

1 **Observation-based modelling of permafrost carbon fluxes**
2 **with accounting for deep carbon deposits and thermokarst**
3 **activity**

4

5 **Thomas Schneider von Deimling^{1,2}, Guido Grosse¹, Jens Strauss¹, Lutz**
6 **Schirrmeister¹, Anne Morgenstern¹, Sibyll Schaphoff², Malte Meinshausen³, Julia**
7 **Boike¹**

8 [1]{Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research, Periglacial
9 Research Unit, Potsdam, Germany}

10 [2]{Potsdam Institute for Climate Impact Research, Potsdam, Germany}

11 [3]{School of Earth Sciences, The University of Melbourne, Victoria, Australia}

12 Correspondence to: T. Schneider von Deimling (Thomas.Schneider@awi.de)

13

14 **Abstract**

15 High-latitude soils store vast amounts of perennially frozen and therefore inert organic matter.
16 With rising global temperatures and consequent permafrost degradation, a part of this carbon
17 stock will become available for microbial decay and eventual release to the atmosphere. We have
18 developed a simplified, two-dimensional multi-pool model to estimate the strength and timing of
19 future carbon dioxide (CO₂) and methane (CH₄) fluxes from newly thawed permafrost carbon
20 (i.e. carbon thawed when temperatures rise above pre-industrial levels). We have especially
21 simulated carbon release from deep deposits in Yedoma regions by describing abrupt thaw under
22 newly formed thermokarst lakes. The computational efficiency of our model allowed us to run
23 large, multi-centennial ensembles under various scenarios of future warming to express
24 uncertainty inherent to simulations of the permafrost-carbon feedback.

25 Under moderate warming of the representative concentration pathway (RCP) 2.6 scenario,
26 cumulated CO₂ fluxes from newly thawed permafrost carbon amount to 20 to 58 petagrams of
27 carbon (Pg-C) (68% range) by the year 2100 and reach 40 to 98 Pg-C in 2300. The much larger

1 permafrost degradation under strong warming (RCP8.5) results in cumulated CO₂ release of 42
2 to 141Pg-C and 157 to 313 Pg-C (68% ranges) in the years 2100 and 2300, respectively. Our
3 estimates do only consider fluxes from newly thawed permafrost but not from soils already part
4 of the seasonally thawed active layer under pre-industrial climate. Our simulated CH₄ fluxes
5 contribute a few percent to total permafrost carbon release yet they can cause up to 40% of total
6 permafrost-affected radiative forcing in the 21st century (upper 68% range). We infer largest CH₄
7 emission rates of about 50 Tg-CH₄ per year around the middle of the 21st century when simulated
8 thermokarst lake extent is at its maximum and when abrupt thaw under thermokarst lakes is
9 taken into account. CH₄ release from newly thawed carbon in wetland-affected deposits is only
10 discernible in the 22nd and 23rd century because of the absence of abrupt thaw processes. We
11 further show that release from organic matter stored in deep deposits of Yedoma regions does
12 crucially affect our simulated circumpolar CH₄ fluxes. The additional warming through the
13 release from newly thawed permafrost carbon proved only slightly dependent on the pathway of
14 anthropogenic emission and amounts to about 0.03-0.14°C (68% ranges) by end of the century.
15 The warming increased further in the 22nd and 23rd century and was most pronounced under the
16 RCP6.0 scenario with adding 0.16 to 0.39°C (68% range) to simulated global mean surface air
17 temperatures in the year 2300.

18

19

20 **1 Introduction**

21 Soils in high northern latitudes represent one of the largest reservoirs of organic carbon in the
22 terrestrial biosphere, holding an estimated 900 to 1700 petagrams (Pg) of organic carbon
23 (Hugelius et al., 2014). While portions of this carbon pool are already affected by seasonal thaw
24 in the active layer, substantial amounts are locked in perennially frozen deposits at depths
25 currently exceeding the seasonal thaw depth. Zimov et al. (2006) have estimated that an amount
26 of 450 Pg-C is stored in deep Siberian organic-rich frozen loess and have speculated that this
27 carbon stock could significantly contribute to global carbon fluxes when thawed. A more recent
28 study based on updated observations estimates a total of 211 (58 to 371) Pg-C being stored in
29 ice- and carbon-rich deep deposits in Siberia and Alaska (Strauss et al., 2013). As long as frozen
30 in the ground, permafrost organic matter is not part of the active carbon cycle and can be

1 considered mainly inert. With sustained warming and subsequent degradation of deeper
2 permafrost deposits, a part of this carbon pool will become seasonally thawed. Consequently, it
3 will become prone to microbial decomposition and mineralization. By ultimately increasing the
4 atmospheric concentration of the greenhouse gases CO₂ and CH₄, the carbon release from
5 thawing permafrost regions is considered a potentially large positive feedback in the climate-
6 carbon system (Schaefer et al., 2014, Schuur et al., 2015). Given the long millennial timescale
7 processes leading to the build-up of old carbon in permafrost soils, future rapid releases from
8 these deposits are irreversible on a human timescale.

9 However, the magnitude and timing of carbon fluxes as a consequence of permafrost degradation
10 are highly uncertain. This is mainly due to incomplete observational knowledge of the amount of
11 organic matter stored in permafrost deposits, of its quality and decomposability, as well as due to
12 the challenge of modelling the full chain of processes from permafrost thaw to carbon release.
13 Furthermore, conceptual and numerical permafrost landscape models also require suitable
14 upscaling methods ranging from local to global scales, based on field-based knowledge of the
15 surface characteristics, key processes and data collection of key parameters (Boike et al., 2012).
16 The vulnerability of permafrost carbon and its fate when thawed will be strongly determined by
17 various environmental controls (Grosse et al., 2011) such as soil type and soil moisture, which
18 both affect soil thermal conductivity and therefore determine the timescale of heat penetration
19 into the ground. Additionally, surface conditions such as organic-rich soil surface layers,
20 vegetation cover and snow exert strong controls on subsurface temperatures by insulating the
21 ground from surface air temperatures (Koven et al., 2013a). In the absence of conditions for
22 abrupt permafrost thaw, mineral permafrost soils are typically more vulnerable to degradation
23 than carbon-rich organic soils. The difference in vulnerability results from the insulating
24 properties of thick organic layers which slow down permafrost degradation (Wisser et al. 2011).
25 Further, the often higher ice-content of organic as compared to mineral soils requires a larger
26 energy input for phase transition, and the usually anaerobic environments in organic soils slow
27 down carbon mineralization. Yet, organic soils which are prone to ground subsidence and
28 impoundment can be highly vulnerable and thus reveal permafrost degradation at increased rates
29 (e.g. Camill et al., 2005; Johnson et al., 2013).

30 Therefore, for capturing site-specific pathways of carbon release from permafrost degradation, it
31 is important to consider the differing soil environments under which the organic matter will be

1 thawed. Of key importance is the impact of hydrological and redox conditions which determine
2 whether mineralized carbon will be emitted as CO₂ or CH₄ (Olefeldt et al., 2013). Future changes
3 in hydrological conditions in permafrost regions will therefore crucially affect the high latitude
4 carbon balance. Especially regions of ice-rich late Pleistocene deposits (Yedoma) are considered
5 to become potential hot spots for intensive thermokarst lake formation with consequent increases
6 in the fraction of permafrost-affected sediments under anaerobic environments (Walter et al.,
7 2007a). Apart from affecting hydrological conditions, thermokarst lakes also exert a strong
8 warming of sub-lake sediments and thus enhance abrupt permafrost degradation. If thermokarst
9 lake depths exceed the maximum thickness of winter lake ice, these lakes retain liquid water
10 year-round and provide a strong warming and thawing of the underlying sediments (Arp et al.,
11 2012). As a consequence, mean annual temperatures of thermokarst lake-bottom sediments can
12 be up to 10 °C warmer than mean annual air temperatures (Jorgenson et al., 2010).

13 So far, permafrost carbon dynamics are not included into standard climate model projections,
14 possibly due to only recent recognition of the large vulnerable permafrost carbon pool and given
15 the complexity of processes involved. The complexity arises not only from the need to simulate
16 physical changes in soil thermal conditions and phase transitions of water as a consequence of
17 various environmental controls (e.g. interactions among topography, water, soil, vegetation and
18 snow (Jorgenson et al., 2010)). It also arises from the challenge of describing the full chain of
19 biogeochemical processes for eventual carbon decomposition in the soils and release to the
20 atmosphere. Therefore, various aspects of permafrost physics and biogeochemistry are only
21 recently being implemented into current global climate models (formulated e.g. in Lawrence and
22 Slater, 2008; Koven et al., 2009; Lawrence et al., 2011; Dankers et al., 2011; Schaphoff et al.,
23 2013; Koven et al., 2013b; Ekici et al., 2014). First modelling results suggest a very large range
24 in predicted soil carbon losses from permafrost regions under scenarios of unmitigated climate
25 change (about 20 to 500 Pg-C by 2100, see Schaefer et al. (2014) for an overview). This large
26 range demonstrates the current uncertainty inherent to predictions of the timing and strength of
27 the permafrost carbon feedback.

28 Yet, these studies are based on models which still miss important mechanisms to capture the full
29 complexity of the permafrost carbon feedback. Grosse et al. (2011) and van Huissteden and
30 Dolman (2012) note that none of the current permafrost models consider the spatially
31 inhomogeneous and potentially rapid degradation of ice-rich permafrost by thermokarst lake

1 formation. This omission of abrupt thaw processes may result in underestimating an important
2 part of anaerobic soil carbon decomposition. Studies have also underlined the importance of
3 considering small scales: not only large Arctic lakes, but also the smaller Arctic thaw ponds, are
4 biological hotspots for the emission of CO₂ and CH₄ (Abnizova et al., 2012; Laurion, 2010). A
5 recent expert assessment has emphasized the importance of abrupt thaw processes and so far
6 unaccounted carbon stored in deep deposits below three meters (Schuur et al., 2013). Evidence
7 for rapid and abrupt thaw on decadal scale is already widespread (Jorgenson et al., 2006; Sannel
8 and Kuhry, 2011; Kokelj et al., 2013; Raynolds et al., 2014), is likely to increase with future
9 warming, and thus needs to be considered to make realistic projections of carbon dynamics in
10 permafrost regions.

11 Our study aims to estimate the range of potential carbon fluxes from thawing permafrost by
12 accounting for some abrupt thaw processes which can accelerate the degradation of frozen
13 ground beyond what is inferred by standard modelling approaches that consider gradual thaw. By
14 allocating permafrost organic matter into pools governed by different environmental controls, we
15 describe different pathways of carbon release and we especially account for carbon released as
16 CH₄. By explicitly modelling carbon releases from deep carbon stores below three meters, we
17 contribute to a more complete quantification of the permafrost-carbon feedback. Permafrost
18 carbon release from deep deposits has mostly not been taken into account previously, although
19 first-order modelling studies have considered the contribution of permafrost carbon in Yedoma
20 regions (Koven et al., 2011; Schaphoff et al., 2013). Yet in these studies the deep deposits have
21 not contributed significantly to simulated carbon release because the models did not describe
22 abrupt thaw processes which may affect great depths. Khvorostyanov et al. (2008) have inferred
23 a large contribution from Yedoma carbon deposits after the year 2300 when assuming that
24 microbial heat strongly speeds-up permafrost degradation. To the best of our knowledge, our
25 modelling approach is the first to globally quantify the permafrost-carbon feedback for the
26 coming centuries under considering carbon release from deep deposits and accounting for abrupt
27 thaw processes.

