact age of 'ancient OC' is unknown as it will have the potential, particularly through glacial and fluvial processes to come from many sources. That from eroded and weathered sedimentary rock will reflect the age of that rock. 'Ancient' OC is of interest, due to its ubiquitous presence in many SPM's described within this review such as till and glacial deposits, where it exists as the diluted OC concentrations of the eroded rock and pre-Holocene soil, with organic rich sedimentary rocks producing SPM deposits with greater OC. 'Ancient' OC may consist of some kerogen (Tissot and Welte, 1978; Hedges, 1992) formed during sedimentary diagenesis from plant, microbial, algal, planktonic or marine organic matter leading to the isomerization and aromatization of organic matter (Hood et al., 2015), and also will likely comprise a range of high molecular weight compounds of carboxyl and aromatic C (e.g. hydrocarbons), along with some phenol compounds, tannins, lignins, chitins and polysaccharides (Dümig et al., 2012). Typically, the incorporation of 'ancient OC' into SPM leads to low (< 0.5%) background concentrations in the parent (C horizon) material (Dümig et al., 2012; Eckmier et al., 2013; Lafreniére and Sharp, 2004; Yang et al., 2020). The OC in Holocene deposits of interests (alluvium, colluvium) are defined for the purposes of this review as 'recent OC'. This will predominantly be made of OC that is up to ~10,000 yrs. old (although it may still contain the 'ancient OC' from the eroded material pool which forms the SPM).

4. Deposits of interest, their formation and spatial distribution

A diverse array of processes leads to OC burial across landscapes. A multidisciplinary understanding of the particularities associated with each of these affords us a more holistic and precise understanding of where and how OC is buried, the thickness and properties of the burial material, and how susceptible buried OC is to being reincorporated into the global C cycle. Here, we review four key geomorphological realms – glacial, fluvial, aeolian and colluvial – each of which can entrain, transport and deposit sediment to bury OC.

4.1. SPM deposits formed from glacial deposits

Glacial processes are recognized in the development of significant SPM throughout the UK (Catt, 1979). Within the UK-LGM limits, glacial related deposits are typically the most spatially widespread and Devensian glacial till is ubiquitous (Fig. 1a; Table 2), its extent representing the maximum ice coverage with respect to the spatial retreating and re-advances of the ice sheet during MIS2 (Fig. 2). Previously glaciated landscapes also contain large areas of hummocky terrain and characteristic features such as moraines and glacial fans. These are often described in geological mapping as diamicton (poorly sorted deposits) and are often distributed on valley sides and bottoms.

4.2. SPM associated with glacio-fluvial, glacio-lacustrine and fluvial processes

Devensian glacio-fluvial sands are products of deposition by glacial meltwaters and often exist as sheet or delta like deposits of sand and gravels to form SPM. One of the major areas of mapped glacio-fluvial sands in the UK is in the Humber Estuary region of the UK (McEvoy et al., 2004), and their spatial distribution is consistent with the LGM limits (Fig. 1c; Table 2). Other deposits associated with glacio-fluvial processes are glaciolacustrine lake muds with a major deposit being found at the site of the Devensian Lake Humber / Fenland (Bateman et al., 2008).

The most widespread glacio-fluvial deposits forming SPM are river terraces created via the interplay between GIA and RSL, along with climate-driven surface processes, moderated by vegetation. Climate and vegetation is reflected in terrace development by the nature and OC content of the sediments. For the purposes of this review, interest focuses on terraces that date back to Marine Isotope Stage (MIS) 2 (11–24 K yr BP), many of which were cut and partially infilled by sand and gravel sequences during the later MIS 2. Ages are often <15 K yr BP (Lewin and Gibbard, 2010). They now predominantly underlay Holocene alluvium (e.g. Brown et al., 1994). The continued existence of these

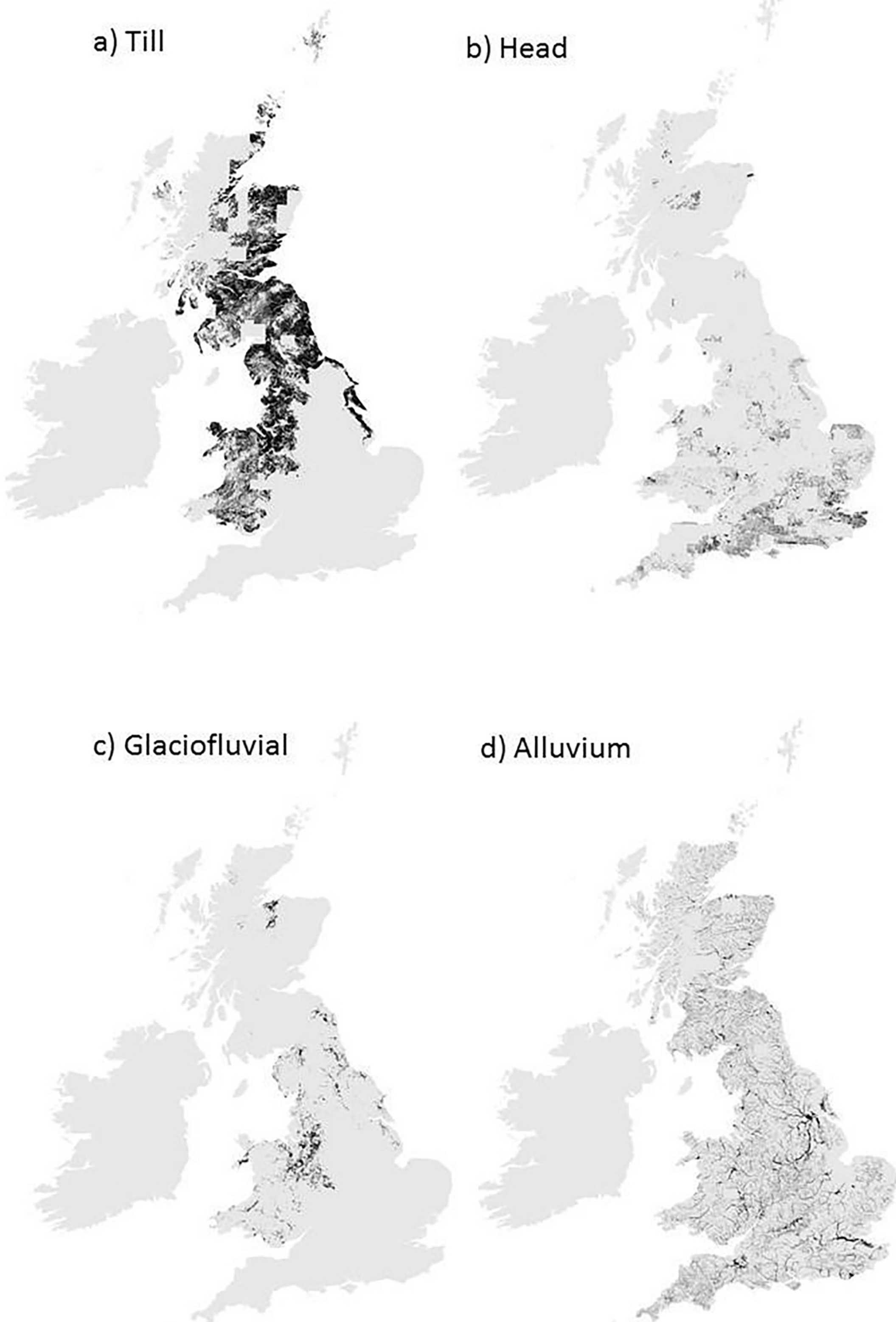

Fig. 1. Spatial distribution of (a) Devensian till, (b) Devensian head, (c) Devensian glacio-fluvial and (d) Holocene alluvium deposits across Great Britain as mapped by the British Geological Survey a scale of 1:50 K.

terraces implies that they have survived the transition from post-glacial to Holocene river systems (e.g. Lewin and Gibbard, 2010; Brown et al., 2009). The mechanisms of formation suggest a climate–vegetation response; reduced vegetation cover during colder phases increases the susceptibility of erosion, leading to increased sediment supply, while seasonal increases in fluvial discharge associated with spring thaw amplifies sediment transport capacity (Bridgland, 2000). The relative position of catchments in relation to the margins of the UK-LGM ice sheet is

Table 2
Area of coverage of post UK-LGM Devensian and Holocene deposits based on geological mapping undertaken by the British Geological Survey.

Deposit	Area (km^2)	Area (%)
OS high water polygon area of GB	230,050.13	100.000
Head deposits	4897.50	2.129
Colluvium Flandrian	74.85	0.033
Loess and Blown Sand	1040.57	0.452
Alluvium	11,126.75	4.837
Peat	12,101.94	5.261
Devensian (MIS2) Glaciofluvial	2784.98	1.211
Devensian (MIS2) Till	39,833.03	17.315
Devensian (MIS2) River Terraces	Not Calculated	-

a critical factor in determining the nature of the terraces and modern floodplains. Bridgland et al. (2010) reported on the neighbouring Trent and Ouse river catchments. Both discharge into the Humber Estuary but exhibit markedly different terrace sequences. The Trent catchment was unglaciated during the Late Devensian and contains a well-constrained terrace archive that extends back into the Middle Pleistocene (Howard et al., 2007; White et al., 2010). By contrast, the Ouse catchment was glaciated during the Late Devensian and contains one valley bottom terrace that forms a basal lag beneath the modern Holocene alluvium. These scenarios are common for adjacent river systems that straddle the Late Devensian ice limit. The area of MIS 2 terraces has not been calculated for GB, as in geological databases they are part of a wider grouping of Quaternary Terraces, reflecting the complexity of dating often partially eroded, and inconsistent sediment deposits.

Holocene alluvium deposits provide direct evidence of active soil erosion and deposition. They overlay late Devensian terrace or early Holocene basal deposits (Lewin and Gibbard, 2010; Brown et al., 2009) and their development as modern floodplain deposits on MIS 2 terraces increased rapidly ~1 K yr BP, coinciding with the agricultural revolution of the Middle Ages both in the UK (Macklin et al., 2010) and NW Europe (Hoffmann et al., 2008). Holocene alluvium deposits based on BGS maps (Fig. 1d; Table 2) cover 11,120 km^2. Dated alluvial sediments provide time-constrained information on the changing rates of erosion in a catchment. Brown et al. (2013) suggest that since catchment size has been maintained during the Holocene, changes in floodplain deposition rates must reflect the difference in the inputs and exports from that reach. Thus, changes in the rate of sediment deposition must then reflect changes in the nature of sediment inputs. Patterns of floodplain development and alluvium deposition are also complex, often resulting in OC rich abandoned paleochannels (Brown et al., 1994) and roddens (Smith et al., 2012). The widespread deposition of alluvium has been dated for many rivers. For example, in lowland England, dates of non-palaeo-channel alluvium floodplain deposition started ~3.5–2.1 K yr BP for the Soar and Nene rivers (Brown et al., 1994), reflecting the stabilization of the Holocene river systems.

