What and where are periglacial landscapes?

Article  (Accepted Version)

Murton, Julian B (2021) What and where are periglacial landscapes? Permafrost and Periglacial Processes. pp. 1-27. ISSN 1045-6740

This version is available from Sussex Research Online: http://sro.sussex.ac.uk/id/eprint/96458/

This document is made available in accordance with publisher policies and may differ from the published version or from the version of record. If you wish to cite this item you are advised to consult the publisher’s version. Please see the URL above for details on accessing the published version.

Copyright and reuse:
Sussex Research Online is a digital repository of the research output of the University.

Copyright and all moral rights to the version of the paper presented here belong to the individual author(s) and/or other copyright owners. To the extent reasonable and practicable, the material made available in SRO has been checked for eligibility before being made available.

Copies of full text items generally can be reproduced, displayed or performed and given to third parties in any format or medium for personal research or study, educational, or not-for-profit purposes without prior permission or charge, provided that the authors, title and full bibliographic details are credited, a hyperlink and/or URL is given for the original metadata page and the content is not changed in any way.
What and where are periglacial landscapes?

Julian B. Murton

Permafrost Laboratory, Department of Geography, University of Sussex, Brighton BN1 9QJ, UK

Abstract

Uncertainties about landscape evolution under cold, non-glacial conditions raise a question fundamental to periglacial geomorphology: what and where are periglacial landscapes? To answer this, with an emphasis on lowland periglacial areas, the present study distinguishes between characteristic and polygenetic periglacial landscapes, and considers how complete is the footprint of periglaciation? Using a conceptual framework of landscape sensitivity and change, the study applies four geological criteria (periglacial persistence, extraglacial regions, ice-rich substrates, and aggradation of sediment and permafrost) through the late Cenozoic to identify permafrost regions in the Northern Hemisphere. In limited areas of unglaciated permafrost regions occur characteristic periglacial landscapes whose morphology is adjusted essentially to present (i.e., Holocene interglacial) process conditions, namely thermokarst landscapes, and mixed periglacial–alluvial and periglacial–deltaic landscapes. More widespread in past and present permafrost regions are polygenetic periglacial landscapes, which inherit ancient landsurfaces on which periglacial landforms are superimposed to varying degrees, presently or previously. Such landscapes comprise relict accumulation plains and aprons, frost-susceptible and non-frost-susceptible terrains, cryopediments and glacial–periglacial landscapes. Periglaciation can produce topographic fingerprints at mesospatial scales ($10^3$–$10^5$ m): (1) relict accumulation plains and aprons form where long-term sedimentation buried landsurfaces; and (2) plateaux with convexo-concave hillslopes and inset with valleys, formed by bedrock brecciation, mass wasting, and stream incision in frost-susceptible terrain.

Keywords: Beringia, geomorphology, ground ice, permafrost, periglacial landscapes, periglaciation
1. Introduction

“The reality is that the nature of landscape evolution under cold non-glacial conditions remains largely neglected in cold region geomorphology.” (French and Thorn 2006, p.171).

“In spite of the fact that ~20% of the Earth’s land surface currently experiences periglacial conditions, field experience suggests that truly periglacial landscapes are rare. This is because the majority of landscapes in the cold non-glacial regions of the world are paraglacial in nature, having only recently emerged from beneath Quaternary and Holocene glaciation.” (French 2016, p.219)

H. M. French (2017a, p.265) highlighted unglaciated landscapes in northern interior Yukon (Canada) as “some of the closest approximations to periglacial landscapes that exist today.” (French and Harry 1992; French 2017a), and his earlier research identified distinctive periglacial landscapes in northwest Banks Island (French 1971) and the chalklands of southern England (French 1972, 1973) (Figure 1). The fourth edition of his textbook (French 2017b, p.12) suggested that “the complete footprint of periglaciation is rarely achieved.” (cf. André 2003) and concluded (French 2017b, p.298) that periglacial landscapes in geomorphic equilibrium comprise (1) cold-climate deltas currently forming at the mouths of northern rivers, (2) unglaciated Pleistocene lowlands in high northern latitudes, and (3) some ice-free areas of Antarctica.

The comments above raise a question fundamental to periglacial geomorphology and our understanding of periglaciation: what and where are periglacial landscapes? The attributes and distribution of periglacial landscapes can be elucidated by placing them in a conceptual framework of landscape sensitivity and change. These concepts reveal that the ability and rates of landscape change vary as a result the differing efficacy of geomorphic processes operating on landforms of differing sensitivity (Brunsden and Thornes 1979; Brunsden 1990, 2001). Some landforms and landscapes—e.g., active silty badlands—are highly sensitive to change and investigated through process or functional studies, whereas others—e.g., ancient erosion surfaces—are insensitive and
studied through historical approaches (Chorley et al. 1984; André 2003). The first type illustrates characteristic forms (cf. Peltier 1950), which embody the proposition that “For any given set of environmental conditions, through the operation of a constant set of processes, there will be tendency over time to produce a set of characteristic landforms.” (Brunsden and Thornes 1979, p. 464). The second type illustrate transient forms, whereby the landforms and landscapes experience perturbations due to changing environmental conditions or structural instabilities that result in transient behaviour of the system over timescales of $10^2$–$10^5$ years or longer.

The concept of constant processes leading to characteristic landforms can be upscaled to characteristic landscapes, where the terrain morphology at micro- and mesospatial scales (see definition of landscape in section 2) is adjusted to the present process conditions. Likewise, the concept of transient forms in varying states of adjustment to changing controls (e.g., climate) can be upscaled to polygenetic landscapes or topography (cf. Peltier 1950), where only partial adjustment of the morphology has occurred, usually in highly localized areas, i.e., at the microform scale (de Boer 1992). Thus, polygenetic topography results where some topographic effects of previous climatic or environmental regimes persist in the present climatic regime (cf. Peltier 1950), and therefore the landscape form expresses a mixture of past and present geomorphic conditions and processes. The past influences are morphologically revealed by landforms that are relict or inactive. In summary, characteristic landscapes express an equilibrium with extant geomorphic conditions and processes, whereas polygenetic landscapes express—in varying degrees and locations—a disequilibrium with them.

The main aim of the present study is to distinguish qualitatively those landscapes considered characteristic of periglaciation from those that are polygenetic, with an emphasis on lowland periglacial areas that evolve relatively quickly in geological terms, over timescales of $10^1$–$10^4$ years. A second aim is to examine the variable impacts of periglaciation: how complete is its footprint? The objectives are (1) to identify in general terms the distribution of fundamental...
periglacial landscapes, based on geological criteria that distinguish permafrost regions; and (2) to outline the main attributes of periglacial landscapes in these regions. The study focuses on empirical observations from the late Cenozoic of the Northern Hemisphere (NH), where studies of periglaciation are best known. Pre-Cenozoic periglaciation and computational models of periglacial landscape evolution (e.g., Egholm et al. 2015) are beyond the scope of the study. Likewise, mountain periglaciation and its interplay with glaciations, where landscape evolution may extend over timescales of $10^5$–$10^6$ years, are not considered in detail. Certain landscapes studied by H. M. French, to whom this article is dedicated, are given particular attention: the chalklands of southern England and the cryopediments and granite landscape of northern Yukon, which he focused on, respectively, during his PhD studies and in the latter part of his career. First, terms are defined before the study is placed in the context of historical attempts to identify periglacial geomorphic regions.

2. Definitions

The term *periglacial* is used hereafter to “describe the climatic conditions, processes, landforms, landscapes, sediments and soil structures associated with cold, nonglacial environments.” (Ballantyne 2018, p.2). The discipline of *periglacial geomorphology*, at its core, highlights the processes driven by ground ice—perennial, seasonal and diurnal—on the initiation and development of landforms and landscapes (cf. Thorn 1992; French and Thorn 2006). Azonal processes relating to snow, wind, liquid water and gravity can exhibit distinctive behaviour and produce distinctive features in periglacial regions and so constitute periglacial geomorphology in a broad sense (cf. French and Thorn 2006; Berthling and Etzelmüller 2011). *Periglaciation* denotes “the collective and cumulative effects of periglacial processes in modifying the landscape” (Ballantyne and Harris 1994, p.3), both ground-ice and azonal processes. A *landscape*, as used by the present author, describes the form of the land surface integrated across the two spatial scales of micro- and mesorelief forms (in the sense used by Karte (1981, 1982). Periglacial microforms
such as patterned ground and gelifluction features typically develop at scales of $10^0$–$10^2$ m, whereas topographic mesoforms such as staircases of cryoplanation terraces, plateaus, escarpments, hills, valleys, large thermokarst basins and river terrace sequences commonly develop at scales of $10^3$–$10^5$ m. A periglacial landscape is therefore defined as an association of periglacial landforms at micro- to mesospatial scales.

Relict is used to refer to landforms and landscapes that developed under climatic and environmental conditions very different for those existing at present. For example, many convexo–concave hillslopes underlain by gelifluction deposits in southern England (section 6.2.2), and northwest France exemplify such relict periglacial terrain because gelifluction is no longer active, and a different process regime now dominates (e.g., soil creep in temperate conditions).

Inactive refers to landforms and landscapes that are stable but remain unchanged under the present climatic and environmental conditions. To illustrate this distinction, many cryoplanation features in Beringia (sections 5.3 and 7.1) may be considered not relict but inactive or weakly active because they currently experience periglacial processes but to a lesser intensity than during full-glacial periods. In other words, they persist in a holding pattern with respect to intense or rapid morphological evolution (F. Nelson, pers. comm. 2020).

3. Historical attempts to identify and classify periglacial geomorphic regions

Historically, several researchers have attempted to identify and classify periglacial geomorphic regions. Tricart and Cailleux (1972, pp. 237–238) used three criteria to identify and subdivide the periglacial morphoclimatic zone: (1) ground frost duration and periodicity (perennial, seasonal frost or diurnal) in the present climate, but including the past climate; (2) vegetation resistance (polar desert/barren grounds and tundra); and (3) total precipitation. From these criteria, five provinces in the periglacial morphoclimatic zone were distinguished: (1) hyperperiglacial (e.g., polar deserts of Antarctica and northern Greenland), (2) mesoperiglacial (e.g., other polar deserts
in North America and Eurasia), (3) tundra, (4) steppe periglacial (e.g., Mongolia and northeast Iceland), and (5) taiga province on residual Pleistocene permafrost (continuous or discontinuous).

More recently, J. Karte (1981, 1982) delineated and subdivided the present arctic and subarctic periglacial zone on the basis of the spatial distribution of active periglacial microrelief features and their climatic threshold values. The microforms, particularly gelifluction landforms, express the active geomorphic effects of frozen ground on diurnal, seasonal or perennial timescales.

Accordingly, regional types of the arctic and subarctic periglacial zone comprise subarctic maritime, subarctic continental, boreal, arctic tundra, arctic frost-debris and high arctic frost-debris zones.

In Russia, the morphological structure of some landscapes in the Arctic zone has been distinguished from the tundra zone based on criteria of (1) sparse vegetation cover, (2) absence of evidence for gleying in soil profiles, and (3) highly localized complexity in soil–vegetation cover (Mikhailov 1981). For example, on Bolshevik Island (part of the Severnaya Zemlya Archipelago) and Faddeev Island (in the New Siberian Islands) (Figure 1) the topography of the Arctic zone comprises regular complexes of different types of landscape, which are divided into complexes of ‘facies’, and these in turn are divided into ‘nanofacies’. The nanofacies constitute “Segments of land surface that have a uniform soil–vegetation mantle and an area varying from several square centimeters to one square meter and that alternate regularly within the boundaries of the facies...” (Mikhailov 1981, p.156). Both the nanofacies and facies comprise a continuum from abiogenic (non-vegetated) to biogenic (vegetated) types of land surface.

Overall, the classifications of Karte (1981, 1982) and Mikhailov (1981) identified periglacial zones characterized by microforms, nanofacies and facies adjusted to present-day climatic and environmental conditions. Though the morphoclimatic map of Tricart and Cailleux (1972) distinguishes the approximate extent of relict Pleistocene permafrost, the map is really one of
landform controls (Stoddart 1969). Fundamentally, therefore, none of the maps or classifications reveal the wider geomorphic impact of periglaciation in modifying the landscape at the mesospatial scale. In other words, they exemplify a functional approach to geomorphology and geoeecology at the expense of an evolutionary or historical approach. An exception is the morphoclimatic map of J. Büdel (1982).

Büdel (1982) developed a conceptual model of frost shattering of bedrock in the upper metres of permafrost, to produce an ice-rich layer (the ‘ice rind’ or, in modern parlance, the ‘transition zone’; Shur et al. 2005). The ice rind preconditioned the ground for intensive thermal erosion—both vertically and laterally—by water flowing along river channels. These processes, Büdel concluded, produced a subpolar zone of excessive valley cutting that extended across (glacier) ice-free areas northern Alaska, much of northern Canada, ice-marginal areas of Greenland, Svalbard and the far north of Eurasia. Additionally, a taiga valley-cutting zone consisted of boreal regions North America and Eurasia underlain by permafrost. Extrapolation of valley cutting from Büdel’s field observations of the ice rind in southeast Svalbard to the wider Arctic and Subarctic (Figure 1), however, has been problematic, because at some localities within permafrost regions such as western Svalbard, west and southeast Greenland deposition of colluvium dominates over erosion in valley bottoms, and in some glaciated permafrost terrains, river activity is controlled by glacial landforms and sediments (Karte 1981) and thus is largely paraglacial in nature. Other problems with the valley-cutting model are (1) the seasonal timing of maximum river discharge in many arctic and subarctic streams occurs months before the active layer has reached is maximum, late-summer depth; and (2) deep rivers (i.e., where water depth exceeds the maximum thickness of river ice, ~2 m) have a talik beneath them, precluding the occurrence of a layer of near-surface ice-rich permafrost (Karte 1981). Finally, accurate dating of landscape evolution over timescales in which rivers incise and planate significantly remains difficult at present. Thus, it is timely to re-evaluate the criteria to identify periglacial landscapes over timescales that can be assessed
reasonably from geological evidence. This is most easily examined in landscapes that evolve relatively quickly as a result of the geomorphic processes of ice segregation and thermokarst.

4. Criteria

Four empirical geological criteria serve to distinguish landscapes that express varying degrees of periglaciation: (1) persistent or recurrent periglacial conditions, (2) extraglacial regions, (3) ice-rich substrates, and (4) aggradation of sediment and permafrost.

4.1 Persistent or recurrent periglacial conditions

The first criterion requires that periglacial conditions have persisted over glacial–interglacial timescales or recurred multiple times, maximising the duration of periglacial processes to produce a “uniform assemblage of landforms” (Brunsden 1993, p.25). First, the Cenozoic history of NH periglacial and permafrost conditions is briefly summarised in order to provide context for presenting a rationale for mapping fundamental periglacial regions.

4.1.1 Cenozoic history of Northern Hemisphere (NH) periglacial and permafrost conditions

Since the Early Eocene Climatic Optimum (~52–50 Ma), global temperatures have generally declined (Zachos et al. 2001), leading to recurrent ice ages of the Quaternary Period. In total, 103 marine isotope stages (MISs) have been identified for the 2.58 Ma of the Quaternary (Head 2019), of which 51 were cold stages (even-numbered MISs) (Figure 2). The terrestrial record of cold stages, by contrast, is limited by erosion and non-deposition, with far fewer cold stages identified so far: 13 since ~3 Ma in the case of the North American environs of the Arctic Ocean (Harris 2005).