28

1 **2 Multi-pool permafrost model**

2 Building on previous work (Schneider von Deimling et al., 2012), we have developed a
3 simplified large-scale two-dimensional model with parameters tuned to match observed
4 permafrost carbon characteristics. The model calculates permafrost degradation and eventual
5 CO_2 and CH_4 release under differing environmental conditions. The newly developed model is
6 shortly described in the following sections while more details are given in the supplementary
7 material.

8 The model accounts for several processes which are crucial to the permafrost carbon feedback:

9 1. Depending on soil-physical factors, hydrologic conditions, and organic matter quality,
10 permafrost carbon inventories were sub-divided into a total of 24 pools.

11 2. Permafrost thaw was calculated for various scenarios of global warming to determine the
12 amount of carbon vulnerable to eventual release. Anaerobic soil fractions were
13 calculated to determine the amount of organic matter stored in wetland- and
14 thermokarst-affected sediments.

15 3. Permafrost carbon release as either CO_2 or CH_4 was calculated based on typical rates for
16 aerobic and anaerobic carbon release.

17 4. By using a simplified climate-carbon model, we have determined the additional increase
18 in global mean temperature through the permafrost carbon feedback.

19 The computational efficiency of our model allows us to explore the range of simulated
20 permafrost carbon feedbacks by running large ensembles. Our analysis expresses the uncertainty
21 inherent to current knowledge of permafrost carbon release. Our framework allows identification
22 of key model parameters and processes and thus enables us to assess the importance of these
23 factors in affecting the strength and timing of the permafrost carbon feedback.

24 **2.1 Model structure**

25 The magnitude and timing of carbon release from thawing permafrost soils will be strongly
26 determined by soil-physical factors such as soil texture and organic matter decomposability,
27 hydrologic state, and surface conditions. To account for these factors, we have developed a
28 simplified but observationally constrained and computationally efficient two-dimensional model

1 which allocates permafrost soil organic matter into various carbon pools. These pools describe
2 carbon amount and quality, soil environments, and hydrological conditions (Fig. 1). To account
3 for deposit-specific permafrost carbon vulnerability, we divide our carbon inventory into two
4 near-surface pools (mineral and organic, 0 to 3m) and into two deep-ranging pools (Yedoma and
5 refrozen thermokarst (including taberl sediments), 0 to 15m, see next section and Table 1). We
6 allocate soil carbon contents according to the inventory estimates of the Northern Circumpolar
7 Soil Carbon Database (Hugelius et al., 2013). Hereby, we describe our mineral soil pool by the
8 sum of SOC contents from orthels and turbels, and our organic pool by the SOC content from
9 histels (see supplementary material for details and for soil classification definitions). We define
10 taberl deposits as permafrost sediments that underwent thawing in a talik (a layer of year-round
11 unfrozen ground in permafrost areas, such as under a deep lake), resulting in diagenetic alteration
12 of sediment structures (loss of original cryostructure, sediment compaction) and biogeochemical
13 characteristics (depletion of organic carbon). In addition, taberl deposits may be subject to
14 refreezing (e.g., after lake drainage) (Grosse et al., 2007).

15 We describe differing hydrological controls by further subdividing each carbon pool into one
16 aerobic fraction and two anaerobic fractions. Hereby we account for anaerobic conditions
17 provided in wetland soils and by water-saturated sediments under thermokarst lakes. We put our
18 model focus on the formation of new thermokarst lakes. We do not consider the contribution of
19 lake areas which existed already under pre-industrial climate. The scarcity of observational data
20 hampers an estimate of circumpolar lake ages. Therefore, estimates of the fraction of sub-lake
21 sediments, which were thawed by past talik formation and growth, are highly uncertain.

22 In the following we define wetland soils from a purely hydrological viewpoint, i.e. by assuming
23 that these soils are water-saturated and not affected by thermokarst. We further assume that
24 anaerobic soil fractions are not stationary but will increase or decrease with climate change.

25 Therefore, we re-calculate the wetland and thermokarst fraction for each time step (see
26 supplementary material for model details). Given the large-scale dominance of aerobic over
27 anaerobic Arctic landscapes, we assume that wetland or thermokarst affected soils can be seen as
28 isolated patches surrounded by aerobic soils. Our model describes the expansion of wetlands or
29 thermokarst lakes into aerobic environments by an increase in the anaerobic area fractions at the
30 expense of the aerobic area fraction. Vice versa, a decrease in simulated anaerobic fractions

1 describes the falling dry of previously water-saturated ground and leads to an increase in the
2 aerobic fraction. We do not consider the case of a thermokarst lake which develops into a
3 wetland by terrestrialization. We neither consider the reverse case of a wetland becoming a
4 thermokarst-affected terrain.

5 The magnitude of fractional area changes determines the amount of carbon which gets
6 transferred between the aerobic and anaerobic pools. Carbon transferred is then subject to
7 environmental control of thaw and decomposition of the corresponding new pool.

8 We assume a linear increase in wetland extent with global warming with mean maximum
9 increases up to 30% above pre-industrial wetland extent (see Table 1). We stress that future
10 changes in wetland extent are subject to large uncertainty. While e.g. Gao et al. (2013)
11 investigate future CH₄ release from Arctic regions based on simulating future increases in
12 saturated areas, Avis et al. (2011) consider a scenario of a reduction in future areal extent and
13 duration of high-latitude wetlands.

14 To capture the growth and decline of newly formed thermokarst lakes, we have developed a
15 conceptual model by making the simplifying assumption that future increases in high latitude
16 surface air temperatures are the main driver for thermokarst formation. We hereby assume that
17 future warming results in a gradual increase in newly formed thermokarst lake areas (Smith et
18 al., 2005; Plug and West, 2009; Walter et al., 2007b) until a maximum extent is reached (see
19 Table 1). With further warming our model describes a decrease in thermokarst lake extent as we
20 assume that lake drainage is becoming a key factor which strongly limits thermokarst lake area
21 (van Huissteden et al., 2011; Smith et al., 2005; Jones et al., 2011; Morgenstern et al., 2011); see
22 also supplementary Fig. S1).

23 As the quality of organic matter is a further key determinant for the timescale of carbon release
24 (Strauss et al., 2015) we subdivide the carbon of each individual pool into a fast and a slowly
25 decomposing fraction, with annual or respectively decadal timescales (Table 1). We do not
26 describe permafrost organic matter of low quality (passive pool) which decays on a multi-
27 centennial to millennial timescale. The partitioning of permafrost organic matter results in a total
28 of 24 separate carbon pools which all contribute individually to simulated carbon fluxes (Fig. 1).

29 All pools and processes are stratified along latitudinal bands that provide a simplified gradient of
30 climate and permafrost types. To describe the climate control exerted by surface-air and ground

1 temperatures in each latitudinal band, we assume that large-scale climate effects can be described
2 by a general north-south temperature gradient. We acknowledge that longitudinal patterns can
3 also be pronounced, but with a focus on large-scale regional rather than local changes we expect
4 that the dominant climate control can be described by a profile of coldest permafrost
5 temperatures at the northern limit and warmest temperatures at the southern limit (Romanovsky
6 et al., 2010; Beer et al., 2013). Our model also resolves vertical information to account for
7 varying carbon density with depth and to track active layer changes (see section 2.2). We chose a
8 model resolution of 20 latitudinal bands (which range from 45°N to 85°N with a 2° gridding) and
9 of 27 vertical soil layers (corresponding to layer thicknesses of 25cm for the upper four meters,
10 and of 1m for the depth range 4 to 15m).

11 **2.2 Model initialization**

12 The flexibility of our model allows us to tune model parameters to observed data, e.g. to
13 permafrost carbon inventories, carbon qualities, or active layer depths. This approach assures
14 that our simulations do not suffer from an initial bias in the amount of modelled permafrost
15 carbon. This is contrary to model studies, which fully simulate soil thermal conditions with
16 potentially large biases in initial permafrost extent (Slater and Lawrence, 2013). Such biases
17 result in a large spread in simulated initial permafrost carbon stocks (Mishra et al., 2013;
18 Gouttevin et al., 2012). Based on updated Arctic soil carbon data (Hugelius et al., 2013; Hugelius
19 et al., 2014; Strauss et al., 2013; Walter Anthony et al., 2014) we allocate permafrost carbon
20 pools (latitudinally and vertically resolved) in different regions: two deep-ranging pools (0 to
21 15m) in regions with Yedoma (80 Pg-C) and refrozen thermokarst deposits (240 Pg-C), and two
22 near-surface pools (0 to 3m) in remaining regions with mineral soils (540 Pg-C) and organic
23 soils (120 Pg-C), see the supplementary material and Table 1. We describe the vertical soil
24 carbon distribution separately for each meter of near-surface permafrost based on the Northern
25 Circumpolar Soil Carbon Database (Hugelius et al., 2013). For deep soils below three meters we
26 assume a constant vertical carbon density (see Strauss et al., 2013, Strauss et al., 2015).

27 We then initialize each latitudinal band with a mean annual ground temperature between -0.5°C
28 and -10°C based on summer air temperature climatology data from the Berkeley Earth dataset
29 (<http://berkeleyearth.org/data>; see supplementary material). The above temperature range is
30 consistent with observed ground temperatures of continuous and discontinuous permafrost in the

1 northern hemisphere (Romanovsky et al., 2010). We do not consider permafrost temperatures
2 below -10°C (observed in the Canadian Archipelago and Northern Russia) which we consider in
3 the outer tail of permafrost temperature distributions.

4 By assuming that the equilibrium active layer depth is determined by mean annual ground
5 temperature and by the seasonal cycle of soil temperatures (see Koven et al., 2013a), we
6 calculate typical minimum seasonal thaw depths of about 30cm (northernmost permafrost
7 regions) and maximum seasonal thaw of about 250 to 300cm (southernmost regions) for present-
8 day climate conditions (see supplementary material). Although topography, soil type, as well as
9 organic layer, vegetation cover, and snow cover variability can lead to spatially very
10 heterogeneous patterns of active layer thicknesses, our scheme describes a latitudinal tendency of
11 a strong north-south gradient of both subsoil temperature and active-layer thickness that
12 generally matches observations (Beer et al., 2013).