4.3. SPM deposits associated with slope processes

A variety of slope mass movement processes have formed SPM and buried OC since the UK-LGM. Slope processes are often cyclical and reflect continual interactions of climate, vegetation coverage and later anthropogenic impacts. 'Head Deposits' are a cold climate, mass movement deposit dominant across the UK, being formed where gelifluction, solifluction, and frost creep processes were active in transporting sediment downslope. Harrison et al. (2010) describe solifluction sheets as the 'most widespread sediment landform assemblage on the lower slopes of valleys in the deglaciated areas of Britain'. Although they can take the form of isolated patches, solifluction sheets can be near continuous debris mantles up to 300 m wide, covering large areas of valley floors and lower slopes, as mapped by the BGS (Fig. 1b; Table 2). Many of the reported dates for Head deposits coincide with the Younger Dryas (cooling phase) and have depths >1 m. In addition, recognized erosion processes such as soil creep, sheet-wash, and gully erosion also contribute to the accumulation of these slope deposits. Typically, deposits consist of admixtures of gravel, sand, silt and clay, which are graded through the transport process. Hutchinson and Coope (2002) examined a solifluction sheet in southern England dated to the Younger Dryas. These formed under conditions of bare soil with sparse vegetation and an annual mean temperature of −3 °C, with it forming over successive summers when the active layer was mobile. Periglacial mudsliding may have added further contributions.

Whereas head deposits exist as major individual deposition events across the UK, they also form part of longer-term cycles of erosion, where climate and anthropogenic impacts interact. Two geographical areas in the UK have been studied extensively to reveal the cyclical nature of erosion; the valley infills on the chalk downs in southern Britain (Kerney, 1963; Kerney et al., 1964; Preece and Bridgland, 1999) and gully erosion in the uplands of Britain, particularly around the Solway Forth (Chiverrell et al., 2007). Chalk valley infill often demonstrates a sequence of climate and anthropogenic signals. An exemplar site was logged at Holywell Coombe (Preece and Bridgland, 1999). They estimated that Late Glacial aged deposits accounted for over 70% of the infill. Initially, during the post UK-LGM period (up to ~13 K yr BP), chalk clasts accumulated in many valley bottoms through intense periglacial slope activity. In the subsequent Bølling interstadial (July max temp 17–18 °C, minimum winter −2 to −7 °C; 13–12 K yr BP; (Coope, 1998)), humic chalk-rich and silt muds formed on the valley floors suggesting the development of marshy environments, with clay from landslides impeding drainage. These thin muds were subsequently overlain by a series of thick chalk muds and silts caused by solifluction when the climate deteriorated at the end of the Bølling period, prior to the start of the cooler Older Dryas. As the temperature increased again during the later Allerød interstadial (July max 13–15 °C and winter minimum −18 to 4 °C, ground temperature ~ 0 °C for much of the year; peak ~13 K yr BP), a paleosoil formed which has been widely recognized across the chalk downs (Evans, 1966; Peake, 1971; Kerney, 1963; Preece, 1994). Burial of the Allerød soil by solifluction deposits occurred during the Younger Dryas stadial. Whilst initial deposits in the valley bottoms are pre-Holocene, later sediment deposits are associated with Holocene land-use change and particularly the Neolithic and Bronze Ages forest clearances (Preece, 1993). The deposits found at Holywell Coombe are considered representative of the wider chalk downland in southern England, although considerable local variability in the spatial erosion patterns is likely. Wilkinson (2003) reviewed seven chalk dry valley systems and concluded that whilst the patterns of sedimentation can be similar, the mechanisms and chronology of deposition vary between valleys. Factors influencing the deposition of sediments are likely to include sediment texture, variations in past land-use, the impact of storm activity, and local topography.

Considerable work regarding slope processes has also been undertaken in the areas (e.g. Howgill Fells, Bowland Fells) between North West England and Southern Scotland (e.g. Chiverrell et al., 2007). Three widespread landforms that may bury OC result from interactions between hillslopes and streams; gullies, alluvial fans/debris cones, and valley-floor deposits. Hillslope gullies dissect steep, glacially or periglacially-modified hill slopes. In the Howgill Fells in North-West England, a representative area of the regional uplands, two types of hillslope gullies are present; older stabilized gullies and currently active gullies. Both have formed in previously deposited solifluction deposits and are primarily related to basal stream activity, either through lateral erosion by the main streams or incision of steep tributary channels. Harvey (1992) examined modern gully formation to understand the mechanisms of formation. Eroded sediment is deposited in storage areas (such as gully channels or basal zones). Storage in the basal zone often results in the formation of debris cones at the foot of the gully or as debris aprons below open slopes. The periodic removal of the sediment by axial streams, particularly during storm events leads to the rejuvenation of erosional activity on the slopes. Thus, cyclical erosion is

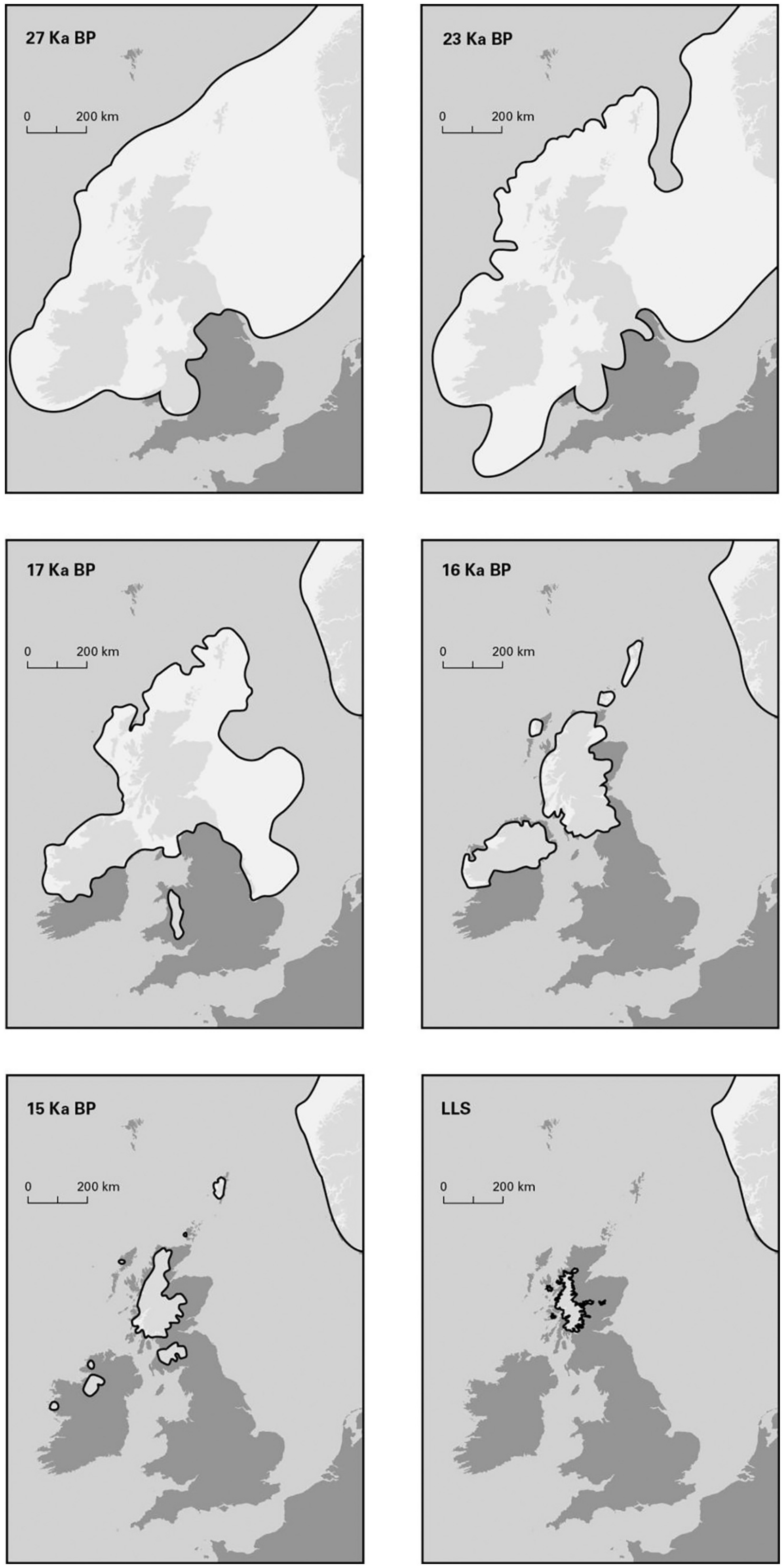

Fig. 2. Descriptions of the extent of ice cover at various times since the Last Glacial Maximum 27 Ka BP. Position of ice sheet is obtained from descriptions of Clark et al., 2012.

initiated by intrinsic channel migration or storm induced erosion. However, once the relationship with the axial streams is severed through sediment accumulation or channel migration, the gullies become vegetated and stabilize within a maximum period of c. 150 yr (Harvey, 1992; Harvey, 2001; Chiverrell et al., 2007). The formation of alluvial fans and debris cones at the junctions of stream tributaries in the Bowland Fells have been dated to between 5400 and 1900 and ~ 900 BP using ^{14}C (Harvey and Renwick, 1987).

Finally, since the Mesolithic-Neolithic transition (approx. 7 K yr BP), soil erosion (colluvium) has increased due to vegetation clearance and the intensification of agriculture. It is a widespread process and several attempts to classify the severity of potential current soil erosion have been made (e.g. Morgan, 1985; Evans, 1990), the highest categories being linked with soils of sandy or silty texture, often with the SPM, being formed from previously deposited sediment (e.g. loess, coversands).

4.4. Wind-blown erosion and deposits

Wind-blown sediments, including loess and cover-sand deposits, cover a combined area of 1040 km^2 or 0.45% of the area of Great Britain (Table 2). The extent to which they act as SPM depends on initial deposit thickness, texture and later redistribution. Loess deposits typically have a clay content <20%, and a sand content of not >2%, with the remainder being silt. Widespread aeolian transport of loess requires a large supply of source material (produced by glacial grinding), dry periglacial conditions and minimal vegetation. Great Britain was situated on the western edge of the European loess belt. The primary loess deposition event was in the late Devensian (~18 K yr BP) and was considered to have covered large parts of the landscape of England. Jefferson et al. (2003) estimated that loess deposited across southern and midland England was 20 to 50 cm thick; this being based on the volume of sediment currently making up reworked large loess bodies in southern England. Some original loess deposits are still found. On the Chilterns Hills, a distinct thin loess deposit of ~0.3 m covers the Plateau drift, whilst on the North and South Downs, loess deposits of 0.4–0.9 m can be found (Catt and Hodgson, 1976). However, most remaining loess represents reworked deposits formed after widespread erosion and fluvial transport of the initial loess covering, particularly within the Wealden loess trap (Catt, 1977; Catt, 1978). The spatial distribution of loess has been mapped and the major remaining deposits are found in the south-eastern counties of England (Antoine et al., 2003; Clarke et al., 2007; Scheib and Lee, 2010), with smaller deposits found in Devon and Cornwall (Roberts, 1985; Cattell, 1997). Knowledge of their depth is generally poor but reworked loess deposits in Kent and Essex can be up to 4 m in depth (Pitcher et al., 1954; Northmore et al., 2008).