Periglaciation in mid- to high-latitudes and at high altitudes of the NH became an important driver of landscape evolution during the late Pliocene to Early Pleistocene. For example, periglacial processes began to exert a dominant effect on landscape change and production of coarse-grained
sediments in susceptible regions of Britain from the beginning of the Quaternary (Rose 2010). The role of periglacial processes in landscape evolution probably increased after the Mid-Pleistocene Transition (~1.2–0.8 Ma), when the dominant orbital cycles changed from ~41 ka (obliquity) to ~100 ka (eccentricity) (Elderfield et al. 2012). The 100 ka cycles resulted in prolonged glacial stages associated with expansion of NH ice sheets and southwards extension of permafrost and periglacial environments. A further increase in the amplitude of 100 ka climate cycles occurred at ~450 ka (after MIS 13), a transition known as the mid-Brunhes Event (Candy et al. 2011).

Permafrost in the NH developed as early as the late Pliocene or Early Pleistocene, and has persisted since the Middle Pleistocene in some arctic and subarctic regions (Figure 2). Past permafrost—i.e., permafrost that no longer exists at a particular locality (French 2008)—is inferred to have developed as early as the late Pliocene, ~3 Ma, based on ice-wedge pseudomorphs from the Klondike, Yukon, Canada (Westgate and Froese 2001) (Figures 1 and 2). The onset of permafrost during the late Pliocene has also been inferred indirectly from pollen spectra consistent with a change in vegetation toward tundra and cold-adapted larch–birch forest from samples in and near Lake El’gygytgyn in Chukotka, northeastern Russia (Glushkova and Smirnov 2007; Andreev et al. 2013; Brigham-Grette et al. 2013) and from a reduction of Ca\(^{2+}\) flux into the lake, attributed to reduced chemical weathering in the catchment, after 3.3 Ma, during the M2 cooling event (Wennrich et al. 2014). The oldest known evidence for past permafrost in Alaska is based on ice-wedge pseudomorphs in central Alaska dating to ~2 Ma or shortly before, i.e., Early Pleistocene (Matheus et al. 2003; Péwé et al. 2009). Past permafrost may have developed on the highest summits of the Qinghai–Tibet Plateau (QTP), China, during the Early Pleistocene, although no direct evidence for it is reported prior to MIS 6 (~150–130 ka) (Chang et al. 2017). According to these authors, past permafrost likely developed on some highlands of the QTP around the margins of glaciers sometime around 780–560 ka (e.g., MIS 16), in the early Middle Pleistocene, and again around 480–420 (MIS 12), thawing out during warmer interglacial conditions after both periods.
Ancient permafrost—i.e., permafrost that has persisted at a locality since the Pleistocene or earlier—containing ground ice as old as \(~740\) ka has been reported from the Klondike, Eastern Beringia (Froese et al. 2018), and \(~650\) ka from Batagay megaslump, Western Beringia (Murton et al. in review) (Figures 1 and 2). Such persistence is consistent with the absence of observed speleothem growth—as permafrost freezes the karst vadose system in caves—since \(~429\) ka (MIS 11) in a Siberian cave (Ledyanaya Lenskaya) near the boundary between continuous and discontinuous permafrost (see Vaks et al. 2013, 2020). More generally, it is highly probable that ancient permafrost persists in some mountain regions, where it has survived beneath or adjacent to cold-based glaciers (B. Etzelmüller, pers. comm. August 2020).

In summary, the evidence above suggests that permafrost may have thawed and re-formed multiple times during the late Pliocene and Early Pleistocene and became persistent, at least in certain arctic regions, at some time in the Early Pleistocene. Thus, persistent permafrost conditions can reasonably be inferred for unglaciated arctic and subarctic continental regions such as the Klondike in northwest Canada and Batagay in northeast Siberia since \(~750–650\) ka (early Middle Pleistocene), and earlier phases of past permafrost—at least during cold MISs, and possibly some warm MISs—are likely in such regions since \(~3.2–2.0\) Ma (Figure 2).

4.1.2 Mapping periglacial environments

To map the spatial limits of present and past periglacial environments, a conservative approach is adopted, based on the spatial extent of permafrost. According to T. L. Pévé (1969, p. 4), “… permafrost is the common denominator of the periglacial environment, and is practically ubiquitous in the active periglacial zone.”, whereas H.M. French’s (2017b, p.4) fourth edition of his textbook noted that “… all periglacial environments experience either seasonally-frozen or perennially-frozen ground.” The present study excludes seasonally frozen ground in non-permafrost areas because identifying permafrost areas is more tractable for mapping purposes. This approach thus represents a first step towards identifying periglacial environments, and it is
acknowledged that a more complete mapping of periglacial environments should include areas of seasonally frozen ground.

The approach used in the present study allows distinction of permafrost regions, which can be considered the fundamental areas in which periglacial landscapes are most likely to exist. A map of present-day permafrost extent in the NH is shown in Figure 1 (Brown et al. 2002). The map divides permafrost into zones delineated by the spatial extent of permafrost. The zones vary—along a transect from colder northern areas to warmer southern areas—from continuous permafrost (where permafrost underlies 90%–100% of the land surface), through discontinuous permafrost (50%–90%), to sporadic permafrost (10%–50%) or isolated permafrost (0%–10%).

Knowledge of the extent of permafrost before MIS 2 is rudimentary, and so the maximum extent of past permafrost reconstructed for the Last Glacial Maximum (LGM), ~26.5–19 ka (Clark et al. 2009), is adopted here. This indicates to a first approximation the maximum extent of past permafrost during the main cold MISs of the Middle and Late Pleistocene. The extent of LGM permafrost in the NH is shown by the pink area in Figure 3, which is based on the most recent synthesis of empirical evidence (Lindgren et al. 2016, figure 1). The LGM permafrost region encompasses continuous, discontinuous, sporadic and isolated permafrost as defined above, because the individual extents of each zone before the present day are highly uncertain (Lindgren et al. 2016). Thus, the LGM permafrost is mapped very broadly and includes numerous areas lacking permafrost. Together, Figures 1 and 3 approximate the overall region of recurrent and persistent permafrost associated with the 100 ka eccentricity cycles characteristic of the Middle and Late Pleistocene.

4.2 Extraglacial regions

The second criterion requires that periglacial conditions existed beyond the maximum limits of glaciation. Definition of the extraglacial regions maximises the area for subaerial cold-climate conditions free of glacial erosion and deposition. This criterion therefore excludes the direct
geomorphic effects from glacier ice on periglacial terrains, though it does not exclude the indirect
g geomorphic effects of glacial processes acting in the proglacial zone, for example on the
development of ground ice supplied from pressurised groundwater (e.g., Murton 2005) or on
surface meltwater activity (Benn and Evans 2010). Equally, the criterion does not exclude the
preservation of preglacial landforms (e.g., tors, blockfields and patterned ground) and weathering
profiles beneath cold-based glacial ice (Waller et al. 2012), as a result of protection within
subglacial permafrost.

The maximum limits of glaciation by NH glaciers and ice sheets are late Cenozoic (i.e., Pliocene–
Pleistocene) in age. The onset of widespread glaciation recorded in the NH by terrestrial evidence
was around 2.5–2.4 Ma (Ehlers et al. 2011; Joyce et al. 1993; Balco and Rovey 2010), and that
recorded by ice-rafted debris in ocean cores, around 2.6–2.7 Ma (Haug et al. 2005; Bailey et al.
2013)(Figure 2). Permafrost probably existed beyond the margins of some these ancient Plio–
Pleistocene ice sheets, though evidence reported for it is limited (Harris 2005).

The maximum limits of Plio–Pleistocene glaciation by the major NH ice sheets are shown in
Figure 3. This map is a first attempt to depict the outermost glacial limits obtained from the LGM
(Hughes et al. 2016) and the best-estimates for 17 pre-LGM time-slices compiled by Batchelor et
al. (2019) for the periods between the late Gauss palaeo-magnetic Chron (3.6–2.6 Ma) and 30 ka.
As a result, the areas of present-day permafrost or LGM permafrost beyond the outermost glacial
limits can be taken to represent periglacial landscapes unglaciated during the late Cenozoic.

4.3 Ice-rich substrates
The third criterion requires the substrate on which periglacial processes operate to be sensitive
to change, enabling characteristic landforms (Brunsden and Thornes 1979) to develop rapidly and
resurface landscapes. Such sensitivity in periglacial regions arises where the upper metres (near-
surface) of materials are highly erodible, either because the substrate is rich in ground ice (i.e.,
contains excess ice) and/or in unconsolidated silt or sand. Excess ice in the upper metres of
permafrost tends to be vulnerable to melting and landscape disturbance by thermokarst activity, producing a range of characteristic thermokarst landforms (Jorgenson 2013; French 2017b). Excess ice is particularly common in silt-rich sediments with an abundant moisture supply (Morse et al. 2009; O’Neill and Burn 2012; O’Neill et al. 2019), as silt is highly high frost susceptible. Excess ice also occurs in frost-susceptible bedrock such as marly limestone, arkose (Büdel 1982) and shale (French et al. 1986). The focus on ground ice highlights its fundamental role in periglacial geomorphology in terms of “landform and landscape initiation and development” (Thorn 1992, p.1). It also echoes Büdel’s (1982) concept of the ice-rind (i.e., transition zone; Shur et al. 2005) as important to periglacial landscape evolution. To show regions rich in ground ice, the *Circum-Arctic Map of Permafrost and Ground-Ice Conditions, Version 2* (Brown et al. 2002) is used, as this distinguishes regions of high (>20%) volumetric ground-ice content in the upper 10–20 m of ground. Figure 4 exemplifies such regions (in red) in unglaciated terrains.

In regions of past permafrost—where excess ice is no longer present—the former presence of excess ice can be inferred from the frost susceptibility of the substrate as well as cryostratigraphic evidence for former ground ice (Murton 2013). In mid-latitude regions such as northwest Europe, the former occurrence of excess ice is commonly indicated by near-surface brecciation that represents relict weathering profiles developed in rocks such as chalk, slate, shale and silty sandstone in France (Cailleux 1943), Germany (Büdel 1982, p.84) and Britain (Murton and Ballantyne 2017).

### 4.4 Aggradation of sediment and permafrost

The fourth criterion focuses on aggradation of sediment and permafrost, which together create new lowland periglacial landsurfaces; therefore, the effects of landscape inheritance (Brunsden 1993) are minimised or avoided. Resurfacing of periglacial landscapes by sedimentation operating at timescales of $10^3$ years was widespread during cold stages of the Pleistocene. Silt and silt–sand intergrades—deposited primarily by wind (loess) and sometimes partially reworked—blanketed
huge areas of unglaciated permafrost terrains in Beringia and past permafrost terrains in southern Siberia, central and northwest Europe (Murton et al. 2015). In Beringia, the silt is characteristically ice-rich, and forms regionally widespread cryostratigraphic units known as ice complexes, of which the most recent, Late Pleistocene, one is termed yedoma. Hence, the yedoma is included with the high volumetric ground-ice category shown in Figure 4. Such sedimentation resulted in long-term growth of syngenetic permafrost. Other types of sedimentation in modern permafrost settings—notably river floodplains, deltas and alluvial fans—can result in growth of syngenetic, quasi-syngenetic or epigenetic permafrost, depending on continuity of deposition of clastic sediment, accumulation of organic material on the ground surface or the timing of channel switching (avulsion). Criterion 4 applies primarily to periglacial lowlands. In periglacial uplands, however, erosional rather than depositional processes may be more important agents of periglaciation, for example by modifying bedrock and leading to formation of erosional periglacial landforms such as cryoplanation terraces and staircases of terraces (section 7.5).

5. Permafrost regions

From these criteria, four types of permafrost region are identified based on glaciation and permafrost status: (1) glaciated permafrost, (2) glaciated past permafrost, (3) unglaciated permafrost, and (4) unglaciated past permafrost.

5.1 Glaciated permafrost regions

Glaciated permafrost regions are those that meet criterion 1 (persistence) to some extent, and sometimes meet criterion 3 (ice-rich). In such regions, glaciers or ice sheets previously advanced over lowlands and uplands where permafrost persists at present. Permafrost was commonly present before the glacial advance and, at least beneath the ice-marginal areas, commonly persisted beneath cold-based ice, with the result that glacier–permafrost interactions were common (Waller et al. 2012). In lowlands, regional burial of basal ice and glaciotectonic deformation of permafrost occurred in western Siberia (overridden by the Barents–Kara Ice Sheet;
Astakhov et al. 1996), northwest Canada (overridden by the Laurentide and Innuitian ice sheets; Murton et al. 2004) and the northeastern New Siberian islands (overridden by the Eastern Siberian Ice Sheet; Tumskoy 2012) (Figure 3). Glacial deposits, ground ice and ice-cored moraines are common in such areas, and the landscape is in a state of delayed or arrested deglaciation (Astakhov and Isayeva 1988; Dyke and Evans 2009), though recent climate warming is renewing deglaciation and driving postglacial permafrost landscape evolution in northwest Canada by rapid thermokarst activity (Kokelj et al. 2017). In mountains and uplands, cold-based glacial ice is thought to have had limited to no geomorphic impact on preglacial landforms, resulting in preservation of features such as weathering zones, trimlines and tors (Waller et al. 2012), for example in the Torngat Mountains of northern Labrador (Evans 2016), where permafrost today is continuous to extensive discontinuous (Way and Lewkowicz 2016) (Figure 1).

5.2 Glaciated past permafrost regions

Glaciated past permafrost regions commonly meet criterion 1 (persistence) and locally meet criteria 3 (ice-rich) and 4 (aggradation). In these regions, glaciers or ice sheets advanced over terrain that presently lacks permafrost, though in many mid-latitude regions past permafrost occurred there during times of glacial advance or retreat. Such regions include glaciated parts of northern Europe, northern USA, and southern and western Canada (Ehlers et al. 2011; Figs. 1 and 3). In these regions, glacial deposits, moraines and glaciotectonic deformation are common, as is evidence of past permafrost (e.g., ice-wedge pseudomorphs; e.g., Rose et al. 1985) and periglaciation (e.g., involutions, patterned ground, vertical stones; Ballantyne and Harris 1994). On hillslopes, the upper metres of tills have commonly been reworked by gelifluction. Signs of past periglacial activity tend to be more muted on glacially eroded bedrock terrains such as the southern parts of the Canadian Shield, where the rock types tend to be non-frost-susceptible. In some regions, accumulation of aeolian sand, silty sand and loess blanketed glaciated past permafrost terrains in NW Europe (Bertran et al. 2016; Andrieux et al. 2018).
5.3 Unglaciated permafrost regions

Unglaciated permafrost regions meet criteria 1 (persistence), 2 (extraglacial) and 3 (ice-rich) everywhere, and commonly meet criterion 4 (aggradation). Such regions are dominated spatially by Beringia, the unglaciated Pleistocene subcontinent that stretched from the Verkhoyansk Mountains of northeast Siberia eastward across the emergent Bering land bridge through central and northern Alaska to the Klondike region of Yukon Territory, Canada (Hopkins et al. 1982; Hoffecker and Elias 2007) (Figure 4). Also included within Beringia were continental shelves to the north of the present coastlands, for example the East Siberian Arctic Shelf, which formed the northwest Beringian Plain (Hoffecker et al. 2020). Bodies of glacier ice in the northern areas of Alaska and central and eastern Siberia (between ~136°W and 123°E) were generally restricted to interior uplands and never reached the outer margins of the subaerially exposed continental shelves underlain by LGM permafrost (Figure 4). Glaciation in Beringia was particularly restricted during the LGM compared to some earlier time-slices (e.g., MIS 6, 5 and 4; see Batchelor et al. 2019). During the LGM, Beringian environments, especially to the east and west of the Bering land bridge (central Beringia), were characterised by intense aeolian activity and associated windblown sand and silt deposits (Hopkins 1982). They formed part of a vast, dry and dusty Late Pleistocene permafrost zone extending from northwest Europe across northern Asia to northwest North America (Murton et al. 2015).