13 By calculating the active layer depth for each carbon pool and in each latitudinal band, we can
14 determine the fraction of permafrost carbon below the active layer and therefore the amount of
15 organic matter perennially frozen under our baseline climate conditions (i.e. pre-industrial
16 climate). Large amounts of organic matter in permafrost soils reside in the active layer and were
17 affected by past decomposition and release over millennia. It is unclear to what extent the quality
18 of this seasonally-thawed organic material will allow extensive microbial decay in the future.
19 Therefore we follow a strategy similar to Burke et al. (2012) and Harden et al. (2012) of
20 considering only the part of permafrost carbon which was locked in perennially frozen ground
21 since pre-industrial times and thus was not part of the active carbon cycle for millennia. We
22 hereby assume that our carbon inventory describes organic matter in continuous and
23 discontinuous permafrost. This carbon is likely to represent organic matter perennially frozen
24 since pre-industrial climate. We do not consider soil carbon stored in younger permafrost
25 deposits (sporadic and isolated patches) which likely had been thawed for the majority of the
26 Holocene and therefore is likely depleted in labile organic matter. When accounting for
27 uncertainty in model parameters, we infer a range of about 400 to 1100 Pg of carbon perennially
28 frozen under pre-industrial climate. By combining field information with modelling, Harden et
29 al. (2012) have estimated a total of about 130 to 1060 Pg of carbon perennially frozen under
30 present day climate.

1 Further, we account for the fact that a large part of the permafrost carbon inventory (i.e. the
2 passive pool) will likely be recalcitrant to decay on a multi-centennial timescale (Schmidt et al.,
3 2011). Assuming a passive pool fraction of about 40 to 70%, only about 120 to 660 Pg of
4 permafrost carbon can become vulnerable for eventual carbon release in our simulation setting.

5 To capture uncertainty in modelled carbon fluxes from thawing permafrost deposits, we have
6 independently sampled a set of 18 key model parameters which are subject to either
7 observational or to model description uncertainty. For each warming scenario, we have
8 performed 500 ensemble runs by applying a statistical Monte-Carlo sampling and by assuming
9 uniformly and independently distributed model parameters and initial values.

10 **2.3 Permafrost thaw and carbon release**

11 With increasing high latitude warming the active layer will deepen. We model this process by
12 assuming that climate-driven long-term thaw rates can be described depending on four key
13 factors: physical ground properties, mean annual ground temperatures, depth of the thawed
14 sediment layer, and magnitude of the warming anomaly which drives permafrost degradation
15 (see supplementary material). Hereby we capture factors which strongly affect pool-specific
16 thaw dynamics, e.g. talik formation under thermokarst lakes, dampening of the thaw signal with
17 depth, variable soil-ice contents. We therefore can determine the amount of newly thawed
18 organic matter under various anthropogenic emission scenarios as a consequence of warming
19 above pre-industrial temperatures. We hereby assume carbon emissions proportional to the
20 amount of newly thawed carbon in each pool. Eventual carbon emission as CO_2 or CH_4 is
21 determined through calculated aerobic and anaerobic emission rates (see supplementary
22 material).

23 Finally, the permafrost model was coupled to a simple multi-pool climate-carbon cycle model to
24 close the feedback loop: while the permafrost model simulates permafrost degradation and
25 subsequent carbon release (as CO_2 and CH_4), the climate carbon-cycle model calculates
26 atmospheric changes in CO_2 and CH_4 concentrations and subsequent increases in global mean
27 surface air temperatures. Based on state-of-the-art climate models (CMIP-5, Taylor et al., 2011),
28 we infer polar amplification factors to describe surface air warming in each latitudinal band
29 which then drives permafrost degradation in the next time step.

1 **2.4 Model limitations**

2 Our approach of modelling permafrost thaw relies on the simplifying assumption that the main
3 driver of permafrost degradation is the rise of Arctic air temperatures. Yet soil thermal
4 conditions can be influenced by factors other than temperature (e.g. vegetation cover, snow
5 thickness, topography) (Jafarov et al., 2012; Jorgenson et al., 2010). We motivate our modelling
6 approach by focusing on the large-scale and long-term deepening of active layer thickness under
7 various warming scenarios. Although snow cover is considered a key factor for simulating
8 present day permafrost extent consistent with observations (Koven et al., 2013a; Langer et al.,
9 2013; Osterkamp, 2007; Stieglitz et al., 2003), it is unclear how strongly future changes in high-
10 latitude snow cover will affect permafrost degradation. Given that no high-quality data products
11 are available for a circumarctic mapping of snow cover, snow depth, and snow density – and
12 given that climate models simulate strongly divergent pathways of future snowfall – we here
13 make the simplifying assumption that the long term evolution of permafrost is largely driven by
14 changes in surface air temperatures. Similarly, our simplified approach of describing thermokarst
15 dynamics is based on the assumption that future thermokarst formation is largely affected by
16 increasing surface air temperature. Temperature-unrelated, local factors (such as topography,
17 precipitation changes or wildfire) can also be key determinants for thermokarst dynamics. We
18 understand our approach mainly as quantifying carbon fluxes under different hypotheses of
19 future thermokarst development rather than providing deterministic and explicit predictions of
20 individual thermokarst terrains. An alternative scenario of a reduction in high-latitude inland
21 water surface area under future warming was e.g. investigated by Krinner and Boike (2010).

22 Nutrient limitation in the soils and abrupt carbon release after wildfires are considered two
23 additional and potentially important mechanisms for the carbon balance of thawed permafrost
24 deposits which we do not consider in our model design (Koven et al., 2015; Mack et al., 2004;
25 Turetsky et al., 2011). Probably the largest effect of unaccounted processes on our simulated
26 carbon fluxes comes from the omission of high latitude vegetation dynamics. Increased carbon
27 uptake in a warmer climate through more productive vegetation can strongly affect the Arctic
28 carbon balance (Schaphoff et al., 2013). The capturing of this feedback component requires the
29 implementation of a dynamic vegetation model which is beyond the scope of this study. Also of
30 importance in this respect is the potential restoration of carbon sinks after lake drainage which

1 could, on the long-term, partially compensate for high CH₄ emission (van Huissteden and
2 Dolman, 2012; Kessler et al., 2012; Jones et al., 2012; Walter Anthony et al., 2012).

3 Our simulated wetland CH₄ fluxes describe CH₄ produced from newly thawed permafrost
4 carbon. Yet the full carbon balance of wetlands is rather complex and possibly more affected by
5 future changes in soil moisture, soil temperature, and vegetation composition than by the
6 delivery of newly thawed organic matter through permafrost degradation (Olefeldt et al., 2013).
7 The accounting of these additional factors requires the implementation of comprehensive
8 wetland models (such as suggested by Frolking et al. (2001); Kleinen et al. (2012); Eliseev et al.
9 (2008)).

10

11 **3 Model results**

12 **3.1 Permafrost degradation**

13 We have run our model under various scenarios of future warming, ranging from moderate
14 (RCP2.6) to extensive (RCP8.5). Under RCP2.6, global greenhouse gas emissions peak by 2020
15 and decline strongly afterwards. We simulate subsequent increases in global mean surface-air
16 temperatures which are constrained to below two degrees above pre-industrial levels. In case of
17 unmitigated climate change (RCP8.5), global mean surface air temperatures continuously
18 increase and reach 10°C by the end of the 23rd century at the upper range of our simulations. This
19 pronounced difference in simulated surface air temperatures results in strongly differing
20 pathways of long-term permafrost degradation (Fig. 2).

21 Depending on initial mean annual ground temperatures (MAGT_{t0}), we infer for cold
22 (MAGT_{t0}=-10°C), medium (MAGT_{t0}=-5°C), and warm (MAGT_{t0}=-0.5°C) permafrost mean
23 active layer depths of 20cm, 70cm, and 250cm, respectively. In a recent study, Koven et al.
24 (2013a) have diagnosed observed active layer depths north of 55°N from a circumpolar and a
25 Russian data set (CALM (Brown et al., 2000), and Zhang et al. (2006)). Their analysis suggests a
26 range of measured present-day active layer depths ranging from 30 to 230cm. The authors
27 underline the challenge of comparing modelled with observed active layer depths given the
28 different spatial coverage of models and observations.

1 As projections of surface air temperatures only start to diverge strongly after the middle of the
2 21st century, continuous but slow deepening of the active layer is similar under RCP2.6 and
3 RCP8.5 until 2050 (Fig. 2). We first focus on active layer deepening of the largest pool of
4 permafrost carbon, i.e. organic matter in mineral soils under aerobic conditions (Fig. 2, upper
5 panels). Under moderate warming (RCP2.6), active layer depths stabilize after 2100 for cold and
6 medium permafrost temperatures (blue and green curves). Permafrost in southerly warm regions
7 will degrade in our simulations with disappearance of near-surface (0 to 3m) permafrost before
8 2100 (red curve). Under strong warming (RCP8.5), a sharp increase in thawing rates in the
9 second half of the 21st century can be seen and the majority of model runs suggest a degradation
10 of near-surface permafrost towards the end of the century. In northern and cold permafrost
11 regions, a complete disappearance of near-surface permafrost is only realized after 2150 (blue
12 curve, upper right panel). The sustained long-term warming leads to a continuous deepening of
13 the permafrost table which can reach about 10m (~7 to 13m, 68% range) by the year 2300 in our
14 simulations.

15 Under wetland conditions (i.e. water/ice-saturated sediments), the active layer shows a similar
16 but slower deepening in response to rising surface air temperatures (Fig. 2, mid panels). In
17 contrast, when considering thermokarst lake formation, thaw rates increase sharply (Fig. 2, lower
18 panels) once lakes have reached a critical depth which prevents winter refreeze. As we do not
19 model lake depth expansion we assume that formation of new thermokarst lakes is initiated for
20 any warming above pre-industrial climate, while we assume that critical lake depths are only
21 realized with beginning of the 21st century (see supplementary material). In the first years after
22 intense thermokarst formation, sub-lake talik progression is very pronounced and annual thaw
23 rates amount many decimetres – in line with observational and modelling studies (Ling et al.,
24 2012; Kessler et al., 2012). The abrupt thaw dynamics results in disappearance of near-surface
25 permafrost well before 2050 (Fig. 2, lower panels). By the year 2100, typical talik depths amount
26 to 10 to 15m. The evolution of active layer depths in thermokarst-affected deposits does not
27 strongly differ between moderate and extensive warming (Fig. 2, lower panels). This is because
28 the degradation in thermokarst-affected sediments is driven by lake-bottom temperatures.
29 Averaged over a full year, lake-bottom temperatures do not strongly differ between moderate and
30 strong surface-air warming (see also Boike et al. (2015) and supplementary material).

1 In our model setting, we explicitly account for permafrost carbon in deep inventories (Yedoma
2 and refrozen thermokarst deposits). By the end of the 23rd century, typical depths of the
3 permafrost table in these carbon- and ice-rich sediments reach about 5 to 9m under the RCP8.5
4 scenario if no abrupt thaw is considered (not shown). Thus even under strong surface air
5 warming, our simulations suggest a large part of the deep carbon deposits will remain
6 perennially frozen over the coming centuries if only gradual thaw is considered. In contrast, in
7 most latitudes where ice-rich Yedoma is affected by new thermokarst formation, thaw reaches
8 the maximum model depth of 15m before 2300.

9 **3.2 Permafrost carbon release**

10 We define permafrost carbon fluxes similar to Burke et al. (2012) and Harden et al. (2012) as the
11 release from newly thawed permafrost carbon, i.e. the contribution of perennially frozen soil
12 organic matter which becomes part of the active carbon cycle if warmed above pre-industrial
13 temperatures. We stress that these fluxes do not describe the full carbon balance of permafrost
14 regions which is also affected by changes in vegetation uptake, new carbon inputs into deeper
15 soil layers and carbon release from soil surface layers which were already seasonally thawed
16 under pre-industrial climate (see discussion in section 2.2).