Coversand deposits differ from loess, being coarser and sourced from glacial outwash deposits located in the Southern North Sea Basin, along with deposits located around the margins of the ice sheet such as those from Lake Humber (Bateman, 1998). Coversands are prominent in Norfolk and Lincolnshire (Catt, 1977; Scheib and Lee, 2010) and redistribution has been recorded.

5. Evidence of the interactions of climate, vegetation and deposit age

Vegetation limits both the extent of erosion and SPM formation, along with influencing OC concentrations of buried deposits. Pollen analysis has helped identify the broad scale vegetation changes over the time-period of this review (Pennington, 1969), of which five vegetation phases appear of importance since the UK-LGM. First, SPM directly related to glacial climates (e.g. tills, moraines, diamicton, terraces, loess) reflect low NPP environments with vegetation assemblages associated with those of tundra like conditions. However, based on current knowledge of arctic environments, when NPP is low (particularly during MIS 2), considerable OC may be stored within some areas of tundra and boreal forests, likely due to slow decomposition rates (Jobbágy and Jackson, 2000). However, knowledge of the extent of OC distribution across the UK during the early climatic periods under review, and how this may have influenced OC burial is unknown. Within the post UK-LGM to Holocene period, the warmer Allerød interstadial occurred where NPP increased. Collins et al. (1996) described floodplains populated with birch and grassland, which later disappeared during the Younger Dryas. The third phase represents the period when vegetation stabilized the environment during the warming climate of the early Holocene (Ballantyne, 2008). Landscape stabilization took about ~3 K years in Scotland and Ireland (Edwards and Whittington, 2001). A fourth phase reflects a period with relatively stable vegetated environments where erosion is likely to have been low. The fifth phase represents the onset of Neolithic (~6000 yrs. BP) agricultural intensification and land-use change where vegetation coverage was impacted at different spatial and temporal scales leading to widespread soil erosion. It is likely that land-use change and increased soil erosion reflect both the extent to which land was cultivable, but also regionally (e.g. N-S) where, for example, cooler climates reduce NPP, increasing recovery time after grazing, with the most intensive grazing practices leading to poaching and the creation of erosion pathways such as cattle tracks.

The extended C-V-E hypothesis is used as a concept to assess when SPM formation and subsequent OC burial took place in relation to these five phases following the UK-LGM. The occurrence of dated (^{14}C or OSL) erosion events in climate periods and later human impacts, as well as OC burial depths, are reported in Tables 3-6.

5.1. The UK-LGM to the Holocene

Four major SPM depositional events are identified in this period (Tables 3 & 4). First, ice retreat (see Fig. 2), including localized re-advances and retreats due to warming starting ~27 K yr BP (Clark et al., 2012), resulted in the widespread erosion and remobilisation of glacial sediments. Till deposition is also associated with ice re-advance and retreat during the Younger Dryas (Ballantyne, 2008; Bickerdike et al., 2018). Low NPP in cold climates and an abundance of unconsolidated material were essential in SPM being created during the MIS stage 2 river terrace formation. Various terraces have been identified as being post UK-LGM and of MIS 2 age between ~27 and 11.5 K yr BP including on the Exe between 20 and 13 K yr BP (Brown et al., 2010) and the Thames (Maddy et al., 2000; Maddy et al., 2001), although some may now be submerged in the offshore area (e.g. Westaway et al., 2006). Terrace development has also been identified during the cold Younger Dryas including on the Exe (c.13–11.5 k BP) (Brown et al., 2010) and the basal gravels of the Nene (11.2–10.2 K yr BP and the Soar 28–10.2 K yr BP) (Brown et al., 1994). Third, low NPP and plant cover facilitated the major period of loess and cover sand deposition across parts of the UK between 25 and 16 K yr BP. Dated loess deposits fall into three periods, coinciding with the late Devensian ice sheet (~18 K yr BP; 14 K yr BP; 11 K yr BP). Singhvi et al. (2001) suggested that primary loess deposition stopped in Northern Europe ~15 K yr BP, confirming the consensus that one primary loess deposition phase occurred in England in the late Devensian (Catt, 2008). Later dates of deposition therefore likely represent hillslope and re-worked fluvial deposits.

The Heinrich events in the UK were considered dry and cold; ideal for the distribution and redistribution of sediment through aeolian transport. For example, coversand deposition in Norfolk occurred during stadials (MIS 5d–MIS 2), but particularly during the Younger Dryas (~12.9–11.55 K yr BP) (Bateman et al., 2014). The initial deposition of the Breckland coversand in Norfolk was approximately 12.82 ± 1.91 K yr BP at Leziate (Hoare et al., 2002), whilst Bateman et al. (2000) identified three periods of coversand deposition near Caistor, Lincolnshire, at ~22, 16 and 6 K yr BP.

Table 4 describes the depth at which dated OC is found within a range of erosion deposits cited in the literature. Widespread buried OC coincides with the warmer Allerød interstadial. Coversand and loess

Table 3
Examples of erosion deposits and ages from the UK-LGM to Holocene.

Category	Event	Reference	Age range (K yr BP)	Period name
Till	Major period of BISS ice retreat leaving Till	McMillan et al., 2011	27 –15	Late Devensian
Till	Major period of BISS ice retreat leaving Till	Clark et al., 2010	27–15	Late Devensian
Till	Probable start of Younger Dryas Stadial glacial retreat leaving Till	Ballantyne 2012	12.4–12.1	Younger Dryas
River Terraces	Formation of MIS 2 Terrace 2 – River Exe	Brown at al. 2010	20 –13	Late Devensian
River Terraces	Formation of LLS Terrace 1 – River Exe	Brown et al., 2010	13–11.5	Younger Dryas
River Terraces	Formation of MIS 2 Holme Pierrepoint Terrace, River Trent	Howard et al., 2011	11.25 – 10.93	Younger Dryas
River Terraces	Thames MIS2 terraces (submerged)	Maddy et al., 2001	18	Late Devensian
Aeolian	Period of initial Loess Deposition, GB	Catt, 2008;	26–13	Late Devensian
Aeolian	Loess deposition, Cornwall	Scourse, 1999	15.9 +/- 3.18	Late Devensian
Aeolian	Loess deposition, Isles of Scilly	Scourse, 1991	18.6 +/– 3.7	Late Devensian
Aeolian	Loess deposition, Isles of Scilly Loess deposition, Isles of Scilly	Smith et al., 1990 Smith et al., 1990	20 +/– 7 26+/– 9.5	Late Devensian
Aeolian	Loess deposition, Heathrow Loess distribution, Heathrow	Rose et al., 2000 Rose et al., 2000	17.8+/–1.5 14.3+/–1.2	Late Devensian
Aeolian	Loess solifluction deposit	Wintle & Catt, 1985	17.5+/–1.6	Late Devensian
Aeolian	Loess deposition, Kent	Clarke et al., 2007 Clarke et al., 2007 Clarke et al., 2007 Clarke et al., 2007	18.7 +/–2.29 23.76 +/–1.3 14.99+/–0.86 17.21+/–1.3	Late Devensian
Aeolian	Loess redistribution, North west England	Vincent et al., 2011	14.5–11.1 (3 samples)	Late Devensian
Aeolian	Coversand deposition, Lincolnshire	Bateman et al., 2000	16.4 – 13.8 (4 samples)	Late Devensian
Aeolian	Coversand deposition, Thanet, Kent	Murton et al., 2003	24–21 16.43–14.85 (2 samples)	Late Devensian
Aeolian	Breckland Coversand	Bateman, 1995	12.59–15.98 (2 samples)	Late Devensian
Aeolian	Coversand deposition	Bateman, 2000	12.62+/–1.3 12.78+/–1.2 11.85+/–4.6 12.2+/–1.2	Younger Dryas
Aeolian	Sand filled stripe, Thanet, Kent	Murton et al., 200;	12.23+/–0.39	Younger Dryas
Aeolian	Leziate coversand	Hoare et al., 2002	12.83+/–1.9	Younger Dryas
Aeolian	Breckland Coversand	Bateman, 1995	13.38+/–0.79 14.58+/–1.4 (2 samples)	Late Devensian
Mass Movement	Head -Paraglacial event	Harrison et al., 2010	9.87+/–1.64	Younger Dryas
Mass Movement	Head -Paraglacial event	Hutchinson, 2010	12.7 – 10.3 (inferred) LLS	Younger Dryas

Links to references: Ballantyne, 2012; Bateman, 1995; Howard et al., 2011; Hutchinson, 2010; Murton et al., 2003; Rose et al., 2000; Scourse, 1991; Scourse, 1999; Smith et al., 1990; Wintle and Catt, 1985.

Table 4
Examples of buried OC with ^{14}C dating in the period post UK-LGM to Holocene.

Category	Event	Reference	Age range (K yr BP)	Name	Depth of OC (m below surface) and % OC if data provided
Mass Movement – Buried OC	Late Glacial valley infill - Chalk - Buried soils – different cores	Preece & Bridgland, 1999	11.37+/–0.15 11.52+/–0.90 11.53+/– 0.16 9.76+/– 0.1 (4 samples)	Younger Dryas	1.5 1.6 3.2 3.5
Mass Movement – Buried OC	Late Glacial valley infill - Chalk - Organic silt and chalk muds	Preece & Bridgland, 1999	13.16+/–0.4 to 11.83+/–0.14 10.16+/–0.11 to 9.82+/0.9	Younger Dryas	4.2 3.4
Mass Movement – Buried OC	Late Glacial valley infill - Chalk - Organic detritus muds	Preece & Bridgland, 1999	12.28+/–0.14 9.53+/–0.75 9.76+/–0.1 (2 samples)	Younger Dryas	4.3 2.3 3.5
Buried soils and organic carbon	Buried soil under Devensian Till	Merritt et al., 1989	12.75 +/– 70	Allerød	1.7–3.85
Buried soils and organic carbon	Allerød organic rich soils / peats	Bateman et al., 2000	13.85+/–0.1 12.49+/–0.1 (2 samples)	Allerød	2.1 1.1
Buried soils and organic carbon	Allerød organic silty sand	Collins et al., 1996	12.77–13.05	Allerød	2.0
Buried soils and organic carbon	Allerød humic horizon in slope deposits, Ventnor, IoW	Preece et al., 1999	11.69 +/– 0.12	Allerød?	0.95–1.05 ~0.5% OC
Buried soils and organic carbon	Allerød soil, Armthorpe, Vale of York	Gaunt et al., 1971	11.13–10.86 IntCal13 corrected	Allerød?	0.67

Links to reference: Gaunt et al., 1971.

often exist as a series of layers separated by palaeosols or peat layers, thus defining hiatuses in deposition or re-distribution when stabilized land surfaces developed through warmer temperatures promoting NPP. A prominent peat layer found within these deposits, with ^{14}C dates of 14–12 K yr BP, is the Allerød peat. Buried Allerød OC deposits are also found within the cycles of mass movement deposits such as those found in chalkland valleys (Preece and Bridgland, 1999) and organic rich deposits in some river floodplains due to the increased NPP of this period. One example is the Wasling Sand from the River Kennet (Collins et al., 1996). These buried soils are likely a widespread occurrence across Great Britain, although few deposits have been dated. The fourth large erosion event was in the Younger Dryas when Head deposits were formed (e.g. Harrison et al., 2010), in cooler and lower NPP climates. Table 4 demonstrates that the depths of OC burial can extend down to 4 m.