The map presented in Figure 4 shows that Beringia included not only lowlands (Elias and Brigham-Grette 2013) but some interior uplands—such as those of the Yana—distant from the moderating effects of coastal climates throughout the Quaternary and therefore persistently continental (i.e., large annual temperature range, little precipitation and relatively warm growing season). In much of Beringia, permafrost persisted even during interglacial conditions—though locally it may have disappeared on warmer south-facing slopes in the discontinuous permafrost zone—and was re-invigorated during glacial conditions. The map in Figure 4 also identifies the
unglaciated permafrost region to the west of Beringia as the unglaciated Siberian Platform, which is usually excluded from Beringia (S. Elias and A. Fedorov, pers. comm. October 2020). The unglaciated Siberian Platform includes ice-rich permafrost in the broad ice-free corridor between the western margin of the Verkhoyansk Mountains and the eastern margin of the Central Siberian Plateau. Within it are the river valleys of the Lena, Vilyuy and the lower course of the Anabar.

5.4 Unglaciated past permafrost regions

Unglaciated past permafrost regions meet criteria 1 (persistence) and 2 (extraglacial), and in places meet criteria 3 (ice-rich) and 4 (aggradation). Such regions tend to occur in the mid-latitudes to the south of the Laurentide and Cordilleran ice sheets in North America, and to the south the Fennoscandian, Barents–Kara and British–Irish ice sheets in Eurasia (Ehlers et al. 2011; Figure 3). In the absence of glacial landforms and landscapes, inheritance of non-glacial features such as Neogene or Palaeogene erosion surfaces tends to be more evident (e.g., Mignoń and Goudie 2012), determining the initial conditions on which periglackation has operated.

6 Periglacial landscapes

The permafrost regions outlined above permit distinction of fundamental periglacial landscapes that are characteristic or polygenetic.

6.1 Characteristic periglacial landscapes

Characteristic periglacial landscapes in lowland regions comprise assemblages of landforms whose morphology is adjusted essentially to present (i.e., Holocene interglacial) process conditions. The stipulation of landscape adjustment to present conditions is necessary to delineate characteristic periglacial landscapes in a way that is tractable, given present knowledge about rates of landscape evolution. Such knowledge is greater for periglacial lowlands than for periglacial uplands. In lowlands, multiple dating methods have been applied to cryostratigraphic sequences and landforms, and landscapes evolve more rapidly by growth and melt of ground ice, and by erosion and deposition of sediment (Murton 2001, 2009, 2013). In uplands, erosion of bedrock
tends to be slower, and dating of it appears to be more difficult and is at an earlier stage of
methodological development, as discussed below in the context of cryoplanation terraces (section
7.1). Thus, it is currently difficult to establish beyond reasonable doubt if upland periglacial
landscapes are characteristic of their present geomorphic regimes.

Two types of lowland characteristic periglacial landscapes are identified: (1) thermokarst
landscapes and (2) mixed periglacial landscapes. Thermokarst landscapes comprise landforms
produced by present ground-ice-related (freeze–thaw) processes, and so represent characteristic
periglacial landscapes in a narrow sense. Mixed periglacial landscapes comprise a mixture of
landforms produced by interactions between present ground-ice-related processes and azonal
processes; these represent characteristic periglacial landscapes in a broad sense. The main
lowland regions likely to contain characteristic periglacial landscapes are approximated by the red
areas on Figure 4. The regions outlined are not exhaustive and omit some areas of known ice-rich
permafrost (e.g., in interior Alaska). They simply provide a first attempt to delineate a
geographical framework to narrow the search for characteristic periglacial landscapes.

6.1.1. Thermokarst landscapes
Thermokarst landscapes occur where all four criteria are met. Aggradation of sediment and
accompanying growth of syngenetic permafrost buried pre-existing topography beneath blankets
of silty deposits up to 10s of metres thick and rich in ground ice, thus producing extensive
accumulation surfaces underlain by ice complexes (section 6.2.1). Such environmental conditions
ceased at the end of the Pleistocene, leaving the accumulation surfaces (inter- alas areas) that
persist at the present-day as inherited features. During the Holocene, assemblages of thermokarst
landforms have developed within the accumulation surfaces and are characteristic of present,
interglacial conditions. Thermokarst was driven by natural geomorphic processes during much of
the Holocene, and increasingly by human disturbances in the last century or so. Numerous modes
of permafrost degradation have been distinguished from boreal regions of Alaska (Jorgenson and
Oskerkamp 2005) and thermokarst terrains more generally (Jorgenson 2013). Thermokarst landscapes are illustrated by way of alas terrains and icy badlands.

6.1.1. Alas terrains

Alases are large flat-floored depressions with steep sides that form by thaw settlement of very ice-rich permafrost (Soloviev 1962; Czudek and Demek 1970). In the central Yakutian lowland, Siberia (Figure 4), they vary in area from ~0.5 to >100 km², and in depth from ~3–40 m. Such thermokarst depressions can host a variety of smaller permafrost landforms, including ice-wedge polygons, thermokarst mounds, pingos, and retrogressive thaw slumps (Figure 5). Where alas coalesce, they form thermokarst valleys, which are characterised by unexpected turns, blind spurs, wide sections (alas basins) and—during early stages of development—by a step-like longitudinal profile. As thermokarst valleys develop, their longitudinal profiles become more graded as a result of infilling of the deepest alases with sediment and erosion of the higher alases along thermokarst gullies. The late stage of valley development is characterised by a wide grassy bottom that is flat to gently sloping with basin-like segments. As Czudek and Demek (1970, p.119) concluded, thermokarst “destroys the initial surface of large areas and creates a new lower level of the lowlands, which is initially independent of the main level of erosion. The thermokarst process thus represents a special type of relief development.” Such special relief, it is argued here, represents a truly characteristic periglacial landscape.

Evolution of alas terrains has been elucidated by integrating field observations from regions of yedoma in Russia and Alaska, and revising the conceptual model of thermokarst-lake development. The model originally developed by Soloviev (1962) and presented in English by Czudek and Demek (1970) was revised by Shur et al. (2012). In outline, the revised model has six stages: (1) the active layer deepens and thaw settlement commences; (2) water ponds in ice-wedge troughs; (3) thermokarst ponds form above ice-wedge polygons; (4) thermokarst lakes deepen rapidly, (5) yedoma thaws completely beneath lakes, and shorelines experience thermal
erosion; and (6) lakes drain and ground freezes in drained-lake basins (alases), alas valleys and alas plains. All of these landforms are common in the northern part of the Seward Peninsula, Alaska (Figure 4). Importantly, the early stages of thermokarst (1 and 2) are commonly interrupted by drainage or accumulation of organic matter, which help to stabilize the yedoma landscape and provide a negative feedback to thermokarst development. Thus, at the present day, development of thermokarst lakes (stage 4) is unusual, and yedoma in Alaska and northern Yakutia is degrading mainly by thermal erosion along the shores of rivers and seas (Shur et al. 2012).

A specific example of an alas plain landscape occurs in the Innoko and Koyukuk Flats, a region of lacustrine–loess lowlands in discontinuous permafrost of interior Alaska (Figure 4). Here, cryostratigraphic observations of the upper metres of permafrost deposits, supplemented by radiocarbon ages, support a conceptual model of terrain evolution from the Late Pleistocene to the present-day (Kanevskiy et al. 2014). Starting with initial conditions of a Late Pleistocene accumulation surface underlain by ice-rich silt (yedoma), the model has four stages of yedoma degradation followed by five stages of permafrost aggradation and degradation that produced a complex mosaic of terrain, hydrological and permafrost conditions. In essence, the present terrain is considered to represent “large thaw lake (alas) plains” in which “ground ice dynamics are fundamental to the evolution of the broader landscape.” (Kanevskiy et al. 2014, p.31).

The degree, extent and rates of landscape evolution of alas terrains have started to be quantified based on terrain analysis at regional to local spatial scales. For example, in part of the Kolyma Lowland of the northern Sakha Republic (Yakutia) (Figure 4), thermokarst activity during the Holocene thermal maximum transformed ~50% of the initial yedoma surface underlain by ice-complex deposits into alas terrain, with a further 11% of the yedoma terrain transformed into alases in the late Holocene (Veremeeva and Gubin 2009). Some 80–100% of the area of low-lying northern coastal regions of the yedoma–alas complex of the Kolyma Lowland is dominated by thermokarst basins (Nitze et al. 2017). These authors detected a net loss of 0.51% of lake area
regionally in the lowland between the years 1999 and 2014, though different sub-regions showed different trends, with lake expansion—attributed to thermokarst and flooding—dominant in the northern part of the yedoma–alas complex. Nitze et al. (2017) have also detected recent changes in lake area on the Alaskan North Slope and in the central Sakha Republic, some of which they attribute to thermokarst activity. At the local spatial scale, rapid subsidence at average rates of 2.1 and 3.9 cm yr$^{-1}$ of ground above the melting tops of ice wedges at two sites in the Churapcha area near Yakutsk (Figure 4) has been recorded from high-definition topographic data (Saito et al. 2018). The thermokarst development started in the early 1990s and is visually expressed by growing thermokarst mounds.

In summary, alas terrain represents an expanding characteristic periglacial landscape. Collectively, the alas terrain identified in the new permafrost–landscape map of the Sakha Republic (Fedorov et al. 2018) comprises 3.6% (~112,000 km$^2$) of the total area of the republic.

6.1.1.2. Icy badlands

Icy badlands constitute a distinctive landscape of gullies and interfluv es formed primarily by rapid thermal erosion through ice-rich permafrost (Murton 2009). Such badlands are particularly well developed and extensive in the floor the Batagay Megaslump, the largest known retrogressive thaw slump on Earth, in the Yana Uplands of the northern Sakha Republic, Siberia (Figure 4; Murton et al. 2017). Thermal erosion on the slump floor has formed v-shaped gullies between sharp-crested interfluv es (Figure 6). Active gully sides commonly expose bare sand, often at or near the angle of repose of the sand, due to active mass movement. Accordant summits of the interfluv es as well as baydzherakhs (thermokarst mounds) on top the interfluv es initially mark the base of the upper ice complex, though erosion lowers them through time. The original hillslope gully now forms the primary gully that runs downslope along the middle of the slump floor towards the Batagay River. Tributary gullies feed into the main gully in a dendritic pattern. In summer, thermo-erosional niches commonly develop beside active streams in gully floors.
The badlands of the Batagay Megaslump developed after human-induced disturbance to the taiga vegetation cover in the 1940s to 1960s initiated a gully on the hillslope in the early 1960s (Savvinov et al. 2018). The gully initiated thaw slumping along its central part during the 1980s, with the slump enlarging to megaslump (>0.2 km²) proportions during the 1990s. Between 1991 and 2018, the area of the slump floor increased from 0.19 to 0.78 km², and the maximum depth of the floor below the adjacent terrain surface was ~70 m (Vadakkedath et al. 2020). Since 2010, vegetation appears to have rapidly colonised the slump floor (Vadakkedath et al. 2020), which may help to stabilise some of the icy badlands (Figure 6b) and reduce the rate of slump expansion, because during the hottest summer days vegetated surfaces are cooler than bare ones.

6.1.2. Mixed periglacial landscapes

Some mixed periglacial landscapes also exemplify characteristic landscapes, where the imprint of ground-ice-related processes is mixed with imprints from azonal processes that show distinctive features in periglacial environments. Such landscapes are found where criteria 1 (persistence), 3 (ice-rich) and 4 (aggradation) are met, and criterion 2 (extraglacial) is of less relevance because sediment aggradation may eventually resurface over inherited glacial landforms. Examples of characteristic mixed periglacial landscapes include periglacial–alluvial landscapes and periglacial–deltaic landscapes.

6.1.2.1. Periglacial–alluvial landscapes

Periglacial–alluvial landscapes comprise a mixture of active periglacial and alluvial landforms formed where wetlands or active floodplains of arctic and subarctic rivers cross permafrost terrain (e.g., the Lena, Yana, Indigirka and Kolyma rivers; Figure 4). In these landscapes, hydrologic processes operating in permafrost drainage basins may (1) concentrate hydrologic activities, (2) spatially differentiate water distribution in different parts of basins, (3) store liquid water and ice, (4) exaggerate hydrologic events, and (5) favour high runoff ratios (Woo 2012, pp. 514–515).
Watercourses in periglacial wetlands are modified by fluvial processes, growth and thaw of permafrost, and vegetation changes (Vandenberghe and Woo 2002).

Frozen ground in or adjacent to stream channels can both protect and enhance riverbeds and banks from erosion (Woo 2012, p.432). Ice bonding of unconsolidated material can protect it from erosion from running water, especially during the snowmelt season, limiting downcutting. But ice also impedes infiltration and percolation, promoting surface runoff. Upward growth of permafrost into newly deposited alluvial sediments—e.g., in levées and active floodplains adjacent to river channels—limits or prevents lateral seepage and helps to stabilize the channels (Woo 2012, p.433). Conversely, melt of ground ice may facilitate erosion. For example, thermal erosion occurs where flowing water melts ground ice by the combined effects of heat conduction and convection, and then mechanically erodes newly released sediment or rock fragments. Thermal erosion along riverbanks through ice-rich unconsolidated sediments causes undercutting and rapid bank retreat (Church and Miles, 1982; Lawson, 1983). Undercutting by currents excavates a horizontal cleft (thermo-erosional niche) that may extend 10 m or more laterally into the bank, at about water level. Above the niche, the undermined permafrost episodically collapses in large blocks, often along ice wedges. Additionally, ice wedges can provide routes for gully erosion (Fortier et al. 2007), with gullies sometimes enlarging into creeks (Woo 2012, p.433). Melting wedge ice at nodes where ice-wedge polygons intersect can create beaded streams. In the long-term, it has been suggested that “…channels transport sediment and expand their network more effectively under permafrost conditions than under nonpermafrost conditions…” (Vandenberghe and Woo 2002, p.495).

Although some periglacial–alluvial landscapes can be regarded as characteristic landscapes in equilibrium with the present process regime, others are polygenetic landscapes. This distinction centres on whether or not criterion 4 is met. Characteristic periglacial–alluvial landscapes are those in which aggradation of clastic sediment—with or without accumulation of organic
material—leads to rise of the depositional surface and resulting upward aggradation of permafrost (criterion 4). Permafrost aggradation can be *syngenetic* where clastic sediment aggrades incrementally, *quasi-syngenetic* where organic material accumulates without sedimentation (Shur 1988), or a combination of both. The landsurface includes modern floodplains adjacent to major rivers crossing permafrost terrain (e.g., the Lena and Kolyma rivers) as well as active alluvial fans, for example the Holocene fans in the Zackenberg Valley of northeast Greenland (Figure 1; Cable et al. 2018). Other periglacial–alluvial landscapes, however, do not currently experience aggradation of sediment and permafrost. For example, rivers such as the Nigu, Ikpikpuk and Titaluk draining the North Slope of Alaska have incised their floodplain since the early Holocene. The incision is leading to reworking of alluvial deposits dating from two older phases of valley aggradation—during the Pleistocene–Holocene transition (Mann et al. 2010), resulting in a polygenetic periglacial landscape (section 6.2).