17 Depending on the degree of ground warming and thus on the extent of active layer deepening,
18 differing amounts of newly thawed carbon will be made available for microbial decomposition
19 and eventual release to the atmosphere. Fig. 3 illustrates permafrost carbon thaw and emissions
20 under a scenario of moderate warming (RCP2.6, upper panels) and extensive warming (RCP8.5,
21 lower panels). Under RCP2.6, largest increases in newly thawed permafrost carbon (Fig. 3, first
22 column) are realized until the middle of the 21st century with a total of 167 Pg-C (113 to 239 Pg-
23 C, 68% range) of which 40 to 70% is assumed part of the passive carbon pool and thus
24 recalcitrant on the timescale considered here. In contrast, the pronounced and continuous
25 warming under RCP8.5 results in much larger amounts of newly thawed permafrost carbon. By
26 the year 2100, 367 Pg-C are thawed (233 to 497 Pg-C, 68% range), and through further
27 permafrost degradation in the 22nd and 23rd century, a total of 564 Pg-C (392 to 734 Pg-C, 68%
28 range) of organic matter is newly thawed by the year 2300. Focusing on the top three soil meters
29 and considering a larger uncertainty spread in the permafrost carbon inventory, two recent

1 studies estimated a min-max range of 75 to 870 Pg (Burke et al. 2012) and of 105 to 851 Pg
2 (Harden et al. 2012) of newly thawed permafrost carbon under RCP8.5 until the year 2100.

3 The intensity of carbon release after permafrost thaw differs strongly among the scenarios in our
4 simulations (Fig. 3). While under RCP2.6, maximum annual CO₂ emission rates are constrained
5 to about 0.4 Pg-C/yr (0.2 to 0.6 Pg-C/yr, 68% range), peak emission rates under RCP8.5 amount
6 to 1.7 Pg-C/yr (median) and can reach 2.6 Pg-C/yr (upper 68% range). The decline in emission
7 rates in the 22nd and 23rd century describes the depletion of thawed permafrost carbon through
8 release to the atmosphere. Under all RCPs, peak CO₂ emission rates occur around the end of the
9 21st century.

10 Due to much lower anaerobic CH₄ as compared to aerobic CO₂ production rates (Table 1), and
11 due to the majority of soil carbon being thawed under aerobic conditions, emission from thawing
12 permafrost soils amounts to only a few percent of total permafrost carbon release. Observational
13 and modelling experts have estimated that CH₄ will contribute by about 1.5 to 3.5% to future
14 permafrost carbon release (Schuur et al., 2013).

15 Given the slow progression of permafrost thaw in wetland-affected sediments, CH₄ release from
16 newly thawed permafrost carbon is only discernible after end of this century (Fig. 3). We
17 consider our estimates of wetland carbon fluxes being conservative: we neither account for
18 carbon release from organic matter contained in the active layer which is already thawed since
19 pre-industrial times, nor do we account for enhanced thaw of water-saturated grounds affected by
20 non-conductive heat flow.

21 Our simulations suggest maximum annual CH₄ emission rates of a few Tg-CH₄ for moderate
22 warming, about 16 Tg-CH₄ (8 to 28 Tg-CH₄, 68% range) for strong warming. To the contrary,
23 abrupt thaw under thermokarst lakes results in peak CH₄ emission after the middle of this
24 century. Under RCP2.6, maximum annual CH₄ emissions are constrained to about 5.5 Tg-CH₄
25 (up to 11.5 Tg-CH₄ for the upper 68% range), while under RCP8.5 peak CH₄ emission reach
26 about 26 Tg-CH₄ (14 to 49 Tg-CH₄, 68% range). The strong decline in emission rates towards
27 the end of the century is an expression of the sharp decrease in thermokarst lake extents through
28 increasing drainage under sustained warming (see Fig. S1).

1 Under strong warming, our modelled CH_4 emissions accumulate to 836 to 2614 Tg- CH_4 (68%
2 range) until the year 2100. Maximum contributions until the year 2300 can reach 10.000 Tg- CH_4
3 (upper 68% range, see Table 2).

4 We have additionally analysed the impact of uncertainty in initial MAGT distribution on the
5 calculated carbon fluxes. Soil temperatures affect the magnitude of carbon release in two ways.
6 First, MAGTs determine the initial active layer profile and thus the amount of carbon perennially
7 frozen under pre-industrial climate. Second, soil temperatures determine the vulnerability of
8 permafrost carbon to future degradation. Based on a model ensemble with sampling solely
9 uncertainty in MAGT, we inferred a spread in the year 2100 of $32.5 \pm 23\%$ Pg-C and $81.5 \pm 8\%$
10 Pg-C for the scenarios RCP2.6 and RCP8.5 respectively, which further increase to $60 \pm 33\%$ Pg-C
11 and $235 \pm 6\%$ Pg-C in the year 2300. The factor 3-5 larger fractional uncertainty for the climate
12 mitigation scenario (RCP2.6) illustrates the enhanced sensitivity to initial permafrost
13 temperatures of modelled carbon fluxes under moderate warming.

14 **3.3 Contribution of individual soil pools and of deep deposits**

15 Carbon release discussed in the previous section describes the sum of fluxes over all individual
16 soil types, hydrologic controls, and organic matter qualities (based on a total of 24 individual
17 carbon pools, see section 2.1). We illustrate the contribution of individual fluxes to the total
18 carbon budget in supplementary figures S2 and S3. It can be seen that CO_2 fluxes are largely
19 controlled by contributions from mineral soils, as these soils describe the largest source of
20 organic matter and as they are dominated by aerobic conditions (Fig. S2). In contrast, the total
21 CH_4 balance is influenced by contributions from all soils types. In our simulation setting, 21st
22 century CH_4 fluxes are largely controlled by the formation and expansion of new thermokarst
23 lakes, while discernible CH_4 release from newly thawing permafrost in wetlands results only in
24 the 22nd and 23rd century.

25 We account for a total of 230 Pg of organic carbon buried below three meters in Yedoma and
26 refrozen thermokarst deposits (including talberal sediments). Under aerobic or wetland
27 conditions, our simulations suggest only small contributions of these deep deposits to the total
28 release of newly thawed permafrost carbon even under scenarios of strong warming (Fig. 4,
29 supplementary figures S2 and S3). Discernible contributions are only inferred towards the end of

1 our simulations (23rd century), with fluxes from deep deposits contributing a maximum of about
2 10% to accumulated CO₂ release or about 5% to total wetland CH₄ release (upper 68% ranges).
3 The lagged response of deep carbon release is an expression of the slow penetration of heat into
4 the ground. In most latitude bands under the RCP2.6 scenario, no frozen carbon from deep
5 deposits is thawed as the moderate warming does not result in active layer depths exceeding
6 three meters.

7 Yet if abrupt thaw under thermokarst lakes is accounted for, the fast propagation of sub-lake
8 taliks can unlock large amounts of perennially frozen deep organic matter even within this
9 century (see supplementary figures S2 and S3). Our simulations suggest that until 2100 about 25
10 to 30% of emitted CH₄ from thermokarst lakes stems from contributions of deep permafrost
11 carbon (Fig. 4, lower panel). Maximum contributions until 2300 can amount to 35% (upper 68%
12 range).

13 We have performed additional model simulations to illustrate the extent to which our simulated
14 permafrost carbon fluxes are affected by changes in anaerobic soil fractions and by deep carbon
15 release. For this purpose we have run two further model ensembles under identical parameter
16 settings for each warming scenario in which we 1) fixed anaerobic soil fractions at initial values
17 (i.e. static anaerobic soil fractions), and 2) disregarded soil carbon below 3 meters. Resulting
18 CO₂ fluxes reveal a comparable magnitude under the different simulation settings because our
19 simulated changes in anaerobic soil fractions and contributions from deep carbon deposits do
20 only slightly affect total CO₂ release. Yet these factors were found to exert a strong control on
21 simulated CH₄ release (supplementary figure S4). Especially CH₄ release in the 21st century is
22 largely driven by the contribution from newly formed thermokarst lakes, enhanced by carbon
23 release from deep deposits.

24

25 **3.4 Permafrost-affected warming**

26 To disentangle the warming caused by anthropogenic greenhouse gas emission from warming
27 caused by permafrost-carbon release, we have performed paired-simulations under identical
28 parameter settings – once with the permafrost module activated and once deactivated. The
29 difference in global mean surface-air temperatures between each pair of ensemble simulations is

1 what we define as the additional global warming caused by newly thawed permafrost carbon (i.e.
2 permafrost-affected warming).

3 Although permafrost carbon release increases strongly with rising global temperatures (Fig. 3),
4 our results suggest a permafrost-affected global warming of about 0.05 to 0.15°C (68% range)
5 until 2100 which is only slightly dependent on the anthropogenic emission pathway. (Fig. 5,
6 Table 2). The quasi path-independency of the permafrost temperature feedback is an expression
7 of the decreasing radiative efficiency under high atmospheric greenhouse gas levels. Long-term
8 warming from the release of newly thawed permafrost carbon can add an additional 0.4°C (upper
9 68% range) to global temperatures until the year 2300. Despite CH₄ release contributing only a
10 few percent to total permafrost carbon release, our analyses suggest that it can cause up to about
11 40% (upper 68% range) of permafrost-affected warming. In the 22nd and 23rd century the
12 radiative balance is largely affected by aerobic permafrost carbon release as emitted CO₂
13 accumulates over centuries in the atmosphere – in contrast to the fast decline in CH₄ anomalies
14 with a typical CH₄ life-time of about a decade.

15

16 **4 Discussion and conclusions**

17 This paper presents a new observation-based model for assessing long-term climatic
18 consequences of permafrost degradation. Our simulation strategy consisted in partitioning carbon
19 inventories into different pools of varying soil and surface conditions to model site-specific
20 carbon release. Rather than trying to capture permafrost-carbon dynamics in detail, we instead
21 have aimed at describing in a simplified manner a multitude of processes which are key to
22 permafrost carbon release – such as abrupt thaw in thermokarst-affected sediments. We have
23 especially aimed at accounting for the contribution of carbon release from known deep deposits
24 in the 1.3 million km² large Yedoma region of Siberia and Alaska (Strauss et al., 2013; Walter
25 Anthony et al., 2014), which had been neglected in most previous modelling studies. Our
26 computationally efficient model has enabled us to scan the large uncertainty inherent to
27 observing and modelling the permafrost carbon feedback. In our study we had focused on the
28 contribution of newly thawed permafrost carbon which becomes vulnerable through soil
29 warming above pre-industrial temperatures. However, we stress that the full permafrost carbon
30 feedback is also affected by carbon fluxes from sources not considered in this study, such as the

1 contribution from soil surface layers (seasonally thawed active layer) and changes in high-
2 latitude vegetation. With rising soil temperatures, further contributions will also result from
3 known carbon stocks in permafrost regions, which are not classified as gelisols (e.g. histosols).
4 Finally, abrupt thaw processes other than thermokarst (e.g. caused by wildfires, coastal and
5 thermal erosion) not considered in our study will potentially result in enhanced permafrost
6 carbon fluxes (Grosse et al., 2011).