5.2. Early to Mid-Holocene

During the Holocene, Bond (cooling) events have occurred within a generally warming climate (Table 1), along with possible 'flooding events' which were identified through increased soil erosion manifesting itself as floodplain soil accumulation (Benito et al., 2015). Table 5 summarizes dated erosion events during the Holocene. Evidence suggests that the first part of the Holocene was a period of landscape stabilization and soil formation, with few recorded dated erosion events; with the exception of those soil erosion events which helped identify the previously described 'flooding events'. Edwards and Whittington (2001) examined sediment accumulation in 50 different lakes in Britain and Ireland, identifying reversals in ^{14}C dating in sediment cores as a signal of terrestrial and older soil carbon input. The results indicated that many of the lakes in north and northwest Scotland displayed constant or decreasing sedimentation rates over time. This was attributed to: (i) thin soils forming on hard bedrock, (ii) the sealing effect of blanket peat, and (iii) low population density allowing for a maturation of the landscape where vegetation stabilized the landscape and soil. However, some evidence suggests that aeolian re-distribution events (Table 5) coincided with the cooler Bond events (Table 1). Vincent et al. (2011) examined the age of loess redistribution deposits on the limestone uplands of north-west England using OSL dating where 9 out of 15 dates were coincident with a hypothesized climatic deterioration at 8.5–8.0 K yr BP.

Table 5
Examples of erosion deposits and ages from the Holocene to present.

Category	Event	Reference	Age range (K yr BP)	Coinciding Time period	Depth of OC (m below surface) and % OC if data provided
Soil Erosion	Flooding events indicating increased soil erosion	Benito et al., 2015	10.7 –10.4 9.4 – 9.1 7.8 – 7.7 7.6 – 7.3 6.9 – 6.5 6.3 – 6.2 6.0 – 5.7 5.3 – 5.1 4.9 – 4.5 4.3 – 4.2 3.6 – 3.4 2.9 – 2.8 2.3 – 2.0 1.5 – 1.1 1.1 – 0.5 0.3 – 0		
Soil Erosion	Scottish and Irish Lake Sedimentation cluster	Edwards & Whittingham 2001	5.29 – 4.97	Benito et al., 2015 Flooding Period	
Soil Erosion	Lake Sedimentation cluster	Edwards & Whittingham 2001	4.53 – 4.23	Benito et al., 2015 Flooding Period	
Soil Erosion	Lake sedimentation cluster	Edwards & Whittingham 2001	2.98 – 2.81	Benito et al., 2015 Flooding Period	
Mass Movement	Gully formation NW England & Scotland	Chiverell et al., 2007 & refs therein	2.5 –2.2	Benito et al., 2015 Flooding Period	
Mass Movement	Gully formation NW England & Scotland	Chiverell et al., 2007 & refs therein	1.3 –1.0	Benito et al., 2015 Flooding Period	
Mass Movement	Gully formation NW England & Scotland	Chiverell et al., 2007 & refs therein	1.0 –0.8	Benito et al., 2015 Flooding Period	
Mass Movement	Gully formation NW England & Scotland	Chiverell et al., 2007 & refs therein	0.5	Benito et al., 2015 Flooding Period	
Mass Movement	Alluvial fan / debris cones	Harvey and Renwick 1987	5.4 – 1.9 ~0.9		Various depths down to 1.8m
Aeolian	Loess redistribution, North west England	Vincent et al., 2011	10.4 – 6.1 (12 samples)	Holocene	

5.3. *Mid Holocene to present day*

In the second part of the Holocene soil erosion is a major contributor to OC burial, with identifiable erosion episodes forming the basis of the ‘Flooding Events’ concept, along with human impacts. Table 6 provides information regarding specific dated buried OC deposits and the depth at which they were buried, with depths extending to ~4 m. Prior to the Benito et al. (2015) study, a significant database of UK ^{14}C dated sediments (Macklin and Lewin, 2003; Macklin et al., 2012; Macklin et al., 2013) and extensive meta-analyses had enabled the production of national and regional reconstructions of anthropogenic alluvium deposition (Macklin et al., 2014). A review of dated deposits suggests some soil erosion and mass transport events may have coincided with Bond events at 5.4 and 4.7 K yr BP (e.g. see Curry, 2000; Benito et al., 2015). However, whilst these Bond events were not noted for specific climate cooling (Table 1), they often partially or fully overlapped with dates of

Table 6
Dated Records of organic carbon burial during the Holocene.

Category	Event	Reference	Age range (Ka BP)	Coinciding Time period	Depth of OC (m below surface) and % OC if data provided
Buried soils and organic carbon	Mass Movement buried organic detritus muds in Chalk valley	Preece & Bridgland, 1999	7.65+/–0.08 7.50+/–0.100	Possible Flooding Period (Benito et al., 2015)	1.8 1.6
Buried soils and organic carbon	Late Glacial Chalk valley infill - Buried soil	Preece & Bridgland, 1999	9.76+/– 0.1	??	3.4
Buried soils and organic carbon	Soil erosion Chalk valley - Mass movement buried A horizon	Recio-Espejo et al., 1992;	1.86+/– 0.08	??	0.46 – 0.66
Buried soils and organic carbon	Breckland, Norfolk aeolian Buried organic rich sand	Bateman & Godby, 2004	2.11+/–0.06	Possible Flooding Period (Benito et al., 2015)	0.80 – 1.1 LOI (%) = 4.1
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Western Red Hills	Curry, 2000	0.3 – 0	Possible Flooding Period (Benito et al., 2015)	0.41
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Western Red Hills	Curry, 2000	2.752 – 2.404	??	1.13
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Western Red Hills	Curry, 2000	0.692 – 0.564	Possible flooding Period (Benito et al., 2015)	0.6
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Western Red Hills	Curry, 2000	0.948 – 0.736	Possible Flooding Period (Benito et al., 2015)	0.74
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Western Red Hills	Curry, 2000	1.387 – 1.269	Possible Flooding Period (Benito et al., 2015)	0.83
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Western Red Hills	Curry, 2000	2.061 – 1.821	??	1.09
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Western Red Hills	Curry, 2000	3.680 – 3.398	Possible Flooding Period (Benito et al., 2015)	1.35
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Glen Doherty	Curry, 2000	0.456 – 0.5	??	2.76
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Glen Doherty	Curry, 2000	4.834–4.454	Possible Flooding Period (Benito et al., 2015)	3.69
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Glen Doherty	Curry, 2000	4.871–4.658	Possible Flooding Period (Benito et al., 2015)	1.11
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Glen Doherty	Curry, 2000	5.607–5.330	??	1.13
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Glen Doherty	Curry, 2000	6.490–6.210	Possible Flooding Period (Benito et al., 2015)	1.41
Buried soils and organic carbon	Hillslope debris flow - buried palaeosols – Pass of Drumochter	Curry, 2000	0.623–0.499	Possible Flooding Period (Benito et al., 2015)	0.70
Buried soils and organic carbon	Hillslope debris flow - buried palaepodzols	Reid & Thomas, 2006	5.450+/–45	??	0.8– 0.78
Buried soils and organic carbon	Hillslope debris flow - buried palaepodzols	Reid & Thomas, 2006	4.560+/–75	Possible Flooding Period (Benito et al., 2015)	0.95–0.92
Buried soils and organic carbon	Hillslope debris flow - buried palaepodzols	Reid & Thomas, 2006	1.820+/–45	??	0.60–0.58

'Flooding Events', suggesting increasing difficulties in ascribing erosion to specific climate events. Later Bond events may not have been sufficiently long or intense enough to cause major changes to vegetation cover or that they instigated cooler, yet drier periods. For example, whilst the identification of the 4.2 ka BP event has been observed globally), there is no coherent record of it in the UK and Ireland, suggesting that the atmospheric-oceanic circulation changes produced did not cause a measurable impact or it has not been preserved due to anthropogenic activity (Roland et al., 2014). Evidence of a 2.8 K yr BP event within a period of considerable climate change between c.3.25 to 2.6 K yr BP has been suggested (Roland et al., 2014), characterized by a shift towards wetter conditions, proceeded by a notably drier period, but to the best of our knowledge no direct erosional evidence has been demonstrated to date in the UK.

Prior to the recognition of the 'flooding events of' Benito et al. (2015), several studies had alluded to 'fluvial events' being triggers for increased soil erosion, particularly in northwest England and Scotland. For example, Reid and Thomas (2006) suggested that during the mid- and late Holocene, individual storm events, combined with anthropogenic vegetation removal, were the dominant slope processes, creating paleo-podzols in valleys, whilst Tipping (1995a, 1995b) recorded a series of colluvial and stream in-washings of hillslope sediment to a peat-filled valley bottom at Carn Dubh in Scotland. In addition, buried vegetation underneath solifluction deposits in the Caingorms in Scotland (e.g. Ballantyne, 1986; Sugden, 1971; Mottershead, 1978) have been ^{14}C dated and broadly align with phases of greater precipitation in Scotland (e.g. Tipping, 1995a, 1995b).

While many late Holocene erosion and mass movement dated events partially or fully overlap with 'Flooding Events', many events, particularly on slopes, do not coincide and demonstrate the impact of Neolithic and later deforestation and land-use change processes in the UK. Edwards and Whittington (2001) identified three clusters of increased sediment accumulation in lake cores between 5.29 and 4.97, 4.53–4.23 and 2.98–2.81 K yr. The authors suggest that these periods broadly link to: (i) intensification of Neolithic agriculture in northern Europe and the decline of Elm trees at ~5.1 K yr, (ii) continued expansion of Neolithic agricultural practices and (iii) Bronze Age tree clearances, respectively. However, these dates also partially coincide with 'flooding events' which suggests a possible combined climate and land-use effect.