6.1.2.2. Periglacial–deltaic landscapes

Periglacial–deltaic landscapes develop in arctic and subarctic regions, where rivers draining basins underlain by permafrost discharge water and sediment into the sea. Active deltas in the Arctic range in size from <0.1 ha in recently tapped deltaic lakes to thousands of km², and in age from newly formed to mid Holocene, as the rate of sea-level rise began to decrease (Walker 1998). Such deltas are geologically young and their geomorphic and sedimentary systems reflect recent to modern conditions, allowing them to develop characteristic landscapes. The larger deltas include the Olenik, Lena, Yana, Indigirka and Kolyma in Siberia (Magritsky et al. 2013) and the Colville Delta in northern Alaska (Walker and Hudson 2003) (Figure 4), though not all parts of the deltas are active; for example, ~40% of the overall area of the Lena Delta consists of pre-Holocene plains and bedrock (Are and Reimnitz 2000). Permafrost features in active deltas commonly include ice wedges, ice-wedge polygons, pingos and thermokarst lakes. Ponds in low-centred polygons and in the troughs above ice wedges are particularly common (Walker 1998, 2002).
Thermokarst activity can lead to rapid geomorphic change in arctic deltas. Thermal erosion, particularly during spring breakup of river ice, undercuts distributary banks by carving thermoerosional niches that commonly result in block collapse. In addition, thermokarst lakes can expand (Nitze and Grosse 2016) or drain over decadal timescales. Emergent flats and bars in summer provide sources for wind to rework the sand into dunes in down-drift locations, for example on the Lena River Delta. Ponds in blowouts in dunes or in depressions between dune belts may develop because permafrost impedes drainage.

6.2 Polygenetic periglacial landscapes

As discussed in section 1, polygenetic landscapes express only partial adjustment of the morphology to present environmental conditions and geomorphic processes, usually in highly localized areas, i.e., at the microform scale. Thus, polygenetic periglacial landscapes inherit relict or inactive periglacial landsurfaces that are modified at varying rates and to varying degrees under extant geomorphic conditions and processes, which may or may not be periglacial. Such landscapes occur where criteria 1 (persistence) and 2 (extraglacial) are always met (except in the case of glacial–periglacial landscapes; section 6.2.5), and criteria 3 (ice-rich) and 4 (aggradation) are sometimes met. Five types of polygenetic periglacial landscapes are identified: (1) relict accumulation plains and aprons, (2) frost-susceptible bedrock landscapes, (3) cryopediments in glaciated and unglaciated terrains, (4) non-frost-susceptible bedrock landscapes, and (5) glacial–periglacial landscapes.

6.2.1. Relict accumulation plains and aprons

Relict accumulation plains and aprons developed during past periglacial conditions where sedimentation resurfaced areas of permafrost landscape, prior to environmental changes that left the resurfaced landscapes as relicts in a different process regime operating under present-day periglacial conditions. The plains and aprons are underlain by thick deposits of silt, sand or gravel that contain variable amounts of ground ice and can be mapped regionally as cryostratigraphic
units. The units developed where clastic sedimentation—by wind, water or gravity—occurred on millennial timescales during multiple cold (glacial) stages of the Plio–Pleistocene, enabling aggradation of sedimentary units metres to tens of metres thick, burying underlying landsurfaces. The relict accumulation plains and aprons are widespread in some lowlands and foothills in Beringia, and commonly form stacked sequences that extend back in time through the Late Pleistocene and sometimes to the Middle or Early Pleistocene or even the Pliocene. For example, a number of accumulation levels have been distinguished in the Kolyma Lowland of northeast Siberia (Figure 4), underlain by yedoma and other Pleistocene deposits of varied origin (Sher et al. 1979). Prominent in the present landscape of Beringian lowlands are plains underlain by ice complexes, particularly those complexes dating from the Late Pleistocene, which tend to have modified less than older surfaces. Such ice complexes developed where silt, sand or silt–sand intergrades accumulated incrementally on landsurfaces, burying them beneath metres to tens of metres of sediment in which syngenetic permafrost aggraded and incorporated abundant excess ice. On lowlands, the resulting new landsurfaces formed accumulation plains such as the Omolon–Anyuy yedoma surface (>1000 km²; Vasil’chuk et al. 2001) in the Kolyma Lowland, for example at Duvanny Yar, the yedoma typesite in Russia (Figures 4 and 7) (Murton et al. 2015). On foothills, the new landsurfaces tended to form aprons, for example at the Batagay Megaslump in the Yana Uplands (Figures 4 and 6b) (Murton et al. 2017). The 50-m high vertical sections in the slump headwall expose four major cryostratigraphic units stacked on top of each other—two ice complexes and two sand deposits—that indicate episodic accumulation of sand, ground ice and syngenetic permafrost from the early Middle Pleistocene ~650 ka to the Late Pleistocene (Murton et al. in review). Other examples of ice complexes and intercalated cryostratigraphic units underlie accumulation plains along the northern coastlands of western Beringia (Opel et al. 2017; Wetterich et al. 2019), and occupy valleys of interior Alaska (Westgate et al. 1990; Péwé et al. 2009).
At the end of the Pleistocene, ice complexes ceased forming as the environment of Beringia abruptly changed from dry, dusty and vegetated with steppe plants to moist, boggy and vegetated with tundra shrubs and taiga forest. Thus, many accumulation plains and aprons are relict Pleistocene features, and although they are periglacial in origin, they are not currently forming and so constitute polygenetic periglacial landscapes. Such Pleistocene accumulation plains and aprons have been modified in many places by Holocene thermokarst activity. Thus, the plains are inset with numerous thermokarst landforms such as alas and icy badlands (section 6.1). The alas terrain has evolved by thaw of ice-rich inter-alam areas, which comprise 10.8% (~335,000 km²) of the total area of the Sakha Republic (Fedorov et al. 2018). If in the future the climate cooled again, and ice-age conditions similar to those in Pleistocene cold stages recurred, then the accumulation plains would likely be reactivated.

6.2.2. Frost-susceptible bedrock landscapes

Frost-susceptible bedrock favours development of excess ice in near-surface permafrost (transition zone), which—combined with long-term uplift and stream incision—promotes periglacial mass wasting and shaping of landscapes dominated by convexo–concave hillslopes. For example, the unglaciated landscape of southern England was described in a benchmark paper by M.T. Te Punga (1957, p.410–411; Worsley 2005) as “characteristically whale-backed” and “subdued”, and interpreted as “a typical relict periglacial landscape”, supported by subsequent inferences by French (1972). This landscape is exemplified by rock types that are lithologically different but frost-susceptible: (1) chalk plateaux of Kent to Dorset, and (2) slate and schist lowlands of south Devon (Figure 8a).

The chalklands can be classified geomorphologically into plateaux, sediment-mantled hillslopes, slope-foot and valley landsystems (Figure 9); and geologically the upper metres of their substrate can be distinguished in terms brecciated chalk bedrock, and deposits of silt, diamicton and gravel (Murton and Ballantyne 2017). During recurrent Pleistocene periglaciation, frost-susceptible
bedrock such as chalk was fractured by ice segregation, producing an ice-rich brecciated layer in
the upper metres of permafrost, with additional fracture in seasonally frozen ground during less
severe periglacial conditions (Murton 1996). The ice-rich layer was vulnerable to thaw and
consolidation, which released debris into the active layer (Murton et al. 2006) and, in undrained
conditions, resulted in elevated porewater pressures and sediment deformation. Diffusive
sediment transport processes of mass movement (e.g., gelifluction and frost creep) and overland
flow (slopewash) moved sediment downslope and smoothed the topography, creating extensive
convexo–concave hillslopes (Figure 10a). Similar descriptions and interpretations can be applied to
the slate and schist lowlands of south Devon (Figure 10b; Murton and Lautridou 2003).

Incision of the chalklands and of the slate and schist lowlands occurred as streams cut down
through the uplifting landscape (section 7.3) and formed valley networks. Many of the valleys are
presently dry, especially in the chalklands, which has sparked debate about the dominant
environmental conditions associated with their formation: either (a) ‘normal’ fluvial erosion under
temperate conditions somewhat similar to those at present and/or (b) periglacial conditions that
fundamentally changed the geomorphic system during the Pleistocene (reviewed by Gregory
1971; Ballantyne and Harris 1994, pp.152–155). Some consensus suggests the larger, high-order
valleys initiated and developed by fluvial erosion under temperate conditions in the Neogene to
Middle Pleistocene but have been modified later by periglacial processes, whereas the smallest,
low-order valleys initiated and developed wholly under periglacial conditions as recently as the
last cold stage (Morgan 1971; Gregory 1971; French 2017b, p.364). Incision of many presently dry
valleys was likely driven, at least partly, by streams episodically discharging seasonal snowmelt or
summer rain, while seasonal frost or permafrost ground rendered the bedrock impermeable.

The smallest dry valleys—shallow unchanneled hillslope hollows, commonly bowl- or paddle-
shaped (French 1972)—are known as dells (Gregory 1971; French 2017b, p.364), hollows or zero-
order drainage basins (Dietrich et al. 1986, 1987). The dells are typically underlain by chalky
gelifluction deposits and have been attributed to snowmelt erosion under cold-climate conditions, i.e., concentrated wash processes beneath snowpatches when the subsoil was either seasonally or perennially frozen (French 1976, p.268). French (2017b, p.364) considered them analogous to *nivation hollows* and raised the possibility that they are “truly periglacial in origin”. What seems clear from our growing understanding of hillslope hydrology in permafrost regions is seepage of water through the active layer in summer, perhaps along *water tracks* (McNamara et al. 1999), and flow re-emergence near the base of hillslopes can be important geomorphic processes that likely influence first-order drainage basins (Gardiner 1995). Such processes probably contributed to dell growth and, possibly, to their initiation.

**6.2.3. Cryopediments in glaciated and unglaciated terrains**

Cryopediments are low-angle erosional surfaces that, at their proximal (upslope) end, connect to a steeper backslope such as the edge of a hill, mountain terrace or the steep side of a dry valley—commonly with a break-of-slope—and at their distal (downslope) end they commonly grade into river terraces or valley floors (Figure 11; Priesnitz 1988; Vandenberghe and Czudek 2008; Ballantyne 2018, pp. 221–222). Some cryopediments have more than one level, giving the landscape a stepped appearance. Their gradients are reported to vary from 0.03° to 12°, with the slopes gently concave to rectilinear. Widths (at right angles to the main valley axis) range from a few tens of metres to several kilometres, and lengths (down valley) may attain several tens of kilometres. Beneath the landsurface of cryopediments, the erosional surface truncates underlying bedrock or unconsolidated deposits and may be subaerially exposed (a bare cryopediment) or buried by a veneer of unconsolidated sediments. The veneer rarely exceeds 1–2 m in thickness, and varies texturally from gravel through diamicton to pebbly sand, and various fine-grained deposits. Relict Pleistocene cryopediments occur in both glaciated and unglaciated terrains, including areas in the Czech Republic (Vandenberghe and Czudek 2008), Middle Poland and France (Dylik 1957), lowlands in northwest Germany (Liedtke 1983), Belgium and the Netherlands.
Cryopediments considered largely inactive occur, for example, in unglaciated northern Yukon, Canada (Figure 11; French and Harry 1992) and the Yukon–NWT border, Canada (Fried et al. 1993). In many cases, cryopediments are dissected by gullies, dells or river valleys. Development of cryopediments has been attributed to lowering of the pediment surface, retreat of the backslope, or a combination of both, operating under periglacial conditions. The geomorphic processes driving lowering are thought to be mainly overland flow, rill wash and gully ing, particularly in the lower parts of cryopediments, where runoff is concentrated (Vandenberghe and Czudek 2008). The water source is inferred to be runoff generated largely from snowmelt or heavy rainfall over frozen and therefore impermeable subsoil. Field observations from a glaciated landscape underlain by continuous permafrost at Zackenberg, northeast Greenland (Figure 1) show the importance of snow meltwater from in generating sheetwash and eroding small, shallow channels in unconsolidated glacial sediments, as well as enhancing gelifluction downslope of perennial and seasonal snowpatches (Christiansen 1998). Retreat of the backslope has been attributed to sheetwash and mass movement (e.g., gelifluction) concentrated within and enlarging troughs or dry valleys (Dylik 1957), by active-layer detachment on backslopes or seepage at their feet (Vandenberghe and Czudek 2008) or by frost weathering and debris sliding on rocky slopes (Priesnitz 1988). According to Vandenberghe and Czudek (2008), vertical lowering of cryopediments by surface runoff always occurs and tends to dominate over retreat of backslopes. But where retreat of backslope dominates over lowering of the pediment, then the cryopediments function primarily as surfaces of transportation (Dylik 1957), for example by gelifluction (French 1973). Low frost-susceptibility of sandy to gravelly substrates disfavour gelifluction and promote sheetwash, whereas finer, more silty substrates favour more gelifluction (Liedtke 1983). Overall, it seems likely that some cryopediments developed by periglacial modification of pre-Quaternary
pediments (Figure 11), whereas others, particularly on soft substrates were initiated and developed during cold Quaternary stages (Ballantyne 2018).

6.2.4. Non-frost-susceptible bedrock landscapes

Non-frost-susceptible bedrock hosts little or no segregated ice during seasonal or perennial freezing. Dense crystalline rocks such as unweathered granite or sedimentary rocks whose original pores are now largely filled with cements also tend to be frost-stable on account of their limited or discontinuous unfrozen water films and their inherent tensile strength. As a result of the limited amounts of ground ice and the tensile strength, periglacial tends to modify landscapes underlain by such rocks to a lesser degree than in frost-susceptible substrates. Nonetheless, periglacial conditions operating for many millennia can fashion a variety of distinctive periglacial landforms, as illustrated from granite landscapes of Dartmoor (Devon, southwest England; Figure 8a), and Mount Sedgewick (Buckland Hills, northern Yukon, Canada; Figure 4).

The landscapes of Dartmoor and Mount Sedgewick contain granitic tors and blockfields (Figure 12), the tors occurring on summits and hillsides. The Mount Sedgewick granite area includes a broad flat summit veneered with angular rock fragments (French 2016). Disaggregated bedrock material, consisting of sand- to pebble-sized particles and sometimes clay (grus), is common in both areas. Although grus is attributed primarily to mechanical weathering, strong reddish brown colours indicative of oxidation are sometimes present. On hillslopes, periglacial mass wasting commonly transports grus downslope, incorporating larger pebbles, cobbles and sometimes boulders as well as any other pre-existing materials, producing deposits of diamicton (Gerrard 1990). Overall, the frost-shattered outcrops and boulder-strewn slopes of Dartmoor (Gerrard 1991) and Mount Sedgwick indicate that the hillslopes have been strongly affected by past periglacial activity.

The age of such landscapes, however, is difficult to determine. Lichens are widespread on the tors and blockfields, giving the appearance that these landforms are essentially inactive, both in
the modern periglacial environment of northern Yukon and the largely past periglacial environment of Dartmoor. In terms of the age of granite tors on Dartmoor, $^{10}\text{Be}$ cosmogenic isotope dating of 32 samples from the tops of 28 tors has yielded apparent minimum exposure ages of $9.2 \pm 1.0$ ka to $117.7 \pm 6.6$ ka, with a clustering of apparent exposure ages around $36–50$ ka, in MIS 3 (Gunnell et al. 2013). From these ages, Gunnell et al. suggested that most of the sampled tors are young landforms that did not emerge until regolith was stripped off them at the onset of periglaciation ~115 ka, following the last interglacial (MIS 5e). Tor age remains to be established definitively, however, because of factors that complicate the interpretation of cosmogenic ages, for example the rock shielding effects of putative occupation of northern Dartmoor by a Late Devensian icefield and uncertain rates of micro-erosion of tor surfaces. No such dating is available for the Mount Sedgewick landforms, and so their absolute age is unknown.