7 The large spread in future carbon release from permafrost degradation inferred from modelling
8 studies (see Schaefer et al. (2014) and Schuur et al. (2015) for an overview) is caused by various
9 factors. One key issue are pronounced differences in the strength of simulated permafrost
10 degradation. In a recent observationally-constrained model study, Hayes et al. (2014) suggest a
11 mean deepening of the active layer of 6.8 cm over the period 1970 to 2006. We simulate a
12 deepening by 5.9 to 15.5 cm (68% range) over the same period when focusing on our mineral
13 soil pool under aerobic conditions. By the year 2100, our simulations suggest a mean active layer
14 deepening of this pool by 40 to 76 cm under RCP2.6, and of 105 to 316 cm under RCP8.5. The
15 latter range covers a large part of previous estimates, although some studies suggest lower values
16 (Schaefer et al., 2014). Yet a comparison of aggregated simulated active layer depths should be
17 considered with care as differences in definitions (e.g. of the considered permafrost domain and
18 its vertical extent) or different assumptions of future warming can lead to estimating
19 systematically lower or higher active layer depths.

20 Our simulations suggest that permafrost emissions will be strongly constrained when limiting
21 global warming: under a climate mitigation pathway (RCP2.6), the increase in high latitude
22 temperatures results in a moderate deepening of the active layer which stabilizes in most
23 latitudes after the year 2100 (in line with diagnostics based on complex models (Slater and
24 Lawrence, 2013)). Until end of the century about 36 Pg (20 to 58 Pg, 68% range) of carbon was
25 released as CO₂. Under strong warming (RCP8.5), permafrost degradation proves substantial and
26 cumulated CO₂ emissions reached 87 Pg-C (42 to 141 Pg-C, 68% range) by the year 2100. A
27 release of 87 Pg-C corresponds to a mean loss of about 12% of our initial inventory of 750 Pg of
28 carbon perennially frozen under pre-industrial climate. Other modelling studies have estimated a
29 loss of 6 to 33% of initial permafrost carbon stocks, while the majority of models suggest a loss
30 of 10 to 20% (Schaefer et al., 2014). Incubation of permafrost soil samples suggest a carbon loss

1 from mineral soils under aerobic conditions of 13% and 15% over 100 years when assuming
2 thaw during four months in a year (Schädel et al. 2013; Knoblauch et al. 2013).

3 Our analyses have shown a large potential of reducing uncertainty in simulated carbon fluxes
4 especially for climate mitigation pathways when more and spatially higher resolved data of
5 present day permafrost temperatures will be available.

6 Based on our conceptual model of thermokarst lake formation and drainage, our results suggest
7 that abrupt thaw can unlock large amounts of frozen carbon within this century. We infer a
8 deepening of the permafrost table by several meters in 100 years after thermokarst initiation,
9 with additional talik propagation large enough to fully thaw sediments to our lower pool
10 boundary (15m) in the second half of the 22nd century. Subsequent CH₄ release from newly
11 thawed permafrost under RCP8.5 results in emissions that peak at about 50 Tg-CH₄ per year
12 (upper 68% range) in the 21st century. A pronounced spike in CH₄ emissions as a consequence of
13 rapidly expanding and subsequently shrinking thermokarst lake areas is in line with hypotheses
14 of past rapid thermokarst lake formation and expansion. Walter et al. (2007a) suggest an annual
15 CH₄ release of 30 to 40 Tg-CH₄ from thermokarst lakes to partially explain CH₄ excursions of
16 early Holocene atmospheric CH₄ levels. Brosius et al. (2012) discuss a yearly contribution from
17 thermokarst lakes of 15±4 Tg-CH₄ during the Younger Dryas and 25±5 Tg-CH₄ during the
18 Preboreal period.

19 Our modelled total CH₄ fluxes under strong warming are comparable in magnitude to an
20 estimated current release of 24.2±10.5 Tg-CH₄ per year from northern lakes (Walter et al.,
21 2007b). The majority of our results suggest CH₄ fluxes from newly thawed permafrost carbon
22 are an order of magnitude smaller than the contribution from all current natural (about 200 Tg-
23 CH₄/yr) and anthropogenic (about 350 Tg-CH₄/yr) sources (Environmental Protection Agency
24 (EPA), 2010). Focusing on thermokarst lakes in ice-rich deposits (i.e. on Yedoma and refrozen
25 thermokarst deposits), we infer 21st century averaged median emission rates of 6.3 Tg-CH₄/yr
26 which are about double compared to a recent estimate based on a stochastic thaw-lake model for
27 Siberian ice-rich deposits (van Huissteden et al., 2011). Using an integrated earth-system model
28 framework, Gao et al. (2013) estimate that increases in CH₄ emissions until 2100 from inundated
29 area expansion and soil warming range between 5.6 to 15.1 Tg-CH₄/yr. In contrast to our
30 analyses, their simulated CH₄ fluxes are largely dominated by wetland CH₄ release because they

1 assume a fixed value of 3.35 for the wetland:lake ratio in regions north of 45°. Even under
2 assumptions of maximum increases in saturated areas, Gao et al. (2013) simulate future
3 thermokarst lake extents which cover only a few percent of Arctic landscapes. In our model
4 setting (see table 1), we have investigated the scenario of a potential large transformation of
5 northern landscapes, considering up to 50% of ice-rich regions being affected by newly formed
6 thermokarst lakes – and therefore we simulate a much larger CH₄ contribution from permafrost
7 sediments affected by thermokarst.

8 Burke et al. (2012) infer 21st century annual CH₄ emission rates from permafrost wetlands and
9 lakes below 53 Tg-CH₄ for the majority of their model runs. Although our CH₄ release estimates,
10 which are inferred by an independent modelling approach, are comparable in magnitude with
11 recent work, a direct comparison with studies extrapolating observed CH₄ fluxes (e.g. van
12 Huissteden et. al (2011); Gao et al. (2011)) should be considered with care. Observed CH₄ fluxes
13 describe the full carbon balance, including contributions from soil surface layers and vegetation
14 cover, which we do not consider in our model setting.

15

16 In contrast to abrupt thaw and fast release under thermokarst lakes, CH₄ release from newly
17 thawed carbon in wetland-affected soils is slow with discernible contributions only in the 22nd
18 and 23rd century. Although contributing only a few percent to total permafrost carbon release,
19 our simulated CH₄ fluxes from newly thawed permafrost carbon can cause up to 40% of
20 permafrost-affected warming in the 21st century. Given the short lifetime of CH₄, the radiative
21 forcing from permafrost carbon in the 22nd and 23rd century is largely dominated by aerobic CO₂
22 release.

23 Under strong warming, our modelled CH₄ emissions from newly thawed permafrost accumulate
24 to some thousand Tg until the year 2100, with maximum contributions of 10.000 Tg-CH₄ (upper
25 68% range) until the year 2300 (see Table 1). Yet the release of this amount of CH₄ would only
26 slightly affect future atmospheric CH₄ levels under projected RCP CH₄ emissions as the
27 anthropogenic contribution will dominate atmospheric CH₄ concentrations. Based on a carbon
28 mass balance calculation of CH₄ release from Siberian thermokarst lakes, Walter et al. (2007b)
29 suggest a contribution of about 50.000 Tg-CH₄ (or 50 to 100 Tg-CH₄/yr over centuries) in the
30 extremely unlikely case of a complete thaw of the Yedoma deposits.

1 To put into relation the contribution of carbon fluxes from deep deposits to the total, circumpolar
2 release from newly thawed permafrost, we have analysed the contribution of individual pools.
3 Our simulations suggest that the omission of deep carbon stocks is unlikely to strongly affect
4 CO₂ release from permafrost degradation in the coming centuries. In contrast, CH₄ fluxes from
5 newly thawed permafrost are strongly influenced by carbon release from organic matter stored in
6 deep deposits. Although our considered deep pools cover only about 12% of the total area of
7 northern hemisphere gelisols, and despite of the organic matter in these pools being buried deep
8 in the ground, these pools contribute significantly to the total CH₄ balance because abrupt thaw
9 under thermokarst lakes can unlock a large portion of previously inert organic matter. About a
10 quarter of 21st century thermokarst lake CH₄ release stems from newly thawed organic matter
11 stored in deep deposits (i.e. from soil layers deeper than 3m). Further, our analyses revealed that
12 the release from mineralization of labile organic matter contributes disproportionately high to
13 these fluxes. Despite assuming a fast (labile) pool fraction of only a few percent, our simulated
14 CH₄ fluxes from newly thawed labile organic matter account for up to half of the total
15 thermokarst-affected deep CH₄ release in the 21st century. Therefore, improved observational
16 estimates of the share of labile organic matter would help to reduce uncertainty in simulated CH₄
17 release from deep carbon deposits (Strauss et al., 2015). The analysis of individual deep pools
18 revealed a CH₄ release from refrozen thermokarst up to twice the emission from unaltered
19 Yedoma.

20 Our results suggest a mean increase in global average surface temperature of about 0.1°C by the
21 year 2100 (0.03 to 0.14°C, 68% ranges) caused by carbon release from newly thawed permafrost
22 soils. Long-term warming through the permafrost carbon feedback (year 2300) can add an
23 additional 0.4°C (upper 68% range) to projected global mean surface air temperatures.

24 Our analyses suggest that the permafrost-induced additional warming is similar under different
25 scenarios of anthropogenic emissions – despite of largest carbon release from permafrost
26 degradation under strong warming. The weak path dependency is a consequence of the
27 decreasing radiative efficiency of emitted permafrost carbon under increasing background CO₂
28 and CH₄ concentrations.

29 In a previous study (Schneider von Deimling et al. (2012) – referred to as SvD2012 in the
30 following – the authors calculated carbon fluxes from degradation of near-surface permafrost

1 based on a model which described permafrost dynamics in less detail but was coupled to a more
2 comprehensive description of climate-carbon cycle feedbacks (MAGICC-6, Meinshausen et al.,
3 2011). The various differences in model description between SvD2012 and our current study
4 (SvD2015) affect simulated permafrost carbon fluxes and the inferred temperature feedback in
5 multiple ways. In contrast to SvD2012, we now resolve vertical model levels and account for
6 depth dependent thaw dynamics and carbon distribution. This allows us to better initialize our
7 model based on observed active layer profiles and soil carbon concentrations. As a consequence
8 of our improved thaw rate parametrization (see section 2.2 of the supplement), in our new study
9 we simulate increased permafrost thaw (compared to SvD2012), especially under moderate
10 warming. Therefore, we now generally simulate larger carbon fluxes in the 21st century which
11 are also due to an improved tuning of soil carbon decomposition. Yet in our current study, we
12 model smaller cumulated carbon fluxes in the 22nd and 23rd century under RCP8.5 because we
13 consider a smaller fraction of permafrost carbon being available for long-term release.