Further examples include gulley formation in northwest England and the Southern Uplands of Scotland, where the formation of the older stabilized gullies have been heavily linked to land-use and vegetation change. Cundill (1976) suggested that soil erosion was linked to cyclical episodes of major woodland clearances in Howgill Fells, Cumbria. These episodes included ones at 3.84 K yr BP initiated by Bronze age humans, 2.29 K yr BP possibly related to climate deterioration but more likely anthropogenic, after 2.29 K yr BP during the Roman period, and after periods of woodland regeneration during the Dark ages (1.6–1 K yr BP) and the Norse period (1.2–0.95 K yr BP). The latter period coincided with an increase in intensive Medieval sheep farming in the area. Harvey et al. (1984) dated paleosols within debris cones using ^{14}C in the Bowland Fells in Lancashire and Howgills Fells in Cumbria, and suggested deposition occurred between 2.71 and 2.50 K yr BP and ~ 1 K yr BP. Again, it is likely these were due to land clearances and management changes such as imported (intensive) Scandinavian sheep grazing practices. Chiverrell et al. (2007) summarized much of the work undertaken in the upland areas of NW England by collating ^{14}C data from buried OC in alluvial fan deposits from the upland regions surrounding the Solway Firth and Morecombe Bay to understand the widespread controls on gully formation. The ^{14}C data suggested that four main phases of gully erosion occurred between 2.5 and 2.2, 1.3–1.0, 1.0–0.8 and ~ 0.5 K cal yr BP. Whilst these periods often partially overlap with identified 'flooding events', high magnitude storm events are considered the primary driver for initiation of gully erosion. The absence of gully systems developed before 3 K cal. yr BP suggested that the increase in gullying after 2.5 K cal. yr BP was also linked to population expansion and deforestation during Iron Age and Romano-British times. In the highlands of Scotland, Curry (2000) studied debris flows on drift mantled hillslopes and found four reworking phases (6.5–6.2, 5.6–5.3, 4.9–4.6 and after 0.45 K yr BP). A re-evaluation of results in the context of Benito et al. (2015) suggest some overlap to 'flooding-events'. However, the date of the fourth event (0.45 K yr BP) is also coincident with the Little Ice Age suggesting a possible climate interaction along with deforestation. These recorded mid to late Holocene events, demonstrate the complexity of the interactions between possible 'flooding events' and land-use change and intensification on the initiation of soil erosion events in Great Britain. Human impacts on the redistribution of aeolian cover-sand deposits are also evident at Breckland in Norfolk, an inland dune system developed during the Anglian (400 K yr BP) (Bateman and Godby, 2004), where five episodic re-distribution events have been identified: c.6.50, c.1.60–1.10, c.0.50, c.0.40–0.33 and from c.0.20–0.30 K yr BP. It is likely that this redistribution resulted from interactions between intensive sheep grazing and possible climatic (Bond event) oscillations at 1.60–1.10 and 0.5 K yr BP.

Many Holocene OC burial processes associated with alluvium (floodplains) and colluvium (hillslope toes) reflect OC burial processes over a period of time. These deposits will, contain 'recent' OC with a range of ages reflecting the deposition period, with older OC at the bottom of the deposit, but also allow for vegetation inputs from the surface (as discussed in Section 3). In particular, mid-late Holocene erosion processes influenced by human activities are known to have changed the rate of sediment and carbon deposition on floodplains compared to earlier Holocene periods. A visible change in both sedimentation and OC burial is often seen and have been described as 'discontinuities of human-induced alluviation' (Brown et al., 2013). Many of these later 'flooding events' were described as being more related to increased human driven catchment erosion and infrastructure placement.

In the UK the major interest in floodplain carbon deposits has primarily been associated with ^{14}C dating for landscape evolution and later climate and anthropogenic impact studies (Macklin et al. 2014). Results have contributed into the interpretation of 'flooding events' described by Benito et al., 2015). However, Mid to late Holocene increases in sediment and OC burial associated with human impacts are often complex processes, both spatially and temporally. An indication of the variation of current sedimentation rates on floodplains was produced by Walling et al. (2006a, 2006b). They examined floodplain deposition of OC on six English rivers with sediments having mean concentrations of OC ranging between 2.17 and 5.07% OC, and sediment deposition rates of 69–114 g m^{-2} yr^{-1}. Greater interest in the OC budgets caused by human impacts have been undertaken in other countries, and many of these responses would equally apply to GB floodplains. In Europe and the US, late Holocene human impacts in catchments (e.g. deforestation, introducing infrastructure) have been found to influence sediment and OC burial rates on floodplains. Ricker et al. (2012b) suggested that in areas of fast accretion rates, where buried surface horizons existed, greater OC stores were found compared to slower accreting areas of the riparian zones. The accretion was driven by catchment deforestation (Ricker et al., 2012a, 2012b). Manipulation of river channels with dams or other infrastructure for water powered mills in the 17-19th century have also fundamentally changed the nature of floodplains in the US and Europe from small anabranching systems with extensive wetlands and large carbon stores to 'slack water' sedimentation systems that have buried the natural pre-human systems and made floodplains essentially into fill terraces. The final stages of floodplain change that impacts sedimentation and OC burial is seen in the installation of drainage systems and the canalisation of rivers (Walter and Merits, 2008; Brown et al., 2013; Brown et al., 2018).

6. The influence of the climate-vegetation-erosion pathway on deposit thickness

Interactions between climate and vegetation are fundamental controls on erosion within the C-V-E concept, in addition to other known variables involved in the supply and transfer of sediment including topography, sediment texture, sediment availability and transport capacity. These combine to determine erosion and deposition rates, leading to the spatial extent and thickness of deposits. The thickness of deposit is important with respect to buried carbon as it is (i) an indicator of the energy of the erosion process, sediment availability and likely OC concentration and source, (ii) it determines the depth at which OC may be buried and (iii) in some instances determines the depth to which 'ancient' OC is distributed through the SPM profile. For (i) above, there is an additional link between the amount of OC associated with different particle sizes transported, with clay particles usually having greater OC associated with them than sand or silt sized particles. Little work has been undertaken comparing the mean thickness of deposit types discussed in this review. Partly, this relates to the difficulty of constraining estimates of deposit thickness over sufficiently large areas. Fig. 3 reports post UK-LGM deposit thickness from the high-resolution 3D geological models at Talla (Scheib et al. (2008) and Shelford (Tye et al., 2011). In Fig. 4, deposit thickness from lower resolution 3D geological models provide a wider 'first-order' overview across Great Britain.

6.1. The UK-LGM to the Holocene

Deposits in this period reflect cold climate and glacial-related processes within generally low NPP environments, and are generally thicker than those of the Holocene. For example, at Talla (Fig. 3) average till thickness is ~6 m, whilst the moraine deposits are ~6 m. River terraces at Talla (within LGM) have a mean thickness of ~2.7 m, whereas the thickness of the MIS 2 terrace deposit at Shelford is 7.65 m. Deposit thickness in the low resolution 3D models (Fig. 4) for this period are broadly similar to those reported in the high-resolution models, with deposits, such as the Devensian till, having a median thickness of ~8 m. Analysis of the median potential till depth in MIS2 - MIS 1 deposits is ~6 m (McMillan et al., 2011). At Shelford, 'Head' deposits have a mean thickness of 0.56 m. The thickness of solifluction deposits reflect the paraglacial response of the landscape, where it is 'relaxing' and achieving a new dynamic equilibrium with slopes and rivers adjusting to enhanced sediment budgets, seasonal discharge regimes (i.e. seasonal melt) and low vegetation (Matsuoka, 2001).

6.2. The Holocene

As described previously, the early Holocene was likely a period of relative landscape stabilization, in contrast to the late Holocene, which has been dominated by land-use change and soil erosion, leading to colluvial deposits accumulating on hillslope toes, and the formation of soils on modern floodplains. The thickness of colluvial deposits are not well represented in the low-resolution 3D models, as deposits are thin and difficult to identify. However, in the Shelford model (Fig. 3) the average thickness at the bottom of the hillslope is 0.5 m.

Valley deposits with ^{14}C dates allow for comparisons between deposition events created through climate and those initiated by human impact, although slope characteristics (length, slope, aspect) and soil texture need to be considered. Dated deposits have been used to estimate Neolithic soil erosion, particularly on chalk downlands, where colluvial deposits represent the upper part of the record of erosion events to have occurred over time. Preece and Bridgland, 1999 found colluvial deposits up to 2.5 m thick from erosion events occurring between 8 and 3.5 K yr BP. Later erosion events (1.865 +/− 80 K yr BP (OxCal age from 1.989 to 1.610 K yr BP) have been dated to Roman times by Espejo et al. (1992) who measured deposition of 46 cm of a flinty silt loam in chalkland valleys, believed to be associated with the cereal cultivation required to feed the resident Roman army. Modelled estimates of the magnitude of soil erosion on the chalk downs of England were made by Favis-Mortlock et al. (1997) who assessed the impact of anthropogenic activity on the erosion of loess-based soils from 7 K yr BP to the present day. Results identified a major period of soil loss between 4 and 1.8 K yr BP as agriculture slowly became more intensive. An estimated initial loess thickness of 1.2 m was reduced to 0.58 m within the modelling and proved to be a reasonable prediction of present day soil depth and stone content. Ellis and Newsome (1991) suggested that soil erosion occurred on the chalk downs 2–3 K yr BP largely through cultivation. Removal from a gentle, upper valley slope over this period was estimated to be 375 kg m^{-2} or equivalent to a loss of 0.1–0.2 kg m^{-2} yr^{-1}.

Estimates of contemporary net erosion rates for UK soils measured using ^{137}Cs (Quine and Walling, 1991; Walling and Quine, 1991; Owens et al., 1997; Walling et al., 2002; DEFRA, 2005; Walling et al., 2006; Boardman et al., 2009) span across three orders of magnitude with a median erosion rate of ~300 t km^{-2} yr^{-1} for UK arable and grassland soils. Slope lengths ranged between 39 and 278 m with maximum angles between 2.1 and 23°. These estimates do not include erosion rates from catastrophic events such as muddy floods on the South Downs (Boardman, 1995) or gully erosion events (Evans and Nortcliff, 1978). The other major late Holocene deposit is alluvium. The thickness of alluvium can often be approximated by the depth of increased OC concentrations. Alluvium deposit thickness at Talla and Shelford suggest depths are typically <1 m (Fig. 3).