6.2.5. Glacial–periglacial landscapes

Glacial–periglacial landscapes develop in glaciated terrains where periglaciation has modified a pre-existing glacial landscape. The degree of modification varies substantially (Evans 2013, 2016, 2017), depending on factors such as the duration of postglacial periglaciation (criterion 1), the ground-ice content of the substrate (criterion 3) and the nature and degree of sediment and permafrost aggradation (criterion 4). Detailed consideration of these factors is beyond the scope of this study, though some general comments are appropriate.

First, the timing and duration of postglacial periglaciation provide a broad time framework to consider the evolution of glacial–periglacial landscapes. This can be illustrated in terms of periglacial regions distinguished by the timing and duration of periglaciation operating on the landscape of the UK and Ireland (Murton and Ballantyne 2017). Four regions (1 to 4) were conditioned by glaciations of different age, and a fifth region (5) lay beyond the maximum limit of Pleistocene glaciation (Figure 8a). In this framework, region 4—last glaciated ~430 ka, during MIS 12—has had many tens of thousands of years to experience substantial periglacial modification of
landforms during multiple periglacial episodes in MIS 10, 8, 6, and 4 to 2, in contrast to limited periglaciation of region 1—last glaciated during the Younger Dryas Stadial (12.9–11.7 ka).

Second, the ground-ice content of glaciated terrains strongly influences their sensitivity and periglacial modification. In ice-rich permafrost terrain such as the Tuktoyaktuk Coastlands of northwest Canada (Figure 3; Mackay 1963; Rampton 1988; Burn and Kokelj 2009), Pleistocene till sheets have been extensively modified by Holocene thermokarst activity and gelifluction, and numerous permafrost landforms such as pingos and thermokarst lakes have developed. Thus, the region is one of the most intensely studied anywhere by periglacial geomorphologists (see references citations in French 2017b and Ballantyne 2018). By contrast, in ice-poor permafrost regions such as rocky outcrops of the Canadian Shield, the effects of periglaciation are more subtle (e.g., frost-heaved bedrock; Dyke 1984). Even within different regions of the shield (e.g., the Slave Geological Province; Figure 3), ground-ice content and terrain sensitivity vary substantially, with ice-rich hummocky till and silty clay marine sediments being particularly sensitive (Wolfe et al. 2017). Likewise, in palaeo-contexts, glaciated landscapes where fine-grained frost-susceptible till sheets are widespread have been extensively modified by periglacial processes, as in eastern England (regions 3 and 4 in Figure 8a; Ballantyne and Harris 1994). Where the glacial deposits are frost-susceptible, their upper metres can be strongly modified by ground-ice-related periglacial processes, as is widespread in lowlands glaciated during the late Cenozoic (Figure 3). Details about such paraglacial landscape modification are reviewed by Ballantyne (2002, 2003, 2013) and recently illustrated from the Antarctic Peninsula (Ruiz-Fernández et al. 2019).

Third, sediment aggradation under periglacial conditions can bury former glacial landscapes. A case in point is the modern Mackenzie Delta, NWT, Canada (Figure 3), which is a postglacial (Holocene) feature developing under present-day permafrost conditions (Hill et al. 2001; Burn and Kokelj 2009). Holocene deltaic sedimentation has buried the older glacial landscape beneath tens
of metres or more of sediments, and so the delta landscape is geomorphologically similar to that of many other arctic deltas (section 6.1.2.2).

7 Discussion

7.1 Sensitivity of periglacial landscapes

The sensitivity of periglacial landscapes is determined fundamentally by the amount and distribution of ground ice. Ice-rich landscapes are the most sensitive, because near-surface ice can grow, crack and melt over timescales of hours to decades. As a result, sensitive landforms and sedimentary structures such as pingos and ice wedges change sufficiently rapidly to measure over human timescales (e.g., Mackay 1998, 2000; Liljedahl et al. 2016). Near-surface ground ice is particularly vulnerable to melting, which is triggered by multiple surface disturbances and commonly causes thermokarst activity (Murton 2009). For example, icy badlands evolve during the course of single summers, such as the efficacy of thermal erosion in large thaw slumps (Figure 6). In areas of frost-susceptible bedrock (e.g., chalk), dells appear to be a fairly sensitive periglacial landform, though detailed process studies are needed to monitor their initiation and growth, perhaps in arctic regions of ice-rich shale or slate. In contrast, insensitive landforms such as tors, blockfields, cryoplanation terraces and cryopediments tend to be ice-poor, and their change measured in terms of human timescales is commonly imperceptible; not surprisingly, therefore, many blockfields, blockstreams, cryoplanation terraces and cryopediments are apparently relict (Ballantyne 2018). However, in some instances, certain of these periglacial landforms are still active—as indicated by features such as gelifluction forms, frost-jacked boulders, late-lying snowbanks and freshly fractured clasts—but not as intensively as during full-glacial intervals (F. Nelson, pers. comm. 2020). Ideally, thresholds of geomorphic activity need to be defined (as with ice-wedge cracking; Mackay 1992), though this is not possible at present, given current knowledge. Some information, nonetheless, is becoming available on rates of cryoplanation development (Nyland and Nelson 2020a; Nyland et al. 2020).
Sensitivity also varies spatially *within* periglacial landscapes. On the Dartmoor granite landscape, interfluves and plateaux are thought to be largely unaffected by hillslope changes initiated along river valleys (*Gerrard 1991*). Specifically, the inner areas of plateaux seem to have changed little in their general form throughout the Quaternary. By contrast, valley-side slopes are more sensitive geomorphologically, as indicated by shallow, infilled gully systems that developed by alternating episodes of erosion and deposition during the Holocene (*Gerrard 1991*), after Pleistocene periglacial conditions had ceased on Dartmoor.

In summary, sensitive periglacial landscapes in the mid-latitudes support French’s (2017b, p.369) conclusions that the degree of periglaciation depends largely upon the lithology and persistence of periglacial conditions. The degree of landscape change in mid-latitude regions of past permafrost is greatest in low-relief terrain underlain by frost-susceptible Cenozoic or Mesozoic rocks or unconsolidated Quaternary sediments.

### 7.2 Inheritance

Inheritance of ancient, pre-periglacial landsurfaces is of limited or no significance where thick sequences of ice-rich sediments accumulate in syngenetic permafrost conditions, burying former landsurfaces beneath 10s of metres of Pleistocene sediments. But inheritance may be greater in non-frost-susceptible substrates such as granite than in frost-susceptible ones such as chalk that develop largely by denudation of bedrock.

For the granite upland of Dartmoor, the major topographic features were strongly shaped by dynamic etchplanation during the Palaeogene and by sandy grus weathering during the Neogene (*Brunsden 2007*), with periglacial modification in the Pleistocene being visually evident but relatively minor (*Mignoń and Goudie 2012*). The rolling upland topography with its tors and basins resulted from long-term differential denudation driven by deep weathering and stripping of saprolite (*Green 1985*). The locations of tors on hilltops and spur ends can be explained by long-term weathering and stripping (*Brunsden 2007*), in contrast to many tors on hillsides, which are
more readily attributed to periglacial processes (Gerrard 1994). Likewise, the bevelled summit of Mount Sedgewick is interpreted by French (2016, 2017a) as either a structurally controlled bench (which seems unlikely in a granite intrusion) or the remnants of an ancient erosional surface.

Likewise, the bevelled summit of Mount Sedgewick is interpreted by French (2016, 2017a) as either a structurally controlled bench (which seems unlikely in a granite intrusion) or the remnants of an ancient erosional surface. French (2016) suggested that the major elements of the northern interior Yukon landscape date from the Neogene and very early Quaternary, initiated on a “Tertiary-age peneplain”.

For the chalklands of southern England, a low-relief landsurface was inherited largely from the Palaeogene, until uplift during the late Pliocene to Early Pleistocene initiated incision and development of escarpments (Jones 1999a, 1999b). The major topographic feature of the English chalklands is an erosional surface interpreted as an etchplanated Summit Surface that originated during the Palaeogene. The highest parts of this surface (‘Chalk uplands’ of Jones 1999a) have been shaped continuously thereafter by subaerial processes, whereas the lower parts (‘Backslope Bench’ of Jones 1999a) were first buried by marine sediments and later exhumed during the late Neogene and Pleistocene (Mignon and Goudie 2012); thus both areas of the chalklands became exposed to recurrent periglaciation at the beginning of the Pleistocene.

The wider significance of inheritance of ancient, pre-periglacial landsurfaces in upland periglacial regions—e.g., the Andes, ice-free regions of Antarctica, and areas of Beringia beyond northern Yukon—remains to be determined. Thus, the assertion by French (2017b, p.12) that “…the complete footprint of periglaciation is rarely achieved…” at present remains essentially unsupported by evidence, at least in periglacial uplands.

7.3 Tectonic setting

The tectonic setting influences the imprint of periglaciation on landscape evolution. This is illustrated from England, where studies of tectonic–landscape interactions are well developed (Lee et al. 2018), and uplift by as a much as 200 m of the chalklands has occurred during the late Neogene and Quaternary (last ~3 Ma; Mignon and Goudie 2012), promoting long-term incision.

Thus, the relief observed today in southern England developed mainly during the ~2.6 Ma of the
Quaternary Period and was superimposed on the low-relief land surface that extended across much of southern England at the end of the Neogene.

Erosional incision resulted in denudational (erosional) unloading and isostatic (uplift) adjustment (Clayton and Shamoon 1998a). Differences in rock resistance to denudation have contributed substantially to local relative relief in Britain (Clayton and Shamoon 1998b). Such differences are strongly influenced by frost-susceptibility and the role of periglacial processes in differential erosion (Murton and Belshaw 2011), especially in fluvial incision of weak rocks to form clay vales and creation of escarpments in the extraglacial regions of southeastern England (Clayton and Shamoon 1999). Lane et al. (2008) estimated that denudational isostasy accounted for ~50% of the development of the present-day relief of the Cotswold region (Figure 8a) and ~50% was largely inherited topography determined by differential erosion due to variations in lithology. Nonetheless, the general form of the end-Neogene low-relief surfaces on the chalklands (Jones 1999a) and in the slate and schist lowlands of south Devon remain prominent features of the landscape, though the finer detail of the surface has been strongly sculpted by Pleistocene frost weathering, stream incision and gelifluction (section 6.2.2).

In summary, periglacial weathering (ice segregation) and fluvial incision imposed significant relief on the ancient erosion surfaces, sculpting a palimpsest landscape of ancient plateaux inset by Pleistocene valleys (Figure 10). This broadly supports French’s (2017b, p.369) general conclusion about mid-latitude Pleistocene periglaciation that “Landscape modification centered on stream and valley incision and the planation and flattening of slopes with widespread mass wasting.”

7.5 The efficacy of frost-driven processes

The traditional view that highlights the efficacy of frost-driven (freeze–thaw) processes in driving landscape evolution of periglacial areas has been questioned in recent years, as evidence of landscape inheritance as well as operation of other, azonal weathering processes and rainfall-
induced events has been increasingly recognized (e.g., André 2003; French 2016, 2017b). This recognition provides an appropriate note of caution in interpreting periglacial landscapes, but should be qualified for three reasons.

First, it applies foremost to non-frost-susceptible rocks (e.g., granite, gneiss and quartzite) which, being relatively hard, naturally form uplands (e.g., Dartmoor and Mount Sedgewick). It applies much less to rocks and sediments that are frost-susceptible and tend to occur in lowlands (e.g., chalk and slate).

Second, the view is based principally on geomorphological examination of surface features—e.g., microweathering phenomena, individual boulders, tors, blockfields, scree slopes, debris-flow tracks and lobes, and Richter denudation slopes. Although appropriately directing attention to evidence of some azonal processes, this geomorphological view ‘from above’ needs to consider with equal geocryological rigour the operation of subsurface frost weathering processes ‘out of sight’, at depths of tens of centimetres to several metres in regolith and/or bedrock. Such subsurface observations of ground-ice dynamics—central to periglacial geomorphology (Thorn 1992)—are naturally difficult to obtain. Other methods, such as cosmogenic dating, have been used to infer lowering of Neogene plateau surfaces in northern Sweden and Scotland by a few tens of metres during the Pleistocene as a result of subsurface weathering (Ballantyne 2010, 2018, pp.185–188): lowering of the weathering front at the regolith–rockhead contact, likely by freeze–thaw processes, where water concentrates near the base of a former active layer, may progressively renew the supply of coarse angular clasts to the surfaces of autochthonous blockfields. In essence, this model can be considered a slower version of frost weathering of frost-susceptible plateaux in chalk, slate or schists lowlands, where instead of blockfields covering the plateaux, the latter are underlain by brecciated rock and periglacial colluvium (diamicton) derived from them.
Third, there is growing evidence that frost action can contribute towards a characteristic periglacial landscape formed by erosion of cryoplanation terraces in uplands. Such terraces consist of gently sloping treads and steep risers cut into bedrock and tend to form bench-like staircases that dominate certain hillslopes (Ballantyne 2018; Queen et al. 2020). The terraces are particularly well developed in unglaciated highlands of Beringia and show similar regional trends in elevation to cirques from the last major glaciation (Nelson and Nyland 2017). Some field evidence suggests that cryoplanation terraces develop through locally intensified weathering and mass wasting associated with late-lying snow patches (nivation)(Nyland and Nelson 2020b). Although their development may be relatively slow, they are thought to have been active during multiple cold stages of the Pleistocene. Potentially, therefore, they may represent ‘characteristic [upland erosional] periglacial landscapes’ (Nelson and Nyland 2017, p.315). Of course, if the terraces developed primarily during cold stages of the Pleistocene—when vegetation cover was sparse or discontinuous in areas such as the eastern Beringian highlands, predisposing them to nivation (Nelson and Nyland 2017)—then the terraces are at present weakly active or inactive. They are currently subject to periglacial processes, but at a lesser intensity than during full-glacial conditions.

7.6 Limitations and future priorities

A number of limitations are apparent in the present study and point to future priorities for research in periglacial geomorphology. First, seasonal and diurnal freezing and ground ice should be considered. As French and Thorn (2006, p.172) noted “…permafrost [is] a central, but not defining, element of periglacial geomorphology.” Future work should develop maps of seasonal ground ice in non-permafrost regions.

Second, alpine and plateau periglaciation should be explored systematically. Azonal processes of non-frost weathering (see review by Dixon 2006), mass movement, fluvial action complicate and obscure the periglacial imprint in steep rocky mountainous terrain. Unglaciated montane and
steppe environments in the Qinghai–Tibet Plateau may have strong periglacial signatures. Many
mountain ranges in mid- to high-latitudes of the NH have been glaciated during the late Cenozoic
(Figure 3), directing attention for a periglacial signal to other mountain regions, perhaps the dry,
ever glaciated regions of the Southern Hemisphere: too dry for glaciers to develop, but moist
enough for ground ice.

Third, Antarctic periglaciation in particular and Southern Hemisphere periglacial landscapes in
general should be investigated further. Some ice-free areas of Victoria Land, Antarctica, are
thought to have been ice-free for several million years (e.g., Summerfield et al. 1999), though
whether there has been sufficient ground-ice activity to resurface the inherited pre-periglacial
landscape remains to be determined. Additionally, the role of periglacial processes in shaping
paraglacial landscape development in regions such as the Antarctic Peninsula (Ruiz-Fernández et
al. 2019) appear to be important.

Fourth, quantitative terrain analysis is essential to determine the morphometric properties of
periglacial landforms and landscapes. In turn, this may permit their evaluation at different spatial
scales (i.e., ‘scale linkage’; Dixon 2006) and establishing the degrees to which periglacial
landscapes morphologically resemble or differ from non-periglacial landscapes. Recent advances
in recognition and delineation of cryoplanation terraces in eastern Beringia through semi-
automated digital terrain analysis providing promising opportunities for objective mapping and
interpretation of upland periglacial landscapes (Queen et al. 2020).