14 The quantification of additional warming through permafrost carbon release requires a model
15 description of translating permafrost carbon fluxes into atmospheric concentrations of CO₂ and
16 CH₄, and ultimately into global mean temperature anomalies. In SvD2012, these calculations
17 were based on the MAGICC-6 model (Meinshausen et al., 2011), while in our current study we
18 use a more simplified description based on Allen et al. (2009, see supplement section 2.5).
19 Finally, the use of a fully-fledged carbon cycle emulation (MAGICC-6) in SvD2012 results in
20 additional carbon fluxes from non-permafrost terrestrial and oceanic sources which are triggered
21 by additional warming through permafrost degradation – and thus increase the overall
22 temperature feedback. Differences in estimates of permafrost affected warming between
23 SvD2012 and SvD2015 illustrate that factors independent from permafrost dynamics (such as
24 differing model formulations of ocean heat uptake) do affect the strength of the inferred
25 temperature feedback.

26 MacDougall et al. (2012) also modelled a permafrost-carbon feedback largely independent of the
27 emission pathway but inferred larger upper estimates of permafrost-affected warming due to
28 considering a much larger pool available for carbon release triggered by permafrost degradation.
29 An increase in the permafrost temperature feedback with global warming was inferred by Burke
30 et al. (2012) who considered a much larger spread in the near-surface permafrost carbon
31 inventory (~300 to 1800 Pg-C) and who estimated the permafrost temperature feedback by the

1 year 2100 as 0.02 to 0.11°C and 0.08 to 0.36°C (90% ranges) under RCP2.6 and RCP8.5
2 respectively.

3 In conclusion, our results demonstrate that deep carbon deposits and abrupt thaw processes, such
4 as provided by thermokarst lake formation, should be included into future model simulations for
5 an improved representation of the permafrost-carbon feedback.

6

7 **5 Outlook**

8 We consider our estimates conservative because carbon release from further, in this study
9 unaccounted sources, are likely to increase the strength of the full permafrost-carbon feedback.

10 (1) Our study focuses solely on the carbon fluxes resulting from newly thawed soils and deposits
11 in our simulation scenarios, thus excluding carbon fluxes from permafrost-affected soils in the
12 current active layer. These soils will also warm to different levels under RCP scenarios and very
13 likely will be subject to enhanced mineralization of the large already seasonally thawed C pool
14 of about 500 Pg (Hugelius et al., 2014). (2) We do not account for the contribution of newly
15 thawed organic matter of low quality, which we assume recalcitrant on the timescale considered
16 here (i.e. 40 to 70% of thawed organic matter is not available for release). More data and longer
17 time series of incubation experiments, in combination with modelling work of soil-carbon
18 dynamics, are needed to better constrain timescale assumptions for soil organic matter
19 decomposition. Also of importance are improved data-based estimates of CH₄:CO₂ anaerobic
20 production ratios, which determine the share of carbon emitted as CH₄. (3) We do not account
21 for the presence, and potential thaw and mobilization of deep frozen carbon outside the Yedoma
22 and refrozen thermokarst region. Currently no coherent data is available on the distribution and
23 organic carbon characteristics of soils and sediments below three meter depth for large regions in
24 Siberia, Alaska, and Canada. Our model results suggest that these depths will be affected by
25 thaw over the coming centuries and available thawed organic matter would contribute to the
26 permafrost carbon feedback. (4) We do not consider carbon release from degrading submarine
27 permafrost which might result in an underestimation of circumpolar permafrost-affected CH₄
28 fluxes in our study (Shakhova et al., 2010). (6) Extensive permafrost degradation can support a
29 large and abrupt release of fossil CH₄ from below the permafrost cap based on presence of
30 regional hydrocarbon reservoirs and geologic pathways for gas migration (Walter Anthony et al.,

1 2012). We do not consider this pathway of potentially abrupt CH₄ release which could lead to a
2 non-gradual increase in the permafrost-carbon feedback if sub-cap CH₄ increases non-linearly
3 with warming. Likely, the most important omission in our study stems from changes in the high-
4 latitude carbon balance caused by altered vegetation dynamics. Here, an increased carbon uptake
5 through more productive high-latitude vegetation and the renewal of carbon sinks in drained
6 thermokarst basins can considerably decrease the net carbon loss on centennial time-scales
7 (Schaphoff et al., 2013; van Huissteden et al., 2011). Yet this loss can be partially compensated
8 through enhanced respiration of soil-surface organic matter which is stored in large amounts in
9 permafrost regions (but which was not incorporated into permafrost in the past and thus is not
10 considered in this study here). On the other hand, a transition from tundra- towards taiga-
11 dominated landscapes as a consequence of high-latitude warming can strongly decrease surface
12 albedo and therefore additionally warm permafrost regions. We consider the implementation of
13 high-latitude vegetation dynamics into permafrost models a key step towards an improved
14 capturing of the timing and strength of the full permafrost-carbon feedback.

15

16 **Acknowledgements**

17 Special thanks to S. Mathesius for the analysis of CMIP-5 data, G. Hugelius for providing soil
18 carbon data for near-surface permafrost inventories, M. Allen for having provided an earlier
19 version of the climate carbon cycle model, H. Lantuit for discussing aspects of permafrost
20 degradation through coastal erosion, C. Schädel for discussing incubation results of soil carbon
21 lability, and K. Walter Anthony and H. Lee for discussing ratios of CH₄ versus carbon dioxide
22 production rates.

23 This study was supported by the Federal Environment Agency for Germany (UBA) under project
24 UFOPLAN FKZ 3712 41 106 and ERC Starting Grant #338335.

25 J. Strauss was supported by a grant of the Studienstiftung des Deutschen Volkes (German
26 National Academic Foundation), the German Federal Ministry of Education and Research (Grant
27 01DM12011) and the Initiative and Networking Fund of the Helmholtz Association (#ERC-
28 0013). A. Morgenstern was supported by the Helmholtz Postdoc Programme of the Initiative and
29 Networking Fund of the Helmholtz Association (#PD-003). M. Meinshausen was supported by
30 the Australian Research Council grant ARC FT130100809.

1 **References**

2 Abnizova, A., Siemens, J., Langer, M., and Boike, J.: Small ponds with major impact: The
3 relevance of ponds and lakes in permafrost landscapes to carbon dioxide emissions, *Global
4 Biogeochemical Cycles*, 26, GB2041, 10.1029/2011GB004237, 2012.

5 Arp, C. D., Jones, B. M., Lu, Z., and Whitman, M. S.: Shifting balance of thermokarst lake ice
6 regimes across the Arctic Coastal Plain of northern Alaska, *Geophysical Research Letters*, 39,
7 L16503, 10.1029/2012GL052518, 2012.

8 Avis, C. A., Weaver, A. J., and Meissner, K. J.: Reduction in areal extent of high-latitude
9 wetlands in response to permafrost thaw, *Nature Geosci*, 4, 444-448, 2011.

10 Beer, C., Fedorov, A. N., and Torgovkin, Y.: Permafrost temperature and active-layer thickness
11 of Yakutia with 0.5-degree spatial resolution for model evaluation, *Earth Syst. Sci. Data*, 5, 305-
12 310, 10.5194/essd-5-305-2013, 2013.

13 Boike, J., Langer, M., Lantuit, H., Muster, S., Roth, K., Sachs, T., Overduin, P., Westermann, S.,
14 and McGuire, A. D.: Permafrost – Physical Aspects, Carbon Cycling, Databases and
15 Uncertainties, in: *Recarbonization of the Biosphere*, edited by: Lal, R., Lorenz, K., Hüttl, R. F.,
16 Schneider, B. U., and von Braun, J., Springer Netherlands, 159-185, 2012.

17 Boike, J., Kattenstroth, B., Abramova, K., Bornemann, N., Chetverova, A., Fedorova, I., Fröb,
18 K., Grigoriev, M., Grüber, M., Kutzbach, L., Langer, M., Minke, M., Muster, S., Piel, K.,
19 Pfeiffer, E. M., Stoof, G., Westermann, S., Wischnewski, K., Wille, C., and Hubberten, H. W.:
20 Baseline characteristics of climate, permafrost and land cover from a new permafrost
21 observatory in the Lena River Delta, Siberia (1998-2011), *Biogeosciences*, 10, 2105-2128,
22 10.5194/bg-10-2105-2013, 2013.

23 Boike, J., Georgi, C., Kirilin, G., Muster, S., Abramova, K., Fedorova, I., Chetverova, A.,
24 Grigoriev, M., Bornemann, N., and Langer, M.: Physical processes of thermokarst lakes in the
25 continuous permafrost zone of northern Siberia - observations and modeling (Lena River Delta,
26 Siberia), *Biogeosciences Discussions* (accepted), 2015.

27 Brosius, L. S., Walter Anthony, K. M., Grosse, G., Chanton, J. P., Farquharson, L. M., Overduin,
28 P. P., and Meyer, H.: Using the deuterium isotope composition of permafrost meltwater to

1 constrain thermokarst lake contributions to atmospheric methane during the last deglaciation,
2 *Journal of Geophysical Research: Biogeosciences*, 117, G01022, 10.1029/2011JG001810, 2012.

3 Brown J., Hinkel K. M., and Nelson F. E.: The circumpolar active layer monitoring (calm)
4 program: research designs and initial results, *Polar Geogr*, 24, 166–258, 2000.

5 Burke, E. J., Hartley, I. P., and Jones, C. D.: Uncertainties in the global temperature change
6 caused by carbon release from permafrost thawing, *The Cryosphere*, 6, 1063-1076, 10.5194/tc-6-
7 1063-2012, 2012.

8 Camill, P.: Permafrost Thaw Accelerates in Boreal Peatlands During Late-20th Century Climate
9 Warming, *Climatic Change*, 68, 135-152, 10.1007/s10584-005-4785-y, 2005.

10 Conrad, R., Klose, M., and Claus, P.: Pathway of methane formation in anoxic rice field soil and
11 rice roots determined by ^{13}C -stable isotope fractionation, *Chemosphere*, 47, 797-806,
12 10.1016/S0045-6535(02)00120-0, 2002.

13 Dankers, R., Burke, E. J., and Price, J.: Simulation of permafrost and seasonal thaw depth in the
14 JULES land surface scheme, *The Cryosphere*, 5, 773-790, 10.5194/tc-5-773-2011, 2011.

15 Dutta, K., Schuur, E. A. G., Neff, J. C., and Zimov, S. A.: Potential carbon release from
16 permafrost soils of Northeastern Siberia, *Global Change Biology*, 12, 2336-2351, 2006.

17 Ekici, A., Beer, C., Hagemann, S., Boike, J., Langer, M., and Hauck, C.: Simulating high-
18 latitude permafrost regions by the JSBACH terrestrial ecosystem model, *Geosci. Model Dev.*, 7,
19 631-647, 10.5194/gmd-7-631-2014, 2014.

20 Eliseev, A. V., Mokhov, I. I., Arzhanov, M. M., Demchenko, P. F., and Denisov, S. N.:
21 Interaction of the methane cycle and processes in wetland ecosystems in a climate model of
22 intermediate complexity, *Izvestiya Atmospheric and Oceanic Physics*, 44, 139-152, 2008.

23 Environmental Protection Agency (EPA): Methane and nitrous oxide emissions from natural
24 sources EPA 430-R-10-001, US Environmental Protection Agency, Washington, DC, 2010.