7. Organic carbon concentrations in deposits

There is evidence for two hypotheses to describe OC concentrations in deposits since the UK-LGM. First, Milankovitch-scale glacial driven processes have the potential to produce thick SPM (C horizon) deposits that have relatively low concentrations of OC, reflecting the low NPP

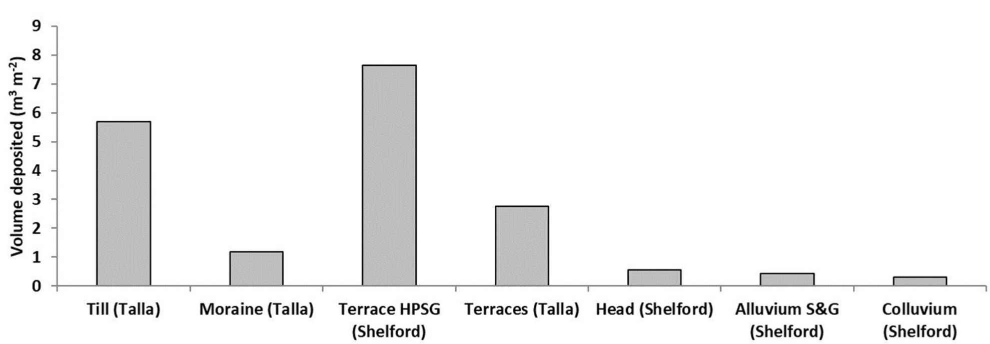

Fig. 3. The estimated thickness of sediment deposited as calculated using high resolution 3D models from a catchment within the UK-LGM (Talla) and one beyond the limits of the UK-LGM (Shelford) for deposits related to the UK-LGM and Holocene. Estimates are determined as one value for all the deposits mapped as a particular type within the 3D model.

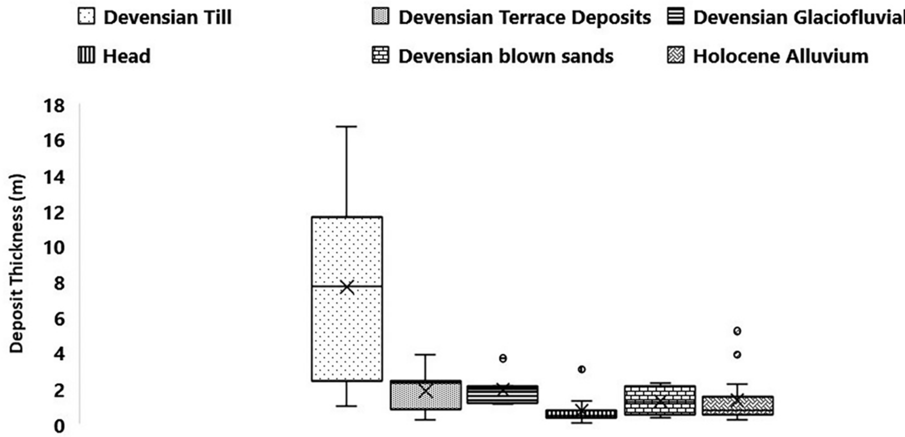

Fig. 4. Depths of erosion deposits collated from a range of BGS 3D models (See Supplementary Information A). The number of models used for data are as follows. For Devensian Till (n = 10 units from models); Devensian River Terrace Deposits (RTD) (n = 11 units from models), Devensian Glaciofluvial (n = 8 units from models); Head (n = 12 units from models); Devensian Blown sands (n = 4 units from models); and Holocene Alluvium (n = 19 units from models).

environments at the time of erosion, and the contribution, dilution and mixing of existing soil organic carbon, along with 'ancient OC' excavated from weathered sedimentary material. Second, Holocene deposits (largely colluvial and alluvial) are likely to be thinner due to greater vegetation coverage ameliorating erosion. However, they are likely to have higher concentrations of OC due to greater NPP, building soil OC.

With regards to the first hypothesis, Table 7 shows concentrations of OC from deposits specifically related to glacial processes (e.g. till, diamicton, and morainic material) and cold climate environments (e.g. terraces, loess, coversand, glacio-fluvial and glacio-lacustrine deposits) which have been reported using OC data collated from examples of UK B/C and C horizons of soil profiles. Typically, these glacially-derived deposits have OC concentrations in their C horizons <0.2%. Any input or leakage of Holocene OC into these SPM deposits will largely depend on later soil development processes based on ecosystem type, landscape position, deposit texture, and climate. Glacial deposits also create paleosols due to deposition over existing soils or sediments. The Torrie silts in Scotland are ~3 m thick organic-rich lacustrine silts that were buried by Younger Dryas glacial till (Merritt et al., 1990). Knowledge of the variation in OC concentration within head deposits is minimal despite their pervasiveness (e.g. Preece et al. 1993; Harrison et al., 2010), with key controls including the time of formation, magnitude, the climate severity inducing solifluction-type processes, and previous levels of soil development as a source material. (See Table 8.)

The second hypothesis suggests that Holocene deposits are likely to be thinner but have greater OC concentrations. In many cases, the depth of elevated OC concentrations within a soil profile will reflect the rate of sediment deposition. Alluvial soils are recognized as potential long-term OC storage sinks (Ferguson et al., 2020; Mayer et al., 2018; D'Elia et al., 2017). Fig. 5 shows OC concentrations of eight typical GB Holocene alluvial or floodplain soils. Elevated concentrations of OC are found at depths >1 m, but will be dependent on the river system and their location within the floodplain. Many floodplain soils in southern Britain have OC concentration profiles that decline with depth (40 cm) but have concentrations >4% OC at the surface (Walling et al. 2006). Within Holocene floodplain soils, a range of deposits are found with elevated OC such as cut-off channel infills, old channels and other relics of river migration across floodplains. Roddens (sand and silt filled tidal creeks deposits) are also found (Smith et al. 2010). Many of these features are deep, and are often highly visible through aerial photography or lidar imagery (Howard et al., 2008).

In Fig. 6 the concepts discussed in this paper are summarized using OC data collected during the production of the 3D model at Shelford (Tye et al. 2010). Fig. 6a shows OC profiles with depth across a *catena* lithosequence (Fig. 6c), predominantly composed of superficial deposits laid down following the UK-LGM. The Holme Pierrepoint sand and gravel river terraces (MIS 2) underlay both the Late Devensian head, where the slope intersects the river terrace, and the later Holocene alluvium deposits that make up the modern floodplain. The modern Holocene floodplain probably represents a cut off deposit reflecting river migration as it is enriched in % OC (1–2%) to depths of 2 m, rather than the more typical Holocene alluvium % OC profiles shown in Fig. 5. In the MIS 2 river terraces, the SPM OC amounts to approx. 0.1%. In Fig. 6b, the SPM formed in the sedimentary bedrock geology underlying the catena shows low concentrations of 'ancient' OC (<0.1%) found at depth. This likely represents the OC that is likely to be found in glacial deposits following glacial excavation.

Colluvium is a major source of Holocene deposition (Leopold and Völkel, 2007). Slope processes are often local and highly complex, often producing a sequence of SPM and organic carbon burial (e.g. Reid and Thomas, 2006; Chiverrell et al., 2007; Preece, 1993). Local factors such as weather events (e.g. storms), vegetation coverage, slope angle, sediment texture, catchment drainage properties and, more recently, land-use (including change) and cultivation practices, all contribute to the magnitude of transport over given time. Whilst contemporary soil erosion across the UK is widespread, surprisingly few assessments have been made of how erosion (water & cultivation) has influenced OC concentrations and stocks along hillslope transects from ridge to toe slope. Soil thickening on toe-slopes is evident (Evans et al., 2019) but there is still much discussion as to whether toe-slopes act as sinks for OC because of the interactions between physical transport, biological, chemical and physiochemical processes as soil is transported downslope (de Nijs and Cammeraat, 2020; Lugato et al., 2018).

8. Estimates of OC storage

Monte-Carlo simulations were used to estimate, on a 'first-order' basis the mass of Total OC stored in deposits after erosion processes, with the aim being to demonstrate the extent that erosion processes are involved in transferring carbon through the landscape and its burial over time. Masses were estimated for UK-LGM Till, Devensian glacio-fluvial deposits and Holocene alluvium as these were the deposits where sufficient data existed to allow calculations. The following conditions were applied for the Monte-Carlo simulations. For, Devensian Till and Glacio-fluvial deposits we use the 'C' or soil parent material concentration for the whole depth of the deposit. This therefore enables us to calculate on a 'first-order' basis the OC burial for the whole depth of deposit, but not count the OC inputted from later vegetation and microbial inputs. By assuming the 'C' horizon concentration any inputs of more recent OC is limited in the calculation.

Table 7
Information regarding the nature of organic carbon distributions and concentrations in soil parent materials in post UK-LGM and pre-Holocene deposits.

Deposit type	Nature of Buried OC	Spatial variability	Buried carbon Magnitude	UK Soil series/Horizon/Depth/% OC	References
Till	Ancient carbon from soils and rocks eroded by glacial movement with inputs from Holocene soil development	Reasonably consistent & well mixed prior to initiation of Holocene soil formation.	Specific to Till type and sediment geological source. Concentrations of OC in C horizon reasonably well understood through Soil Surveys	*Wilcocks-Hafren-Hiarethog 1 / 60–75cm / 0.8% *Wilcocks-Hafren-Hiarethog 1 / 85–100cm /0.6%	Scheib et al., 2008
Loess and cover sands	Allerod peat layers, palaesols buried by wind-blown deposits	Variable spatially and with depth. No systematic mapping of depth of buried OC has been undertaken	Buried peats will be high in OC. Concentrations of OC in surrounding sand and silt particle size will be low reflecting the low surface areas of these size minerals	Everingham /2Cgf /93–151cm / 0.2% Blackwood /Cu / 70–110 cm / 0.2% Naburn /Cu / 65–93 cm / BD	Data from: Avery, 1990
Glacio-lacustrine	Ancient carbon from physical weathering of rocks and old soil profiles	Reasonably consistent & well mixed	Low OC typical of low NPP environments despite particle size being largely clay size	Foggathorpe / Bck(g) / 130–164cm / 0.3%	Data from: Avery, 1990
River Terraces and river terrace gravels (glacio fluvial drift)	Some Ancient carbon from eroded deposits with fresh carbon from Holocene soil development	Reasonably well mixed through terraces	Low at depth with modern soil development at surface. Concentrations of OC in sub soils reasonably well understood through Soil Surveys.	Quorndon/2Cg/84–107 cm/BD Blackwood/Cu(g)/70–110/0.2% Ollerton/Cg/42–90/0.2% Hornbeam/2Bct(g)/110–170/BD Arrow/2Cgf/90–120cm/BD Hurst/3Cg 0r 3Cgf/0.1%	Data from: Avery, 1990
Head	OC from soils that are eroded via solifluction	Found on most slopes. Whilst mapping undertaken very little systematic survey of concentrations undertaken.	Low OC typical of low NPP environments	Salwick/C/30–60cm <0.4 %**	Tye, 2010.
Pre-Holocene Valley Infill	Includes soils eroded and allerod peat layers	Understanding of formation but no systematic mapping of buried OC has been undertaken	Concentrations typical of soil eroded		Preece & Bridgeland , 1999.