Finally, the designation of characteristic landscapes may sometimes be difficult to apply when
climatic and environmental change leads to substantial slowing or cessation of morphological
evolution while periglacial conditions and processes persist, but at lower intensity than previously.
For example, the uplands of Beringia have likely experienced periglacial conditions continuously
during much or possibly all of the Pleistocene, with the result that some periglacial landforms such
as cryoplanation terraces and cryopediments experienced different levels of activity at different times (e.g., greater activity during full-glacial stages versus limited activity during interglacial stages).

8 Conclusions

To answer the two questions posed in section 1, the following conclusions are drawn:

8.1. What and where are periglacial landscapes?

In terms of lowlands, truly periglacial landscapes develop where thermokarst activity has completely resurfaced areas of unglaciated ice-rich permafrost (i.e., alas terrains and icy badlands). In terms of the exposed land currently underlain by permafrost in the NH (~1.2–1.7 x 10^7 km^2) (Zhang et al. 2000), the total area of alas landscapes in the Sakha Republic—a major part of Beringia—is two orders of magnitude less (~1.1 x 10^5 km^2). For uplands, cryoplanation features on the hills and plateaux of Beringia may also represent truly periglacial landscapes, though knowledge of their precise mechanisms and rates of evolution remains limited. Beringian uplands are extensive and form a substantial part of the periglacial realm.

8.2. How complete is the footprint of periglaciation?

Relative rates of periglaciation are fastest and most complete where ground ice is abundant and geomorphologically active (through ice segregation and thermokarst processes) or sediment aggradation is rapid (e.g., in deltas and modern floodplains and wetlands of the Arctic and Subarctic); intermediate and partial where bedrock is frost-susceptible (e.g., chalk and slate); and slowest and partial where it is not (e.g., granite). Thus, periglaciation produces topographic footprints at mesospatial scales (10^3–10^5 m) where: (1) Accumulation plains and aprons underlain by regionally extensive ice complexes developed repeatedly by aggradation of sediment and permafrost during Pleistocene cold stages in Beringia and the unglaciated Siberian platform (Figure 4). (2) Plateaux on which convexo–concave hillslopes inset with valleys (Figure 8) developed in moist, frost-susceptible lithologies in unglaciated lowlands in the mid-latitudes such
as southern England and northern France. Such topography resulted from recurrent development of a palaeo-transition zone of ice-rich brecciated bedrock at the top of past permafrost—supplemented by deep seasonal freezing during milder periglacial conditions—coupled with periglacial mass wasting and stream incision operating on tectonic background of Pleistocene uplift. Finally, the completeness of the topographic footprint of upland cryoplanation and development of cryopediments generally awaits systematic investigation.

Acknowledgements

The present work is dedicated to Hugh French, my former PhD supervisor, friend and colleague. Hugh introduced me to permafrost research and the granite landscape of Mount Sedgewick. Denys Brunsden is thanked for discussions about landscape evolution and supervision of my BSc project on characteristic landforms, some of which was adapted in Brunsden (1990). Discussions about periglacial and glacial landscapes with Hugh French, Colin Ballantyne, Charles Harris, David Evans, Steve Kokelj, Steve Wolfe, Bernd Etzelmüller, Ole Humlum, Hanne Christiansen and Rich Waller are appreciated. Andrea Manica and colleagues are thanked for discussions of cold-climate landscapes in a project reconstructing NH ice-sheet extent through the late Cenozoic (Batchelor et al. 2019). Harold Lovell provided the shape file for the glacial limit in Fig. 5 (see Lukas et al. 2017). Della Murton drafted part of Figure 1, and Ed Holt compiled the basemaps for Figures 3 and 4. Bernd Etzelmüller, Fritz Nelson and Mauro Guglielmin are thanked for their incisive comments on a previous version of the manuscript, which improved it substantially.

Data availability statement:

This article is based primarily on secondary data from published sources listed in the References, supplemented by the author’s field observations.
References

Amante C, Eakins BW. *ETOPO1 1 Arc-Minute Global Relief Model: Procedures, Data Sources and Analysis*. NOAA Technical Memorandum NESDIS NGDC-24. Boulder, Colorado, USA: National Geophysical Data Center, NOAA, 2009. [https://data.nodc.noaa.gov/cgi-bin/iso?id=gov.noaa.ngdc.mgg.dem:316](https://data.nodc.noaa.gov/cgi-bin/iso?id=gov.noaa.ngdc.mgg.dem:316). Accessed December 17, 2019.

André M-F. Do periglacial landscapes evolve under periglacial conditions? *Geomorphology*. 2003;52:149-164.

Andreev AA, Tarasov PE, Wennrich V, et al. Late Pliocene and early Pleistocene environments of the north-eastern Russian Arctic inferred from the Lake El'gygytgyn pollen record. *Clim Past*. 2014;10:1017-1039.

Andrieux E, Bateman MD, Bertran P. The chronology of Late Pleistocene thermal contraction cracking derived from sand wedge OSL dating in central and southern France. *Global Planet Change*. 2018;162, 84-100.

Are F, Reimnitz E. An overview of the Lena River Delta setting: geology, tectonics, geomorphology, and hydrology. *J Coastal Res*. 2000;16(4):1083-1093.

Astakhov VI, Kaplyanskaya FA, Tarnogradskiy VD. Pleistocene permafrost of West Siberia as a deformable glacier bed. *Permafrost Periglac Process*. 1996;7:165-191.

Astakhov VI, Isayeva LL. The 'ice hill': an example of 'retarded deglaciation' in Siberia. *Quat Sci Rev*. 1988;7:29-40.

Bailey I, Hole GM, Foster GL. et al. An alternative suggestion for the Pliocene onset of major northern hemisphere glaciation based on the geochemical provenance of North Atlantic Ocean ice-rafted debris. *Quat Sci Rev*. 2013;75:181-194.
Balco G, Rovey CW II. Absolute chronology for major Pleistocene advances of the Laurentide Ice Sheet. *Geology*. 2010;38:795-798.

Ballantyne CK. Paraglacial geomorphology. *Quat Sci Rev.* 2002;21:1935-2017.

Ballantyne CK. Paraglacial landform succession and sediment storage in deglaciated mountain valleys: theory and approaches to calibration. *Z Geomorph Suppl.* 2003;132:1-18.

Ballantyne CK. A general model of autochthonous blockfield evolution. *Permafrost Periglac Process*. 2010;21:289-300.

Ballantyne CK. Paraglacial geomorphology. In: Elias SA, Mock CJ, eds. *The Encyclopedia of Quaternary Science*. 2nd ed. Vol. 3. Amsterdam, The Netherlands: Elsevier; 2013:553-565.

Ballantyne CK. *Periglacial Geomorphology*. Chichester, UK: Wiley-Blackwell; 2018.

Ballantyne CK, Harris C. *The Periglaciation of Great Britain*. Cambridge, UK: Cambridge University Press; 1994.

Batchelor CL, Margold M, Krapp M, et al. The configuration of Northern Hemisphere ice sheets through the Quaternary. *Nat Commun*. 2019;10:3713.

Benn DI and Evans DJA. *Glaciers and Glaciation*. 2nd edn. London, UK: Hodder Education; 2010.

Berthling I, Etzelmüller B. The concept of cryo-conditioning in landscape evolution. *Quat Res*. 2011;75:378-384.

Bertran P, Liard M, Sitzi L, Tissoux H. A map of Pleistocene aeolian deposits in Western Europe, with special emphasis on France. *J Quat Sci*. 2016;8:844-856.

Boer DH de. Hierarchies and spatial scale in process geomorphology: a review. *Geomorphology*. 1992;4:303-318.
Brigham-Grette J, Melles M, Minyuk P, et al. Pliocene Warmth, extreme Polar Amplification, and Stepped Pleistocene Cooling recorded in NE Russia. *Science.* 2013;340:1421-1427.

Brown J, Ferrians O, Heginbottom JA, Melnikov E. *Circum-Arctic Map of Permafrost and Ground-Ice Conditions, Version 2.* [permaice.shp]. Boulder, Colorado USA. NSIDC: National Snow and Ice Data Center; 2002. [https://nsidc.org/data/ggd318](https://nsidc.org/data/ggd318). Accessed December 17, 2019.

Brunsden D. Tablets of stone: toward the Ten commandments of Geomorphology. *Z Geomorph Suppl.* 1990;79:1-37.

Brunsden D. The persistence of landforms. *Z Geomorph. N.F. suppl.-Bd.* 1993;93:13-28.

Brunsden D. A critical assessment of the sensitivity concept in geomorphology. *Catena.* 2001;42:99-123.

Brunsden D. The geomorphological evolution of Dartmoor. In: André M-F, Étienne S, Lageat Y, Le Coeur C, Mercier D. eds. *Du continent au basin versant. Théories et pratiques en géographie physique. Hommage au Professeur Alain Godard.* Clermont-Ferrand; Presses Universitaires Blaise Pascal; 2007:63-75.

Brunsden D, Thones JB. Landscape sensitivity and change. *Trans Inst Brit Geogr.* 1979;4:463-484.

Büdel J. *Climatic Geomorphology.* Princeton, NJ: Princeton University Press; 1982.

Burn CR, Kokelj SV. The environment and permafrost of the Mackenzie Delta area. *Permafrost Periglac Process.* 2009;20:83-105.

Cable S, Christiansen HH, Westergaard-Nielsen A, Kroon A, Elberling B. Geomorphological and cryostratigraphical analyses of the Zackenberg Valley, NE Greenland and significance of Holocene alluvial fans. *Geomorphology.* 2018;303:504-523.
Cailleux A. 1943. Fissuration de la craie par le gel. Bulletin de la Société Géologique de France. 1943;13: 511-520.

Candy I, Silva B, Lee J. 2011. Climates of the early Middle Pleistocene in Britain: environments of the earliest humans in Northern Europe. In: Ashton N, Lewis SG, Stringer C. eds. The Ancient Human Occupation of Britain. Amsterdam, The Netherlands: Elsevier; 2011:11-22.

Chang XL, Jin HJ, He RX, Lü LZ, Harris SA. Evolution and changes of permafrost on the Qinghai-Tibet Plateau during the Late Quaternary. Sciences Cold Arid Regions. 2017;9(1):1-19.

Chorley RJ, Schumm SA, Sugden DE. Geomorphology. London, UK: Methuen; 1984.

Church MA, Miles MJ. Discussion: processes and mechanisms of river bank erosion. In: Hey RD, Bathurst JC, Thorne CR. eds. Gravel-bed Rivers. New York: Wiley & Sons; 1982:259-271.

Clark PU, Dyke AS, Shakun JD, et al. The Last Glacial Maximum. Science. 2009;325: 710-714.

Clayton KM, Shamoon N. A new approach to the relief of Great Britain: I. The machine-readable database. Geomorphology. 1998a;25,31-42.

Clayton KM, Shamoon N. A new approach to the relief of Great Britain: II. A classification of rocks based on Relative Resistance to Denudation. Geomorphology. 1998b;25:155-171.

Clayton KM, Shamoon N. A new approach to the relief of Great Britain III. Derivation of the contribution of neotectonic movements and exceptional regional denudation to the present relief. Geomorphology. 1999;27:173-189.

Christiansen HH. 1998. Nivation forms and processes in unconsolidated sediments, NE Greenland. Earth Surf Process Landf. 1998;23:751-760.

Czudek T, Demek J. Thermokarst in Siberia and its influence on the development of lowland relief. Quat Res. 1970;1:103-120.
Dietrich W E, Wilson CJ, Reneau SL. Hollows, colluvium and landslides in soil-mantled landscapes. In: Abrahams AD, ed. Hillslope Processes. Boston, MA: Allen & Unwin; 1986:362-388.

Dietrich WE, Reneau SL, Wilson CJ. Overview: “zero-order basins” and problems of drainage density, sediment transport and hillslope morphology. In: Beschta RL, Blinn T, Grant GE, Ice GG, Swanson FJ, eds. Proceedings of the International Symposium on Erosion and Sedimentation in the Pacific Rim held at Oregon State University, Corvallis, Oregon, USA, 3-7 August 1987. Wallingford, UK: International Association of Hydrological Sciences Publication 165;1987:27-37.

Dixon J. Scale in periglacial geomorphology. Géomorphologie-Relief Processus Environnement. 2006;3:175-186.

Dyke AS, Evans DJA. Ice-marginal terrestrial landsystems: northern Laurentide and Innuitian ice sheet margins. In: Evans DJA ed. Glacial Landsystems. London, UK: Arnold; 2003:143-165.

Dyke LD. Frost heaving of bedrock in permafrost regions. Bull Assoc Engin Geol. 1984;21:389-405.

Dylik J. Tentative comparison of planation surfaces occurring under warm and under cold semi-arid climatic conditions. Biul Peryglac. 1957;5:175-186.

Egholm DL, Andersen JL, Knudsen MF, Jansen JD, Nielsen SB. The periglacial engine of mountain erosion – Part 2: Modelling large-scale landscape evolution. Earth Surf Dynam. 2015;3:463-482.

Ehlers J, Gibbard PL, Hughes PD. eds. Quaternary Glaciation Extent and Chronology: a Closer Look. Developments in Quaternary Science 15. Amsterdam, The Netherlands: Elsevier; 2011.

Elderfield H, Ferretti P, Greaves M, et al. Evolution of ocean temperature and ice volume through the Mid-Pleistocene climate transition. Science. 2012;337,704-709.

Elias SA, Brigham-Grette J. Late Pleistocene glacial events in Beringia. In: Elias SA, Mock CJ, eds. The Encyclopedia of Quaternary Science. 2nd ed. Vol. 2. Amsterdam, The Netherlands: Elsevier; 2013:191–201.
European Environment Agency, 2017. *IPA permafrost map*. https://www.eea.europa.eu/data-and-maps/figures/permafrost-in-the-northern-hemisphere. Accessed February 19, 2020.

Evans DJA. Glacial landsystems. In: Elias SA, Mock CJ, eds. *The Encyclopedia of Quaternary Science*. 2nd edn. Vol. 1. Amsterdam, The Netherlands: Elsevier; 2013:813-824.

Evans DJA. Landscapes at the periphery of glacierization—retrospect and prospect. *Scottish Geographical J*. 2016;132(2):140-163.

Evans DJA. Conceptual glacial ground models: British and Irish case studies. In: Griffiths JS, Martin CJ, eds. *Engineering Geology and Geomorphology of Glaciated and Periglaciated Terrains—Engineering Group Working Party Report*. London, UK: The Geological Society of London, Engineering Group Special Publications 28; 2017:369-500.

Fedorov AN, Vasilyev NE, Torgovkin YI et al. Permafrost–landscape map of the Republic of Sakha (Yakutia) on a scale 1:1,500,000. *Geosciences*. 2018;8:465; doi:10.3390/geosciences8120465

Fortier D, Allard M, Shur Y. Observation of rapid drainage system development by thermal erosion of ice wedges on Bylot Island, Canadian Arctic Archipelago. *Permafrost Periglac Process*. 2007;18:229-243.

French HM. Slope asymmetry of the Beaufort Plain, northwest Banks Island, N.W.T., Canada. *Can J Earth Sci*. 1971;8:717-731.

French HM. Asymmetrical slope development in the Chiltern Hills. *Biul Peryglac*. 1972;21:51-73.

French HM. Cryopediments on the Chalk of southern England. *Biul Peryglac*. 1973;22:149-156.

French HM. *The Periglacial Environment*. 1st edn. London, UK: Longman; 1976.