25 Frauenfeld, O. W., Zhang, T., Barry, R. G., and Gilichinsky, D.: Interdecadal changes in
26 seasonal freeze and thaw depths in Russia, *Journal of Geophysical Research: Atmospheres*, 109,
27 D05101, 10.1029/2003JD004245, 2004.

1 Friedlingstein, P., Cox, P., Betts, R., Bopp, L., von Bloh, W., Brovkin, V., Cadule, P., Doney, S.,
2 Eby, M., Fung, I., Bala, G., John, J., Jones, C., Joos, F., Kato, T., Kawamiya, M., Knorr, W.,
3 Lindsay, K., Matthews, H. D., Raddatz, T., Rayner, P., Reick, C., Roeckner, E., Schnitzler, K.-
4 G., Schnur, R., Strassmann, K., Weaver, K., Yoshikawa, C., and Zeng, N.: Climate–Carbon
5 Cycle Feedback Analysis: Results from the C4MIP Model Intercomparison, *Journal of Climate*,
6 19, 3337–3353, 10.1175/JCLI3800.1, 2006.

7 Frolking, S., Roulet, N. T., Moore, T. R., Richard, P. J. H., Lavoie, M., and Muller, S. D.:
8 Modeling northern peatland decomposition and peat accumulation, *Ecosystems*, 4, 479–498,
9 2001.

10 Gao, X., Schlosser, C. A., Sokolov, A., Anthony, K. W., Zhuang, Q., and Kicklighter, D.:
11 Methane Feedback : What is the Worst We Can Expect ?, *Environmental Research Letters*, 8,
12 2013.

13 Gouttevin, I., Menegoz, M., Dominé, F., Krinner, G., Koven, C., Ciais, P., Tarnocai, C., and
14 Boike, J.: How the insulating properties of snow affect soil carbon distribution in the continental
15 pan-Arctic area, *Journal of Geophysical Research: Biogeosciences*, 117, G02020,
16 10.1029/2011JG001916, 2012.

17 Grosse, G., Schirrmeister, L., Siegert, C., Kunitsky, V. V., Slagoda, E. A., Andreev, A. A., and
18 Dereviagyn, A. Y.: Geological and geomorphological evolution of a sedimentary periglacial
19 landscape in Northeast Siberia during the Late Quaternary, *Geomorphology*, 86, 25–51,
20 10.1016/j.geomorph.2006.08.005, 2007.

21 Grosse, G., Harden, J., Turetsky, M., McGuire, A. D., Camill, P., Tarnocai, C., Frolking, S.,
22 Schuur, E. A. G., Jorgenson, T., Marchenko, S., Romanovsky, V., Wickland, K. P., French, N.,
23 Waldrop, M., Bourgeau-Chavez, L., and Striegl, R. G.: Vulnerability of high-latitude soil organic
24 carbon in North America to disturbance, *J. Geophys. Res.*, 116, G00K06, 2011.

25 Harden, J. W., Koven, C. D., Ping, C.-L., Hugelius, G., David McGuire, A., Camill, P.,
26 Jorgenson, T., Kuhry, P., Michaelson, G. J., O'Donnell, J. A., Schuur, E. A. G., Tarnocai, C.,
27 Johnson, K., and Grosse, G.: Field information links permafrost carbon to physical
28 vulnerabilities of thawing, *Geophysical Research Letters*, 39, L15704, 10.1029/2012GL051958,
29 2012.

1 Hayes, D. J., Kicklighter, D. W., McGuire, A. D., Chen, M., Zhuang, Q., Yuan, F., Melillo, J.
2 M., and Wullschleger, S. D.: The impacts of recent permafrost thaw on land–atmosphere
3 greenhouse gas exchange, *Environmental Research Letters*, 9, 045005, 2014.

4 Hugelius, G., Tarnocai, C., Broll, G., Canadell, J. G., Kuhry, P., and Swanson, D. K.: The
5 Northern Circumpolar Soil Carbon Database: spatially distributed datasets of soil coverage and
6 soil carbon storage in the northern permafrost regions, *Earth Syst. Sci. Data*, 5, 3-13,
7 10.5194/essd-5-3-2013, 2013.

8 Hugelius, G., Strauss, J., Zubrzycki, S., Harden, J. W., Schuur, E. A. G., Ping, C. L.,
9 Schirrmeister, L., Grosse, G., Michaelson, G. J., Koven, C. D., O'Donnell, J. A., Elberling, B.,
10 Mishra, U., Camill, P., Yu, Z., Palmtag, J., and Kuhry, P.: Estimated stocks of circumpolar
11 permafrost carbon with quantified uncertainty ranges and identified data gaps, *Biogeosciences*,
12 11, 6573-6593, 10.5194/bg-11-6573-2014, 2014.

13 Jafarov, E. E., Marchenko, S. S., and Romanovsky, V. E.: Numerical modeling of permafrost
14 dynamics in Alaska using a high spatial resolution dataset, *The Cryosphere*, 6, 613-624,
15 10.5194/tc-6-613-2012, 2012.

16 Johnson, K. D., Harden, J. W., McGuire, A. D., Clark, M., Yuan, F., and Finley, A. O.:
17 Permafrost and organic layer interactions over a climate gradient in a discontinuous permafrost
18 zone, *Environmental Research Letters*, 8, 035028, 2013.

19 Jones, B. M., Grosse, G., Arp, C. D., Jones, M. C., Walter Anthony, K. M., and Romanovsky, V.
20 E.: Modern thermokarst lake dynamics in the continuous permafrost zone, northern Seward
21 Peninsula, Alaska, *Journal of Geophysical Research: Biogeosciences*, 116, G00M03,
22 10.1029/2011JG001666, 2011.

23 Jones, M. C., Grosse, G., Jones, B. M., and Walter Anthony, K.: Peat accumulation in drained
24 thermokarst lake basins in continuous, ice-rich permafrost, northern Seward Peninsula, Alaska,
25 *Journal of Geophysical Research: Biogeosciences*, 117, G00M07, 10.1029/2011JG001766, 2012.

26 Jorgenson, M. T., Shur, Y. L., and Pullman, E. R.: Abrupt increase in permafrost degradation in
27 Arctic Alaska, *Geophysical Research Letters*, 33, L02503-L02503, 2006.

28 Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G.,
29 Kanevskiy, M., and Marchenko, S.: Resilience and vulnerability of permafrost to climate change,

1 Canadian Journal of Forest Research-*Revue Canadienne De Recherche Forestiere*, 40, 1219-
2 1236, 2010.

3 Kaufman, D. S., Ager, T. A., Anderson, N. J., Anderson, P. M., Andrews, J. T., Bartlein, P. J.,
4 Brubaker, L. B., Coats, L. L., Cwynar, L. C., Duvall, M. L., Dyke, A. S., Edwards, M. E., Eisner,
5 W. R., Gajewski, K., Geirsdóttir, A., Hu, F. S., Jennings, A. E., Kaplan, M. R., Kerwin, M. W.,
6 Lozhkin, A. V., MacDonald, G. M., Miller, G. H., Mock, C. J., Oswald, W. W., Otto-Bliesner, B.
7 L., Porinchu, D. F., Rühland, K., Smol, J. P., Steig, E. J., and Wolfe, B. B.: Holocene thermal
8 maximum in the western Arctic (0–180°W), *Quaternary Science Reviews*, 23, 529-560,
9 10.1016/j.quascirev.2003.09.007, 2004.

10 Kessler, M. A., Plug, L. J., and Walter Anthony, K. M.: Simulating the decadal- to millennial-
11 scale dynamics of morphology and sequestered carbon mobilization of two thermokarst lakes in
12 NW Alaska, *Journal of Geophysical Research: Biogeosciences*, 117, G00M06,
13 10.1029/2011JG001796, 2012.

14 Khvorostyanov, D. V., Ciais, P., Krinner, G., and Zimov, S. A.: Vulnerability of east Siberia's
15 frozen carbon stores to future warming, *Geophysical Research Letters*, 35, L10703, 2008.

16 Kleinen, T., Brovkin, V., and Schultdt, R. J.: A dynamic model of wetland extent and peat
17 accumulation: results for the Holocene, *Biogeosciences*, 9, 235-248, 10.5194/bg-9-235-2012,
18 2012.

19 Knoblauch, C., Beer, C., Sosnin, A., Wagner, D., Pfeiffer, E. M.: Predicting long-term carbon
20 mineralization and trace gas production from thawing permafrost of Northeast Siberia, *Global
21 Change Biology*, 19, 1160–1172, 2013.

22 Kokelj, S. V., Lacelle, D., Lantz, T. C., Tunnicliffe, J., Malone, L., Clark, I. D., and Chin, K. S.:
23 Thawing of massive ground ice in mega slumps drives increases in stream sediment and solute
24 flux across a range of watershed scales, *Journal of Geophysical Research: Earth Surface*, 118,
25 681-692, 10.1002/jgrf.20063, 2013.

26 Koven, C., Friedlingstein, P., Ciais, P., Khvorostyanov, D., Krinner, G., and Tarnocai, C.: On the
27 formation of high-latitude soil carbon stocks: Effects of cryoturbation and insulation by organic
28 matter in a land surface model, *Geophysical Research Letters*, 36, L21501,
29 10.1029/2009GL040150, 2009.

1 Koven, C. D., Ringeval, B., Friedlingstein, P., Ciais, P., Cadule, P., Khvorostyanov, D., Krinner,
2 G., and Tarnocai, C.: Permafrost carbon-climate feedbacks accelerate global warming,
3 Proceedings of the National Academy of Sciences, 108, 14769-14774, 2011.

4 Koven, C. D., Riley, W. J., and Stern, A.: Analysis of Permafrost Thermal Dynamics and
5 Response to Climate Change in the CMIP5 Earth System Models, Journal of Climate, 26, 1877-
6 1900, 10.1175/JCLI-D-12-00228.1, 2013a.

7 Koven, C. D., Riley, W. J., Subin, Z. M., Tang, J. Y., Torn, M. S., Collins, W. D., Bonan, G. B.,
8 Lawrence, D. M., and Swenson, S. C.: The effect of vertically resolved soil biogeochemistry and
9 alternate soil C and N models on C dynamics of CLM4, Biogeosciences, 10, 7109-7131,
10 10.5194/bg-10-7109-2013, 2013b.

11 Krinner, G., and Boike, J.: A study of the large-scale climatic effects of a possible disappearance
12 of high-latitude inland water surfaces during the 21st century, Boreal Environment Research 15,
13 203-217, 2010.

14 Langer, M., Westermann, S., Heikenfeld, M., Dorn, W., and Boike, J.: Satellite-based modeling
15 of permafrost temperatures in a tundra lowland landscape, Remote Sensing of Environment, 135,
16 12-24, 10.1016/j.rse.2013.03.011, 2013.

17 Laurion, I., Vincent, W. F., MacIntyre, S., Retamal, L., Dupont, C., Francus, P., Pienitz, R.:
18 Variability in greenhouse gas emissions from permafrost thaw ponds, Limnology and
19 Oceanography 55, 115–133, 2010.

20 Lawrence, D. M., and Slater, A. G.: Incorporating organic soil into a global climate model,
21 Climate Dynamics, 30, 145-160, 2008.