Table 8
Information regarding nature of organic carbon distributions and concentrations in soil parent materials in Holocene erosion deposits.

Deposit type	Nature of Buried OC	Spatial variability	Buried Carbon concentration (%)	References
Holocene Gully erosion	Sediment accumulates at bottom of slopes with soil carbon mixed within and palaeosols created.	Depth of OC enriched layer varies. Extensive in upland Britain.	Little is known. However, accumulation of SPM at slope bottoms will reflect soil organic C eroded and transported downslope.	Chiverell et al., 2007 Harvey (1992)
Holocene Alluvial fans	Soil Carbon buried after deposit formation leaving palaeosols	Extensive in upland Britain	Little is known but likely to reflect soil type prior to burial.	Harvey And Renwick ,1987
Holocene Alluvium	Overlays river terraces. Related to eroded soils within catchment	Spatially reasonably well defined through flood plain mapping and soil survey. Depth of OC enriched layer varies (see Figure 5).	Often enriched compared to parent soils as flood deposits often leave clay type deposits with high surface area for carbon sorption	Walling et al., 2006b
Holocene slope deposits	May represent cycles of erosion based on climate and human impacts. Represent carbon from soil development, and may include buried peat deposits.	Widespread on slopes throughout Britain	Few studies have been undertaken to assess OC enrichment due to slope erosion deposits in Britain	Preece & Bridgland, 1999
	Modern day OC deposited with soil erosion			

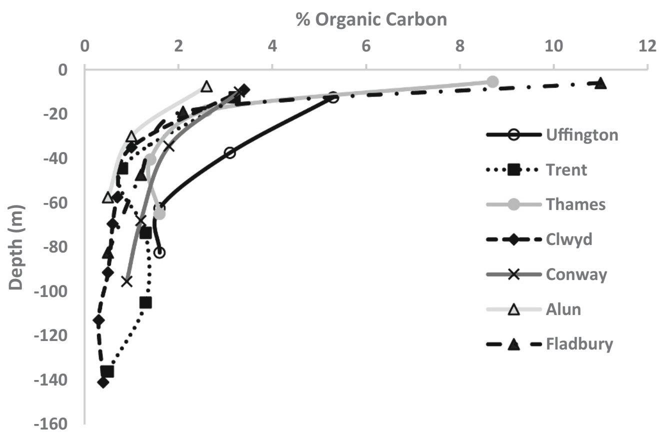

Fig. 5. Organic C concentrations in G.B. Holocene river alluvium. Data collated from Avery, 1990.

Assigning a value for OC concentration stored in the Holocene alluvium was more complicated than for the other deposits. Concentrations of particulate organic carbon in river waters associated with suspended sediment in GB has a median value of ~8% (Worrall et al., 2018). This is comparatively greater than the % OC values reported in alluvial soils (Fig. 7), suggesting that dilution and losses via microbial respiration take place during and after transport. In addition, after deposition, soils will often receive inputs of plant carbon. We therefore assigned a conservative value based on the range at the lowest-most parts of the OC profiles shown in Fig. 7 of between 0.2 and 2%. The range of parameters used to predict OC mass for the three deposits are shown in Table 9.

Results are shown in Fig. 7 and median values for OC transport and storage in post UK-LGM till, alluvium and glacio-fluvial deposits are 328.7, 48.49 and 7.85 MT respectively (a total of 385MT). These values do not include river terraces and those buried deposits resulting from slope erosion processes (e.g. colluvium and gully erosion). However, these values demonstrate the important role of erosion processes in the transport and burial of large amounts of OC over time, of which much will be 'ancient OC'. These values can be placed into context by their comparison with other published figures for OC inventories in soils of the UK, thus demonstrating the magnitude of the erosion and burial process. Part of our estimates would also be included in estimates of SOC when considered to a depth of 1 m, (e.g. they are likely to include the upper part of the 'C' horizon). Bradley et al. (2005) estimated a figure of 4266 MT OC in soils to a depth of 1 m in the UK, and 2370 MT if only the 0–30 cm depth increment was considered. The estimated mass of buried OC from this study, and only three deposit types, therefore represents 9 and 16% of these totals respectively. Rawlins et al. (2011) estimated that the mass of soil inorganic carbon in soils (0–30 cm) of England as 186 MT. Thus, the mass of buried OC reported would represent 206% of this value.

9. Discussion

9.1. Main findings

A 'first-order' understanding of the processes producing the spatial distribution and magnitude of SPM and buried OC since the UK-LGM has been described using an extended C-V-E hypothesis which also assessed the impact of smaller climate variations recorded in the GRIP, along with identified 'flooding events'. SPM deposits with associated OC concentrations demonstrated some correspondence to climate and NPP production, up to the Mesolithic-Neolithic transition. After this, evidence suggests that the 'flooding events' as described by Benito et al. (2015), particularly when combined with human interaction, have been major drivers of erosion and OC burial. The Holocene Bond events, apart

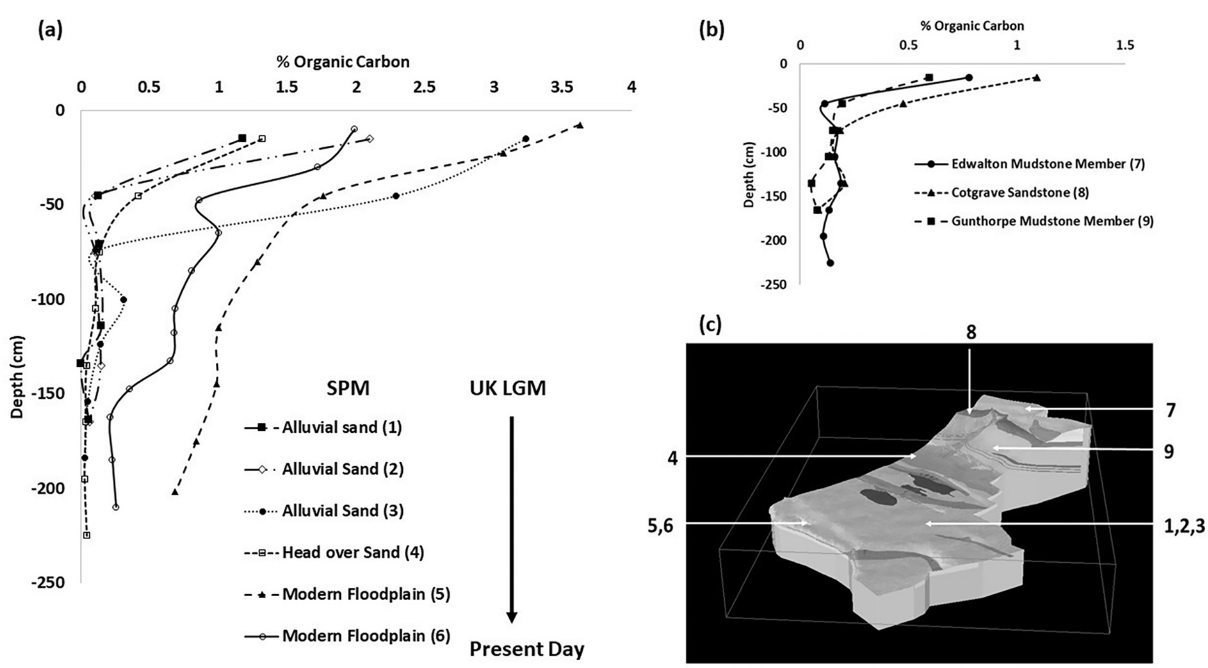

Fig. 6. Summary figure relating to OC concentrations for soils developed into SPM created through post UK-LGM erosion processes. The figure shows SPM of a catena at Shelford close to the river Trent in Nottinghamshire. In Figure (a) SPM with Holocene soil development show the alluvial soils developed into MIS stage 2 river terraces (Holme Pierrepoint Sands and gravels; age ~ 26 K yr BP) where SPM % C is very low; the deposition of Head (and modern Colluvium) over the Holme Pierrepoint Sands and gravels and then later Holocene alluvium deposits of the modern floodplain where % C in enriched to depths of >2 m. Inset (b) shows the % C concentrations of soils and SPM formed in the bedrock geology that underlies the catena, showing low concentrations of OC (probably kerogen like) at depth, this being representative of the OC that would likely be found in glacial deposits such as till if glacially excavated and deposited. The numbers in parenthesis in the legends refer to the lithology of the SPM shown in the 3D model in (c).

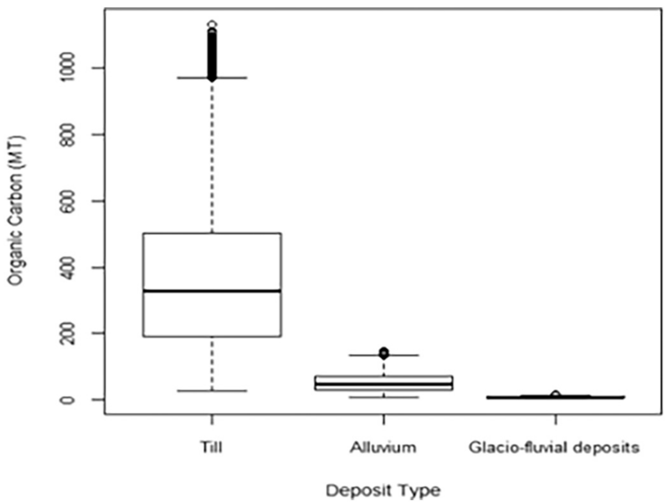

Fig. 7. Boxplot showing output from Monte-Carlo simulations for calculating the potential ranges stored Organic Carbon in post UK-LGM deposits of Till, glacio-fluvial terraces and Holocene Alluvium.

Table 9
Ranges of properties used in the Monte Carlo simulation to determine 'first order' estimates for the quantity of organic carbon transported and buried in selected deposit types in Great Britain since the UK-LGM. Bulk Density and OC data taken from profiles shown in Fig. 5 from Avery (1990) and Tye (2010).

Deposit type	Bulk Density ($g\ cm^3$)	Deposit Thickness (m)	Organic Carbon (kg/kg)	Deposit area (km^2)
Till	1.2–1.6	0.5–6	0.001–0.003	39,833
Devensian Glaciofluvial	1.1–1.6	0.5–6	0.001–0.002	2784
Alluvium	1.1–1.5	0.4–0.9	0.002–0.02	11,126

from the 8.2 k yr event, did not produce a significant erosion signal, although some aeolian redistribution events were identified. However, in parts of eastern Europe, some linkage between Bond events and erosion has been suggested (Dreibrodt et al. 2020 and references therein), but conditions are likely to reflect the more continental locations.