French HM. Recent contributions to the study of past permafrost. *Permafrost Periglac Process*. 2008;19:179-194.
French HM. Do periglacial landscapes exist? A discussion of the upland landscapes of northern interior Yukon, Canada. *Permafrost Periglac Process.* 2016;27:219-228.

French HM. The northern interior Yukon: an example of periglaciation. In: Slaymaker O, ed. *Landscapes and Landforms of Western Canada.* World Geomorphological Landscapes. Cham, Switzerland: Springer; 2017a:257-266.

French HM. *The Periglacial Environment.* 4th edn. Chichester, UK: Wiley-Blackwell; 2017b.

French HM, Bennett L, Hayley DW. Ground ice conditions near Rea Point and on Sabine Peninsula, eastern Melville Island. *Can J Earth Sci.* 1986;23:1389-1400.

French HM, Harry DG. Pediments and cold-climate conditions, Barn Mountains, unglaciated Northern Yukon, Canada. *Geogr Annal A.* 1992;74(2/3),145-157.

French H, Thorn CE. The changing nature of periglacial geomorphology. *Géomorphologie-Relief Processus Environnement.* 2006;3:165-173.

Fried G, Heinrich J, Nagel G, Semmel A. Periglacial denudation in formerly unglaciated areas of the Richardson Mountains (NW Canada). *Z Geomorph.* 1993;92:55-69.

Froese DG, Westgate JA, Reyes AV, Enkin RJ, Preece SJ. Ancient permafrost and a future, warmer Arctic. *Science.* 2008;321:1648.

Gardiner V. Channel networks: progress in the study of spatial and temporal variations of drainage density. In: Gurnell AM, Petts GE, eds. *Changing River Channels.* Chichester, UK: Wiley; 1995:65-85.

Gerrard J. The status of temperate hillslopes in the Holocene. *Holocene.* 1991;1(1):86-90.

Gerrard AJ. Variations within and between weathered granite and head on Dartmoor. *Proc Ussher Soc.* 1994;7;285-288.
Gerrard J. Classics in physical geography revisited. Linton, D.L. 1955: The problem of tors. The Geographical Journal 121, 470–187. Prog Phys Geog. 1994;18(4):559-563.

Green CP. Pre-Quaternary weathering residues, sediments and landform development: examples from southern Britain. In: Richards KS, Arnett RR, Ellis S, eds. Geomorphology and Soils. London, UK: Allen and Unwin: 58-77.

Gregory KJ. Drainage density changes in South-West England. In: Ravenhill WLD, Gregory KJ, eds. Exeter Essays in Geography. Exeter, UK: University of Exeter; 1971:33-53.

Glushkova OYu, Smirnov NN. Pliocene to Holocene geomorphic evolution and paleogeography of the El’gygytgyn Lake region, NE Russia. J Paleolimnol. 2007;37:37-47.

Gullentops F, Janssen J, Paulissen E. Saalian nivation activity in the Bosbeek valley, NE Belgium. Geologie en Mijnbouw 1993;72:125-130.

Gunnell Y, Jarman D, Braucher R, et al. The granite tors of Dartmoor, Southwest England: rapid and recent emergence revealed by Late Pleistocene cosmogenic apparent ages. Quat Sci Rev. 2013;61:62-76.

Harris SA. Thermal history of the Arctic Ocean environs adjacent to North America during the last 3.5 Ma and a possible mechanism for the cause of the cold events (major glaciations and permafrost events). Prog Phys Geogr. 2005;29(2):218-237.

Haug GH, Ganopolski A, Sigman DM, et al. North Pacific seasonality and the glaciation of North America 2.7 million years ago. Nature. 2005;433(7028):821-825.

Head M J. Formal subdivision of the Quaternary System/Period: present status and future directions. Quatern Int. 2019;500:32-51.

Hill PR, Lewis CP, Desmarais S, Kauppaymuthoo V, Rais H. The Mackenzie Delta: sedimentary processes and facies of a high-latitude, fine-grained delta. Sedimentology. 2001;48:1047-1078.
Hoffecker JF, Elias SA. *Human Ecology of Beringia*. New York, NY: Columbia University Press; 2007.

Hoffecker JF, Elias SA, Potapova O. Arctic Beringia and Native American Origins. *PaleoAmerica*. 2020;6(2):158-168.

Hopkins DM. 1982. Aspects of the paleogeography of Beringia during the late Pleistocene. In: Hopkins DM, Matthews JV, Jr, Schweger CE, Young SB. eds. *Paleoecology of Beringia*. New York, NY: Academic Press; 1982:3–28.

Hopkins DM, Matthews JV, Jr, Schweger CE, Young SB. eds. *Paleoecology of Beringia*. New York, NY: Academic Press; 1982.

Hughes ALC, Gyllencreutz R, Lohne ØS, Mangerud J, Svendsen JI. The last Eurasian ice sheets—a chronological database and time-slice reconstruction, DATED-1. *Boreas*. 2016;45:1-45.

Jorgenson MT. 2013. Thermokarst terrains. In: Shroder JF ed-in-chief; Giardino R, Harbor J. eds. *Treatise on Geomorphology, Vol. 8, Glacial and Periglacial Geomorphology*. San Diego, CA: Academic Press; 2013:313-324.

Jorgenson MT, Osterkamp TE. Response of boreal ecosystems to varying modes of permafrost degradation. *Can J Forest Res*. 2005;35:2100-2111.

Jones DKC. Evolving models of the Tertiary evolutionary geomorphology of southern England, with special reference to the Chalklands. In: Smith BJ, Whalley WB, Warke PA, eds. *Uplift, Erosion and Stability: Perspectives on Long-term Landscape Development*. London, UK: The Geological Society of London, Special Publications 162; 1999a:1-23.

Jones DKC. On the uplift and denudation of the Weald. In: Smith BJ, Whalley WB, Warke PA, eds. *Uplift, Erosion and Stability: Perspectives on Long-term Landscape Development*. London, UK: The Geological Society of London, Special Publications 162; 1999b:25-43.
Joyce JE, Tjalsma LRC, Prutzman JM. North American glacial meltwater history for the past 2.3My: oxygen isotope evidence from the Gulf of Mexico. *Geology*. 1993;21:483-486.

Kanevskiy M, Jorgenson T, Shur Y, et al. Cryostratigraphy and permafrost evolution in the lacustrine lowlands of west-central Alaska. *Permafrost Periglac Process*. 2014;25:14-34.

Karte J. Development and present state of German periglacial research in Arctic and alpine environments. *Biul Peryglac*. 1982;29:183-201.

Karte J. Development and present state of German periglacial research in the polar, subpolar and alpine environment. (Entwicklung und gegenwartiger stand der deutschen perigalziarforschung im polaren, subpolaren und alpinen milieu) Technical translation, Ottawa, National Research Council of Canada 1981. [https://doi.org/10.4224/20337838](https://doi.org/10.4224/20337838)

Kokelj SV, Lantz TC, Tunnicliffe J, Segal R, Lacelle D. Climate-driven thaw of permafrost preserved glacial landscapes, northwestern Canada. *Geology*. 2017;45(4):371-374.

Lane NF, Watts AB, Farrant AR. An analysis of Cotswold topography: insights into the landscape response to denudational isostasy. *J Geol Soc Lond*. 2008;165:85-103.

Lawson DE. Erosion of perennially frozen streambanks. Hanover, New Hampshire: U.S. Army Cold Regions Research and Engineering Laboratory Report 83-29; 1983.

Lee JR, Candy I, Haslam R. 2018. The Neogene and Quaternary of England: landscape evolution, tectonics, climate change and their expression in the geological record. *Proc Geol Ass*. 2018;129:452-481.

Liedtke H. Periglacial slopewash and sedimentation in northwestern Germany during the Würm (Weichsel) glaciation. In: *Permafrost, Fourth International Conference, Proceedings, July 17–22, 1983*. Washington, DC: National Academy Press; 1983:715-718.
Liljedahl AK, Boike J, Daanen RP, et al. Pan-Arctic ice-wedge degradation in warming permafrost and its influence on tundra hydrology. Nat Geosci. 2017;9:312-318.

Lindgren A, Hugelius G, Kuhry P, Christensen TR, Vandenberghhe J. GIS-based maps and area estimates of Northern Hemisphere permafrost extent during the Last Glacial Maximum. Permafrost Periglac Process. 2016;27:6-16.

Lisiecki LE, Raymo ME. A Pliocene-Pleistocene stack of 57 globally distributed benthic δ¹⁸O records. Paleoceanogr Paleocl. 2005;20(1):PA1003.

Lukas S, Preusser F, Evans DJA, Boston CM, Lovell H. 2017. The Quaternary. In: Griffiths JS, Martin CJ, eds. Engineering Geology and Geomorphology of Glaciated and Periglaciated Terrains – Engineering Group Working Party Report. London, UK: The Geological Society of London, Engineering Group Special Publications 28; 2017:31-57.

McNamara JP, Kane DL, Hinzman LD. An analysis of an arctic channel network using a digital elevation model. Geomorphology. 1999;29:339-353.

Mackay JR. 1963. The Mackenzie Delta Area, N.W.T. Ottawa, Geographical Branch, Department of Mines and Technical Surveys: Memoir 8; 1963.

Mackay JR. 1992. The frequency of ice-wedge cracking (1967–1987) at Garry Island, western Arctic coast, Canada. Can J Earth Sci. 1992;29:236-248.

Mackay JR. 1998. Pingo growth and collapse, Tuktoyaktuk Peninsula area, western Arctic coast, Canada: a long-term study. Geogr Phys Quatern. 1998;52:271-323.

Mackay, J.R. 2000. Thermally induced movements in ice-wedge polygons, western arctic coast: a long-term study. Geogr Phys Quatern. 2000;54:41-68.

Magritsky D, Mikhailov V, Korotaev V, Babich D. Changes in hydrological regime and morphology of river deltas in the Russian Arctic. In: Young G, Perillo GME, Aksoy H, et al. eds. Deltas:
Mann DH, Groves P, Reanier RE, Kunz ML. Floodplains, permafrost, cottonwood trees, and peat: what happened the last time climate warmed suddenly in arctic Alaska. *Quat Sci Rev.* 2010;29, 3812-3830.

Matheus P, Begét J, Mason O, Gelvin-Reymiller C. 2003, Late Pliocene to late Pleistocene environments preserved at the Palisades site, central Yukon River, Alaska. *Quat Res.* 2003;60:33-43.

Mignon P, Goudie A. Pre-Quaternary geomorphological history and geoheritage of Britain. *Quaestiones Geographicae.* 2012;31:67-79.

Mikhailov IS. Morphological structure of some landscapes in the Arctic zone. In: Govorukha LS, Kruchinin YA. eds. *Problems of Physiographic Zoning of Polar Lands.* New Delhi: Amerind Publishing Company; 1981:155-172.

Morgan RPC. A morphometric study of some valley systems on the English Chalklands. *Trans Inst Brit Geogr.* 1971;54:33-44.

Morse PD, Burn CR, Kokelj SV. Near-surface ground-ice distribution, Kendall Island Bird Sanctuary, western Arctic coast, Canada. *Permafrost Periglac Process.* 2009;20:155-171.

Mortimore RN. *Logging the Chalk.* Dunbeath, UK: Whittles; 2014.

Murton JB. Near-surface brecciation of Chalk, Isle of Thanet, southeast England: a comparison with ice-rich brecciated bedrocks in Canada and Spitsbergen. *Permafrost Periglac Process.* 1996;7:153-164.
Murton JB. Thermokarst sediments and sedimentary structures, Tuktoyaktuk Coastlands, Western Arctic Canada. *Global Planet Change*. 2001;28;175-192.

Murton JB. Ground-ice stratigraphy and formation at North Head, Tuktoyaktuk Coastlands, western Arctic Canada: a product of glacier-permafrost interactions. *Permafrost Periglac Process*. 2005;16:31-50.

Murton JB. Global warming and thermokarst. In: Margesin R. ed. *Permafrost Soils*. Soil Biology Vol. 16. Berlin and Heidelberg: Springer-Verlag; 2009: 185-203.

Murton JB. Ground Ice and Cryostratigraphy. In: Shroder JF ed-in-chief; Giardino R, Harbor J. volume eds. *Treatise on Geomorphology, Vol. 8, Glacial and Periglacial Geomorphology*. San Diego, CA: Academic Press; 2013:173-201.

Murton JB, Ballantyne CK. 2017. Periglacial and permafrost ground models for Great Britain. In: Griffiths JS, Martin CJ, eds. *Engineering Geology and Geomorphology of Glaciated and Periglaciated Terrains – Engineering Group Working Party Report*. London, UK: The Geological Society of London, Engineering Group Special Publications 28; 2017:501-597.

Murton JB, Belshaw R. A conceptual model of valley incision, planation and terrace formation during cold and arid permafrost conditions of Pleistocene southern England. *Quat Res*. 2011;75:285-394.

Murton JB, Edwards ME, Lozhkin AV, et al. Preliminary palaeoenvironmental analysis of permafrost deposits at Batagaika megaslump, Yana Uplands, northern Siberia. *Quat Res*. 2017;87:314-330.

Murton JB, Goslar T, Edwards ME, et al. 2015. Palaeoenvironmental interpretation of yedoma silt (Ice Complex) deposition as cold-climate loess, Duvanny Yar, northeast Siberia. *Permafrost Periglac Process*. 2015;26:207-288.
Murton JB, Lautridou J-P. Recent advances in the understanding of Quaternary periglacial features of the English Channel Coastlands. *J Quat Sci.* 2003;18:301-307.

Murton JB, Opel T, Toms P, et al. (in revision) A multi-method pilot dating study of ancient permafrost, Batagay megaslump, East Siberia. *Quat Res.*

Murton JB, Peterson R, Ozouf J-C. Bedrock fracture by ice segregation in cold regions. *Science.* 2006;314:1127-1129.

Murton JB, Waller RI, Hart JK, Whiteman CA, Pollard WH, Clark ID. Stratigraphy and glaciological structures of a relict deformable bed of permafrost at the northwestern margin of the Laurentide ice sheet, Tuktoyaktuk Coastlands. Canada. *J Glaciol.* 2004;50:399-412.

Natural Earth. *10m rivers lake centerlines.* Made with Natural Earth. Free vector and raster map data @ naturalearthdata.com. 2019a. https://www.naturalearthdata.com/downloads/10m-physical-vectors/10m-rivers-lake-centerlines/. Accessed December 2019.

Natural Earth. *10m admin 0 countries.* Made with Natural Earth. Free vector and raster map data @ naturalearthdata.com. 2019b. https://www.naturalearthdata.com/downloads/10m-cultural-vectors/10m-admin-0-countries/. Accessed December 2019.

Nelson FE, Nyland KE. Periglacial cirque analogs: elevation trends of cryoplanation terraces in eastern Beringia. *Geomorphology.* 2017;293:305-317.

Nitze I, Grosse G. Detection of landscape dynamics in the Arctic Lena Delta with temporally dense Landsat time-series stacks. *Remote Sensing Environ.* 2016;181:27-41.

Nitze I, Grosse G, Jones BM et al. Landsat-based trend analysis of lake dynamics across northern permafrost regions. *Remote Sensing.* 2017;9:640; doi:10.3390/rs9070640

Nyland KE, Nelson FE. Long-term nivation rates, Cathedral Massif, northwestern British Columbia. *Can J Earth Sci.* 2020a;57:1305-1311.
Nyland KE, Nelson FE. Time-transgressive cryoplanation terrace development through nivation-driven scarp retreat. *Earth Surf Process Landf.* 2020b; 45:526-534.

Nyland KE, Nelson FE, Figueiredo PM. Cosmogenic $^{10}$Be and $^{36}$Cl geochronology of cryoplanation terraces in the Alaskan Yukon-Tanana Upland. *Quat Res.* 2020;97:157-166.

O’Neill HB, Burn CR. Physical and temporal factors controlling the development of near-surface ground ice at Illisarvik, western Arctic coast, Canada. *Can J Earth Sci.* 2012;49:1096-1110.

O’Neill HB, Wolfe SA, Duchesne C. New ground ice maps for Canada using a paleogeographic modelling approach. *Cryosphere.* 2019;13:753-773.

Opel T, Wetterich S, Meyer H, Dereviagin AY, Fuchs MC, Schirrmeister L. Ground-ice stable isotopes and cryostratigraphy reflect late Quaternary palaeoclimate in the Northeast Siberian Arctic (Oyogos Yar coast, Dmitry Laptev Strait). *Clim Past.* 2017;13:587-611.

Peltier LC. The geographic cycle in periglacial regions as it is related to climatic geomorphology. *Ann Assoc Am Geogr.* 1950;40(3):214-236.

Péwé TL. The periglacial environment. In: Péwé TL, ed. *The Periglacial Environment.* Montreal, Canada: McGill-Queen’s University Press; 1969:1-9.

Péwé TL, Westgate JA, Preece SJ, Brown PM, Leavitt SW. Late Pliocene Dawson Cut Forest Bed and new tephrochronological findings in the Gold Hill Loess, east-central Alaska. *Geol Soc Am Bull.* 2009;121(1/2):294-320.

Priesnitz K. Cryoplanation. In: Clark MJ, ed. *Advances in Periglacial Geomorphology.* Chichester, UK: Wiley; 1988:49-67.

Queen CW, Nelson FE, Gunn GE, Nyland KE. A characteristic periglacial landform: Automated recognition and delineation of cryoplanation terraces in eastern Beringia. *Permafrost Periglac Process.* 2020; https://doi.org/10.1002/ppp.2083.
Rampton VN. 1988. Quaternary Geology of the Tuktoyaktuk Coastlands, Northwest Territories. Ottawa; Geological Survey of Canada; Memoir 423; 1988.

Rose J. The Quaternary of the British Isles: factors forcing environmental change. J Quaternary Sci. 2010;25:399-418.

Rose J, Allen P, Kemp RA, Whiteman C, Owen N. The Early Anglian Barham Soil of Eastern England. In: Boardman J. ed. Soils and Quaternary Landscape Evolution. Chichester: Wiley; 1985:197-229.

Ruiz-Fernández J, Oliva M, Nývlt D, et al. Patterns of spatio-temporal paraglacial response in the Antarctic Peninsula region and associated ecological implications. Earth-Science Reviews. 2019;192:379-402.

Saito H, Iijima Y, Basharin NI, Fedorov AN, Kunitsky VV. Thermokarst development detected from high-definition topographic data in central Yakutia. Remote Sensing. 2018;10:1579; doi:10.3390/rs10101579

Savvinov GN, Danilov PP, Petrov AA, Makarov VS, Boeskorov VS, Grigoriev SE. Environmental problems of the Verkhoyansk district. Vestnik of North-Eastern Federal University. 2018;6(68):18-33.

Sher AV, Kaplina TN, Giterman RE, et al. Late Cenozoic of the Kolyma Lowland: XIV Pacific Science Congress, Khabarovsk August 1979, Tour Guide XI. Moscow: USSR Academy of Sciences; 1979.

Shur YL. The upper horizon of permafrost soils. In: Senneset K, ed. Proceedings of the Fifth International Conference on Permafrost, Trondheim, Norway, August 2–5, 1988. Trondheim, Norway: Tapir Publishers; 1988:867-871.

Shur Y, Hinkel KM, Nelson FE. The transient layer: implications for geocryology and climate-change science. Permafrost Periglac Process. 2005;16:5-18.
Shur Y, Kanevskiy M, Jorgenson T, Dillon M, Stephani E, Bray M. 2012. Permafrost degradation and thaw settlement under lakes in yedoma environment. In: Hinkel KM, ed. *Proceedings of the Tenth International Conference on Permafrost, Vol. 1 International contributions, June 25-29, 2012, Salekhard, Russia*. Salekhard, Russia: The Northern Publisher:383-388.

Soloviev PA. 1962. Alas relief of central Yakutia and its origin. In: Grave NA, ed. *Permafrost and Related Features in Central Yakutia*. Moscow: The Academy of Sciences of USSR Press; 1962:38-54 (in Russian).

Stoddart D. Climatic geomorphology: a review and re-assessment. *Progress in Geography*. 1969;1:159-222.

Summerfield MA, Stuart FM, Cockburn HAP, et al. Long-term rates of denudation in the Dry Valleys, Transantarctic Mountains, southern Victoria Land, Antarctica based on in-situ-produced cosmogenic 21Ne. *Geomorphology*. 1999;27:113-129.

Te Punga MT. Periglaciation in southern England. *Tijdschrift van het Koninklijk Nederlandsch Aardrijkskundig Genootschap*. 1957;74:400-412.

Thorn CE. Periglacial geomorphology: what, where, when? In: Dixon JC, and Abrahams AD, eds. *Periglacial Geomorphology*. Chichester, UK: Wiley; 1992:1-30.

Tricart J, Cailleux A. *Introduction to Climatic Geomorphology*. London: Longman; 1972. (Translation of *Introduction à la Géomorphologie Climatique*. Kiewiet de Jonge, CJ (translator). Paris: Société d’Edition d’Enseignement Supérieur; 1965).

Tumskoy VE. Peculiarities of cryolithogenesis in northern Yakutia (Middle Neopleistocene to Holocene). *Earth Cryosphere*. 2012;16(1):12-21. (in Russian)
Vadakkedath V, Zawadzki J, Przeździecki K. Multisensory satellite observations of the expansion of the Batagaika crater and succession of vegetation in its interior from 1991 to 2018. *Environ Earth Sci.* 2020;79:150.

Vaks A, Gutareva OS, Breitenbach SFM, et al. Speleothems reveal 500,000 year history of Siberian permafrost. *Science.* 2013;340:183-186.

Vaks A, Mason AJ, Breitenbach SFM, et al. Palaeoclimate evidence of vulnerable permafrost during times of low sea ice. *Nature.* 2020;577:221-225.

Vandenberghe J, Czudek T. Pleistocene cryopediments on variable terrain. *Permafrost Periglac Process.* 2008;19:71-83.

Vandenberghe J, Kasse K. Cryopedimentation on soft-sediment subsoils. *Würzburger Geographische Arbeiten* 1993;87:283-297.

Vandenberghe J, Woo M. Modern and ancient periglacial river types. *Prog Phys Geog.* 2002;26(4):479-506.

Vasil’chuk YK, Vasil’chuk AC, Rank D, Kutschera W, Kim J-C. 2001. Radiocarbon dating of δ\(^{18}\)O–δD plots in Late Pleistocene ice-wedges of the Duvanny Yar (Lower Kolyma River, Northern Yakutia). In: Carmi I, Boaretto E. eds. *Proceedings of the 17th International \(^{14}\)C Conference.* Radiocarbon. 2001;43(2B):541-553.

Veremeeva AA, Gubin SV. 2009. Modern tundra landscapes of the Kolyma Lowland and their evolution in the Holocene. *Permafrost Periglac Process.* 2009;20:399–406.

Walker HJ. Arctic deltas. *J Coastal Res.* 1998;14(3):718-738.
Walker HJ. Landform development in an arctic delta: the roles of snow, ice and permafrost. In: Hewitt K, Byrne M, English M, Young G. eds. *Landscapes in Transition*. Dordrecht, The Netherlands, Kluwer; 2002:159-183.

Walker HJ, Hudson PF. Hydrologic and geomorphic processes in the Colville River delta, Alaska. *Geomorphology*. 2003;5:291-303.

Waller RI, Murton JB, Kristensen L. Glacier–permafrost Interactions: processes, products and glaciological implications. *Sediment Geol.* 2012;255-256:1-28.

Way RG, Lewkowicz AG. Modelling the spatial distribution of permafrost in Labrador–Ungava using the temperature at the top of permafrost. *Can J Earth Sci.* 2016;53:1010-1028.

Wennrich V, Minyuk PS, Borkhodoev V, et al. Pliocene to Pleistocene climate and environmental history of Lake El'gygytgyn, Far East Russian Arctic, based on high-resolution inorganic geochemistry data. *Clim Past.* 2014;10:1381-1399.

Westgate J, Stemper BA, Péwé TL. A 3 m.y. record of Pliocene-Pleistocene loess in interior Alaska. *Geology*. 1990;18:858-861.

Westgate J, Froese DG. Stop 15: Quartz Creek: Pliocene ice wedges/Quartz Creek tephra (63° 46’ N, 139° 03’ W). In: Froese DG, Duk-Rodkin A, Bond JD, eds. *Fieldguide to Quaternary Research in central and western Yukon Territory*. Yukon Heritage Branch Occasional Papers in Earth Science 2, 2001:69-71.

Wetterich S, Rudaya N, Kuznetsov V, et al. Ice Complex formation on Bol’shoy Lyakhovsky Island (New Siberian Archipelago, East Siberian Arctic) since about 200 ka. *Quat Res.* 2019;92(2):530-548.
Wolfe SA, Kerr DE, Morse PD. Slave Geological Province: an archetype of glaciated shield terrain. In: Slaymaker O, ed. *Landscapes and Landforms of Western Canada*. World Geomorphological Landscapes. Cham, Switzerland: Springer; 2017:77–86.

Woo MK. *Permafrost Hydrology*. Berlin Heidelberg: Springer-Verlag; 2012.

Worsley P. Martin Theodore Te Punga (1921–1989) and the periglacial legacy of southern England. *Proc Geol Ass*. 2005;116:177-182.

Zachos J, Pagani M, Sloan L, Thomas E, Billups K. Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science*. 2001;292:686-693.

Zhang T, Heginbottom JA, Barry RG, Brown J. Further statistics on the distribution of permafrost and ground ice in the northern hemisphere. *Polar Geog*. 2000;24:126-131.
Figure 1. Map of permafrost zones in the Northern Hemisphere. Modified from European Environment Agency (EEA) 2017.
Figure 2. Timescale of Quaternary permafrost and periglaciation based on the benthic LR04 $\delta^{18}O$ stack for the past 3.6 Ma. Data source: Lisiecki and Raymo (2005). Blue shading indicates main periods of palaeo-ice-sheet reconstructions by Batchelor et al. (2019). Oldest known onset of past permafrost is based on Westgate and Froese (2001), Matheus et al. (2003) and Péwé et al. (2009). Oldest known permafrost containing ground ice that has never thawed, i.e., ancient permafrost, is based on Froese et al. (2008) and Murton et al. (in preparation). Indirect evidence for permafrost from the Siberian speleothem record is based on Vaks et al. (2020), and from pollen data (Brigham-Grette et al. 2013).
Figure 3. Limits of Cenozoic ice sheets and Last Glacial Maximum (LGM) permafrost in the Northern Hemisphere. Creamy white shading indicates maximum best-estimates of Plio–Pleistocene ice sheets (from Batchelor et al. 2019); pink shading indicates areas of LGM permafrost, including continuous, discontinuous, sporadic and isolated permafrost (from Lindgren et al. 2016); blue lines indicate present-day rivers (from Natural Earth 2019a); elevation and bathymetric data from Amante and Eakins (2009), and country outlines from Natural Earth (2019b). ‘IS’ denotes ‘Ice Sheet’. Map projection is North Pole Lambert Azimuthal Equal Area; central meridian: 0°; datum: D_WGS 1984.
Figure 4. Unglaciated permafrost regions beyond the maximum extent of Cenozoic glaciation showing lowland regions in which characteristic periglacial landscapes (red) are likely to occur. Dotted areas indicate regions of high (>20%) volumetric ground ice in the upper 10–20 m of ground (from Brown et al. 2002); pink shading indicates areas of LGM permafrost, including continuous, discontinuous, sporadic and isolated permafrost (from Lindgren et al. 2016) and these areas were above contemporaneous sea level during the LGM. Other sources of data as in caption of Figure 3. ‘IS’ denotes ‘Ice Sheet’. Map projection is Lambert Azimuthal Equal Area, central meridian: 180°; datum: D_WGS 1984.
Figure 5. Characteristic periglacial landscape of alas terrain, east of Yakutsk, central Yakutia, Siberia. (a). Well-developed baydzeraks (thermokarst mounds) on the far hillslope formed by partial melting of syngenetic ice wedges in the ice complex. Lake partially fills the floor of alas. (b) Pingo with high-centred polygons in floor of alas. Surrounding forest covers Pleistocene accumulation surface in which alas developed.
Figure 6. Characteristic periglacial landscape of icy badlands, Batagay megaslump, Yana Uplands, northeast Siberia. (a) Meltwater stream in gully floor. Slump headwall in distance is ~50 m high and exposes an ice complex. (b) Icy badlands in slump floor (foreground) formed by resurfacing of the original smooth hillslope (accumulation apron) covered in open woodland dominated by Cajander larch (background). Slump headwall exposes ice complex ~25 m thick. Mount Kirgilyakh in distance.
Figure 7. Polygenetic periglacial landscapes in the form of relict landsurfaces of Pleistocene accumulation plains at (a) Duvanny Yar, lower Kolyma River, northeast Siberia; and (b) Syrdah, 70 km northeast of Yakutsk, central Siberia.
Figure 8. Periglacial regions in the UK and Ireland. Region 1 was largely covered by glacier ice during the Younger Dryas Stadial, and periglacial activity within this zone has been confined to high ground during the Holocene. Region 2 was covered by the last (Late Devensian) British–Irish Ice Sheet and experienced periglacial conditions during ice-sheet recession in the Late Devensian and during the Younger Dryas. Region 3 was subject to periglacial conditions during Wolstonian deglaciation and throughout the Devensian. Region 4 experienced periglaciation during Anglian deglaciation and during the Wolstonian and Devensian. Region 5 was unglaciated and experienced recurrent periglacial conditions throughout the Quaternary. (b) Timescale for the periglacial regions during the last 500 ka. Blue shading indicates the even-numbered (cold) marine isotope stages, during which periglacial conditions in the UK tended to be coldest and most extensive. Note that periglacial and even permafrost conditions also occurred during MIS 3. Modified from Murton and Ballantyne (2017).
Figure 9. Relict periglacial landsystems of the English chalklands. (a) Ground model of limestone plateau–clay vale. (b) Dry valley incised into a plateau. (a) and (b) from Murton and Ballantyne (2017); (b) modified originally from Mortimore (2014, fig. 6.65).
Figure 10. Polygenetic periglacial landscapes marked by relict, rounded convexo–concave hillslopes (‘whale-backed landscape’) resulting from extensive modification by frost weathering of the frost-susceptible substrate and downslope transport of the rock debris by gelifluction. (a) South Downs, Sussex, underlain by chalk. (b) South Hams, south Devon, underlain by slate.
Figure 11. Polygenetic periglacial landscape dominated by cryopediments considered to be largely inactive, Buckland Hills–Barn Mountains, northern Yukon. (a) Sharp break-of-slope between the proximal (upslope) end of the cryopediments and the bedrock (creamy coloured) that crops out on the backslopes above them. (b) Cryopediments in the centre of the image are incised by tributaries of the Babbage River. Sleepy Mountain in distance.
Figure 12. Polygenetic periglacial landscapes developed on non-frost-susceptible granite bedrock and considered to be largely inactive. (a) Tors and boulder-covered hillslopes (‘clitter’), Staple Tors, near Merrivale, Dartmoor, SW England. (b) Tors and blockfield, Mount Sedgewick, Buckland Hills, northern Yukon, Canada. Arrows indicate person for scale.