The magnitude of deposit thickness, important as a store for 'ancient OC' and for depth of burial, demonstrated that glacially derived deposits were generally thickest. These deposits were often >1.5 m, the thickness threshold typically accounted for within the UK OC inventory (Bradley et al., 2005). Glacial deposits are generally well mixed and cover large areas, and whilst OC concentrations are low, similar to those found in C horizons (~0.1%), the volume of material, combined with high bulk densities, are likely to contribute to significant OC stocks (e.g. Simo et al., 2019; Harper and Tibbett, 2013). This was demonstrated by the 'first order' estimates produced in Fig. 7. Incorporating measurements of OC concentrations and bulk density for the whole soil profile (i.e., from the soil surface down to the soil-bedrock interface) may more accurately constrain carbon budgets. However, the depth variable deposits with potentially greater OC concentrations that occur as toe slope deposits, valley infills, and alluvial fans, often as part of longer-term cycles of OC burial, are relatively poorly spatially mapped, and their buried OC poorly accounted for.

Additional buried carbon sinks, concomitant with the time-period of this review but beyond its scope, include coastal and estuarine peats formed through burial during GIA and sea level change (Allen and Haslett, 2007; Wheeler and Waller, 1995) and landslides (Boon et al., 2015). Buried OC linked to erosion processes are also found stored in mill ponds, lakes and reservoirs (Butcher et al., 1993; Mendonça et al., 2017). In addition, whilst this review deals with post UK-LGM erosion, the analysis could also be extended back to older UK glaciations including the Anglian. The Anglian was a more extensive and longer glacial event in the UK resulting in greater deposition of glacial till (Clayton, 2000), along with river terrace sequences with buried OC or palaeosols developed during warmer periods (Langford et al., 2007). However, these assessments may only be possible south of the UK-LGM because of later glaciations and would also be complicated by 430,000 years of post-Anglian erosion.

Knowledge of the spatial distribution of OC deposits, along with soil physicochemical properties, would allow for a comparative examination of the biogeochemical and geophysical processes that have contributed to successful long-term sequestration of the OC, thus assessing their suitability for future 'frontier' technologies, and particularly the mechanisms that enhance OC chemical stability. When determining future national OC management strategies, increasing importance needs to be placed on understanding the extent to which buried OC may re-enter the global carbon cycle, to which all buried OC is susceptible if perturbed. Oxidation by microbial processes will impact recent and 'ancient' OC, albeit at different rates. Incorporation of ancient OC weathered from sedimentary rocks during soil formation, and subsequent oxidation processes, can lead to significant loss of ancient or petrogenic OC (Hemingway et al., 2018). Current glacial systems have been identified as a source of CO_2 as microbial communities oxidise petrogenic OC (Horan et al., 2017). However, buried OC oxidation rates are generally limited due to physical and chemical isolation from microbial activity (De Nobilia et al., 2001; Hamer and Marschner, 2005; Ewing et al., 2006; Baisden and Parfitt, 2007; Fontaine et al., 2007). Oxidation rates also reflect soil properties, such as mineralogy and texture, which control drainage and O_2 exclusion, and influence the physical protection of buried OC. Within the context of changing environmental pressures, experimental work has shown (i) deep OC pools are often more responsive to temperature increases (Schwendenmann and Veldkamp, 2006) and (ii) the evidence of the potential for 'priming' old carbon with new carbon. Bernal et al. (2016) demonstrated that substrate additions to mimic plant root exudates (glucose, alanine) and leaf litter could increase the mineralization of OC buried to depths of 2.7 m. This 'priming effect' could complicate 'frontier' technologies, such as the use of deep-rooted plants, to sequester OC deeper in soils. Relatively little work has been undertaken to understand these processes in different lithological and subsoil environments.

Buried OC is likely to be disturbed where land-use is most intensive, leading to perturbations which reduce separation between old and new OC. For example, soils formed within glacially derived SPM are some of the most productive agricultural soils and are subject to regular cultivation. These soils may be favoured for hosting 'frontier technologies' as they are accessible. As populations grow, there is expected to be increased encroachment, engineering and construction onto these soils which will further accelerate their erosion (Price et al., 2011; Barthel et al., 2019). Furthermore, excavation or human disturbance activities may lead to the formation of new hydraulic pathways, allowing fresh carbon (e.g. DOM) to be eluviated deeper into profiles or lost through the land-ocean continuum (Butman et al., 2015). Buried layers such as the Allerød peat or OC in paleosols may be exposed, or brought closer to surface environments, through perturbations such as erosion or excavation, whilst the OC in colluvial and alluvial deposits is concentrated so perturbation may increase reintroduction to the global C cycle.

Changing climates may also add to perturbation of buried OC. Increased rainfall intensity may increase the entrainment and transport of colluvium downslope. Where soils thin on backslopes, increased weathering and release of OC from bedrock may increase as soil formation rates increase as soils thin (Evans et al., 2019). For alluvial soils, bank erosion by rivers can re-introduce both 'recent' and 'ancient' OC (Adams et al., 2015), typically differentiated by measuring ^{14}C and ^{13}C ratios (Blair et al., 2003), back into the C cycle. Whilst, Cole and Caraco (2001) suggest that ancient terrestrial carbon (1000–5000 yrs. old) may be an important source of labile carbon, the extent and mechanisms of aquatic (water and sediment) priming effects are still being debated (Bengtsson et al., 2018; Gintikaki & Witte, 2019).

9.2. Uncertainties and limitations

With respect to OC budgets, this review has revealed the paucity of knowledge regarding buried OC beyond that recorded through national soil surveys, which for large areas is only mapped at 1:250 or 1:50 K and to a maximum depth of 1.5 m. In addition, where buried OC has been found in landscape evolution studies (e.g. paleosols), its position within a sediment profile is known, but there is little information relating to its spatial extent and concentration. This is particularly true where externally transported material has resulted in buried soils, both in cold (e.g. talus and aeolian deposits) and warm (e.g. alluvial fans) climates. Significantly, the lack of information likely results from the perceived unimportance of buried OC during the times of major geological and soil survey in the UK (1873–2000). This reflects the changing priorities of surveys. A further example was the initial priority given to the geological mapping of bedrock, at the expense of thinner superficial deposits in the past (Booth et al., 2015), which may lead to underestimation of the spatial extent of some deposits (e.g. head, colluvium, alluvium). In landscape and geomorphological reconstruction, OC age and its position within profiles has usually been the driver of interest, and not OC concentration. A systematic survey of these deposits is required to understand the extent and concentrations of OC in these deposits, and the controls of NPP, slope and texture.

9.3. Going forward - implications and future research needs

The importance of buried OC is that it exists as a store of C, but with the potential that it may be re-introduced into the active global carbon cycle in the future, as DOC (Butman et al., 2015), POC (Worrall et al., 2016) and if oxidised as CO_2. This presents a significant challenge to scientists tasked with conceiving ways of maintaining or increasing the long-term storage of buried OC. Despite the limitations in understanding OC concentrations and often their highly localized spatial extents, sufficient evidence exists to (i) produce a 'first order' spatial framework, or national model, to identify those different landscapes where buried OC may be found and its type, and (ii) provide evidence that a more detailed terrestrial OC budget for the UK, including the inclusion of deeper buried carbon, should be considered. Initially the framework could adapt the 'Quaternary Provinces and Domains' approach the BGS have developed (Booth et al., 2015). Advances in digital soil mapping techniques and geomorphological analysis may help in identifying buried carbon sinks on slopes by assessing transport pathways and deposition zones. The question of how policy makers best formulate strategies for the management and preservation of buried carbon needs to be addressed, particularly towards alluvial and colluvial soils as both are currently active processes as well as representing historical stores. How drainage and land development may alter water tables depth, essential for controlling access of air to buried OC would be a key question (Yabusaki et al., 2017).

Within this 'first order' framework, considerable knowledge gaps exist. These represent areas where the disciplines of biogeochemistry and geomorphology need to be bought together (Hoffmann et al., 2013). A key focus for further work is understanding in greater detail the net balance of OC resulting from modern day soil erosion and deposition on slopes as this is the dominant current erosion process. Firstly, do transport processes result in OC sequestration on toe-slopes or mineralization and release of CO_2 (Wang et al., 2014a; Wang et al., 2014b), and how do these vary with soil types? Secondly, does increasing the thickness of the soil profile on toe-slopes through erosion increase the amount of OC stabilized within a thickened soil profile over time (Doetterl et al., 2015). Thirdly, losses of soil on backslopes through erosion are particularly significant given that natural soil formation rates have been shown to be slow (Evans et al., 2019; Evans et al., 2021). This is particularly important where erosion is accelerated by modern land use, leaving backslope soil profiles shallower, and with a commensurately smaller OC storage potential. In addition, the thinning of upslope profiles may bring OC, once stored at depth, closer to the surface. However, slope erosion processes may also lead to periods of accelerated soil formation, serving to reverse the trajectory of soil thinning. Within this process, fresh oxide and clay minerals may be released that could help protect future OC from microbial decomposition, but again this may be balanced by an increase in 'ancient' OC released from sedimentary rocks and moving into areas where there is potential for 'priming'. Integrating these processes into global C models for slopes remains a key research objective.

The geomorphological processes described above need to be considered in conjunction with an increased understanding of how the range of erosion processes (e.g. rate of deposition, SPM texture and aggregation), and the time scales they operate over, lead to the successful sequestration of OC through burial. Fundamental research is required to understand preservation processes of buried and ancient OC, including the interactions between OC and mineralogy. Recent work has focused on the nanoscale (Basile-Doelsch et al., 2015) but understanding how aggregation, which affords OC preservation in the topsoil (Oades and A.G., 1991), can be enhanced in deeper soils and the effects of age is another key topic. Interactions of buried OC also need to be considered with respect to other potential technologies that could sequester CO_2 such as the use of basalt mineral weathering when applied to soils (Kelland et al., 2020).

10. Conclusions

This review was based on data associated with the specific climatic, geomorphological and geological conditions of Great Britain. However, the role of C-V-E and the processes resulting in buried OC will be applicable to other countries, but will vary according to their specific conditions. Several broad categories of buried OC have been identified, but generally little work has been undertaken regarding understanding how SPM properties have contributed to the preservation and protection of buried OC through time. Understanding where buried OC exists within landscapes provides a first step towards greater understanding as well as being an archive of material for such studies. SPM exists as a potential sink for OC associated with 'frontier technologies' but understanding how fresh carbon burial may interact with existing buried OC is key. It is also apparent that greater understanding is required of how human perturbation, particularly the ability to excavate large areas and masses of SPM, along with modifying hydrological pathways, impacts the existing buried OC store.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements

The work on this project by AT and JL was funded by the British Geological Survey as part of its NERC funded National Capability program for projects NEE2089 and NEE7185. We publish with the permission of the Executive Director, British Geological Survey (UKRI). The authors would like to thank Martin Hurst for discussions at the beginning of this project.

Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.earscirev.2022.104126.

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