22 Lawrence, D. M., Oleson, K. W., Flanner, M. G., Thornton, P. E., Swenson, S. C., Lawrence, P.
23 J., Zeng, X., Yang, Z.-L., Levis, S., Sakaguchi, K., Bonan, G. B., and Slater, A. G.:
24 Parameterization improvements and functional and structural advances in Version 4 of the
25 Community Land Model, Journal of Advances in Modeling Earth Systems, 3,
26 10.1029/2011ms000045, 2011.

27 Lee, H., Schuur, E. A. G., Inglett, K. S., Lavoie, M., and Chanton, J. P.: The rate of permafrost
28 carbon release under aerobic and anaerobic conditions and its potential effects on climate, Global
29 Change Biology, 18, 515-527, 10.1111/j.1365-2486.2011.02519.x, 2012.

1 Ling, F.: Numerical simulation of permafrost thermal regime and talik development under
2 shallow thaw lakes on the Alaskan Arctic Coastal Plain, *Journal of Geophysical Research*, 108,
3 4511-4511, 2003.

4 Ling, F., Wu, Q., Zhang, T., and Niu, F.: Modelling Open-Talik Formation and Permafrost
5 Lateral Thaw under a Thermokarst Lake, Beiluhe Basin, Qinghai-Tibet Plateau, *Permafrost and*
6 *Periglacial Processes*, 23, 312-321, 10.1002/ppp.1754, 2012.

7 MacDougall, A. H., Avis, C. A., and Weaver, A. J.: Significant contribution to climate warming
8 from the permafrost carbon feedback, *Nature Geosci*, advance online publication, 2012.

9 Mack, M. C., Schuur, E. A. G., Bret-Harte, M. S., Shaver, G. R., and Chapin, F. S.: Ecosystem
10 carbon storage in arctic tundra reduced by long-term nutrient fertilization, *Nature*, 431, 440-443,
11 2004.

12 Marcott, S. A., Shakun, J. D., Clark, P. U., and Mix, A. C.: A Reconstruction of Regional and
13 Global Temperature for the Past 11,300 Years, *Science*, 339, 1198-1201,
14 10.1126/science.1228026, 2013.

15 Mishra, U., Jastrow, J. D., Matamala, R., Hugelius, G., Koven, C. D., Harden, J. W., Ping, C. L.,
16 Michaelson, G. J., Fan, Z., Miller, R. M., McGuire, A. D., Tarnocai, C., Kuhry, P., Riley, W. J.,
17 Schaefer, K., Schuur, E. A. G., Jorgenson, M. T., and Hinzman, L. D.: Empirical estimates to
18 reduce modeling uncertainties of soil organic carbon in permafrost regions: a review of recent
19 progress and remaining challenges, *Environmental Research Letters*, 8, 035020, 2013.

20 Morgenstern, A., Grosse, G., Günther, F., Fedorova, I., and Schirrmeister, L.: Spatial analyses of
21 thermokarst lakes and basins in Yedoma landscapes of the Lena Delta, *The Cryosphere*, 5, 849-
22 867, 10.5194/tc-5-849-2011, 2011.

23 Myhre, G., Highwood, E. J., Shine, K. P., and Stordal, F.: New estimates of radiative forcing due
24 to well mixed greenhouse gases, *Geophys. Res. Lett.*, 25, 2715-2718, 1998.

25 Olefeldt, D., Turetsky, M. R., Crill, P. M., and McGuire, A. D.: Environmental and physical
26 controls on northern terrestrial methane emissions across permafrost zones, *Global Change*
27 *Biology*, 19, 589-603, 10.1111/gcb.12071, 2013.

28 Osterkamp, T. E.: Causes of warming and thawing permafrost in Alaska, *Eos, Transactions*
29 *American Geophysical Union*, 88, 522-523, 10.1029/2007EO480002, 2007.

1 Plug, L. J., and West, J. J.: Thaw lake expansion in a two-dimensional coupled model of heat
2 transfer, thaw subsidence, and mass movement, *Journal of Geophysical Research: Earth Surface*,
3 114, F01002, 10.1029/2006JF000740, 2009.

4 Reynolds, M. K., Walker, D. A., Ambrosius, K. J., Brown, J., Everett, K. R., Kanevskiy, M.,
5 Kofinas, G. P., Romanovsky, V. E., Shur, Y., and Webber, P. J.: Cumulative geoecological
6 effects of 62 years of infrastructure and climate change in ice-rich permafrost landscapes,
7 Prudhoe Bay Oilfield, Alaska, *Global Change Biology*, 20, 1211-1224, 10.1111/gcb.12500,
8 2014.

9 Romanovsky, V. E., Smith, S. L., and Christiansen, H. H.: Permafrost thermal state in the polar
10 Northern Hemisphere during the international polar year 2007–2009: a synthesis, *Permafrost and*
11 *Periglacial Processes*, 21, 106-116, 10.1002/ppp.689, 2010.

12 Sannel, A. B. K., and Kuhry, P.: Warming-induced destabilization of peat plateau/thermokarst
13 lake complexes, *Journal of Geophysical Research: Biogeosciences*, 116, G03035,
14 10.1029/2010JG001635, 2011.

15 Scanlon, D., and Moore, T.: Carbon dioxide production from peatland soil profiles: The
16 influence of temperature, oxic/anoxic conditions and substrate, *Soil Science*, 165, 153-160, 2000.

17 Schädel, C., Schuur, E. A. G., Bracho, R., Elberling, B., Knoblauch, C., Lee, H., Luo, Y.,
18 Shaver, G. R., and Turetsky, M. R.: Circumpolar assessment of permafrost C quality and its
19 vulnerability over time using long-term incubation data, *Global Change Biology*, 20, 641-652,
20 10.1111/gcb.12417, 2014.

21 Schaefer, K., Lantuit, H., Romanovsky, V. E., Schuur, E. A. G., and Witt, R.: The impact of the
22 permafrost carbon feedback on global climate, *Environmental Research Letters*, 9, 085003, 2014.

23 Schaphoff, S., Heyder, U., Ostberg, S., Gerten, D., Heinke, J., and Lucht, W.: Contribution of
24 permafrost soils to the global carbon budget, *Environmental Research Letters*, 8, 014026, 2013.

25 Schirrmeister, L., Grosse, G., Wetterich, S., Overduin, P. P., Strauss, J., Schuur, E. A. G., and
26 Hubberten, H.-W.: Fossil organic matter characteristics in permafrost deposits of the northeast
27 Siberian Arctic, *Journal of Geophysical Research: Biogeosciences*, 116, G00M02,
28 10.1029/2011JG001647, 2011.

1 Schmidt, M. W. I., Torn, M. S., Abiven, S., Dittmar, T., Guggenberger, G., Janssens, I. A.,
2 Kleber, M., Kogel-Knabner, I., Lehmann, J., Manning, D. A. C., Nannipieri, P., Rasse, D. P.,
3 Weiner, S., and Trumbore, S. E.: Persistence of soil organic matter as an ecosystem property,
4 *Nature*, 478, 49-56, 2011.

5 Schneider von Deimling, T., Meinshausen, M., Levermann, a., Huber, V., Frieler, K., Lawrence,
6 D. M., and Brovkin, V.: Estimating the near-surface permafrost-carbon feedback on global
7 warming, *Biogeosciences*, 9, 649-665, 2012.

8 Schuur, E. A. G., Bockheim, J., Canadell, J. G., Euskirchen, E., Field, C. B., Goryachkin, S. V.,
9 Hagemann, S., Kuhry, P., Lafleur, P. M., Lee, H., Mazhitova, G., Nelson, F. E., Rinke, A.,
10 Romanovsky, V. E., Shiklomanov, N., Tarnocai, C., Venevsky, S., Vogel, J. G., and Zimov, S.
11 A.: Vulnerability of permafrost carbon to climate change: Implications for the global carbon
12 cycle, *Bioscience*, 58, 701-714, 2008.

13 Schuur, E. A. G., Abbott, B. W., Bowden, W. B., Brovkin, V., Camill, P., Canadell, J. G.,
14 Chanton, J. P., Chapin, F. S., III, Christensen, T. R., Ciais, P., Crosby, B. T., Czimczik, C. I.,
15 Grosse, G., Harden, J., Hayes, D. J., Hugelius, G., Jastrow, J. D., Jones, J. B., Kleinen, T.,
16 Koven, C. D., Krinner, G., Kuhry, P., Lawrence, D. M., McGuire, A. D., Natali, S. M.,
17 O'Donnell, J. A., Ping, C. L., Riley, W. J., Rinke, A., Romanovsky, V. E., Sannel, A. B. K.,
18 Schädel, C., Schaefer, K., Sky, J., Subin, Z. M., Tarnocai, C., Turetsky, M. R., Waldrop, M. P.,
19 Walter Anthony, K. M., Wickland, K. P., Wilson, C. J., and Zimov, S. A.: Expert assessment of
20 vulnerability of permafrost carbon to climate change, *Climatic Change*, 119, 359-374,
21 10.1007/s10584-013-0730-7, 2013.

22 Schuur, E. A. G., McGuire, A. D., Schadel, C., Grosse, G., Harden, J. W., Hayes, D. J.,
23 Hugelius, G., Koven, C. D., Kuhry, P., Lawrence, D. M., Natali, S. M., Olefeldt, D.,
24 Romanovsky, V. E., Schaefer, K., Turetsky, M. R., Treat, C. C., and Vonk, J. E.: Climate change
25 and the permafrost carbon feedback, *Nature*, 520, 171-179, 10.1038/nature14338, 2015.

26 Segers, R.: Methane production and methane consumption: a review of processes underlying
27 wetland methane fluxes, *Biogeochemistry*, 41, 23-51, 1998.

1 Shakhova, N., Semiletov, I., Salyuk, A., Yusupov, V., Kosmach, D., and Gustafsson, O.:
2 Extensive Methane Venting to the Atmosphere from Sediments of the East Siberian Arctic Shelf,
3 Science, 327, 1246-1250, 2010.

4 Shindell, D. T., Faluvegi, G., Koch, D. M., Schmidt, G. A., Unger, N., and Bauer, S. E.:
5 Improved Attribution of Climate Forcing to Emissions, Science, 326, 716-718,
6 10.1126/science.1174760, 2009.

7 Sitch, S., Smith, B., Prentice, I. C., Arneth, A., Bondeau, A., Cramer, W., Kaplan, J. O., Levis,
8 S., Lucht, W., Sykes, M. T., Thonicke, K., and Venevsky, S.: Evaluation of ecosystem dynamics,
9 plant geography and terrestrial carbon cycling in the LPJ dynamic global vegetation model,
10 Global Change Biology, 9, 161-185, 2003.

11 Slater, A. G., and Lawrence, D. M.: Diagnosing Present and Future Permafrost from Climate
12 Models, Journal of Climate, 26, 5608-5623, 10.1175/JCLI-D-12-00341.1, 2013.

13 Smith, L. C., Sheng, Y., MacDonald, G. M., and Hinzman, L. D.: Disappearing arctic lakes,
14 Science, 308, 1429-1429, 2005.

15 Stieglitz, M., Déry, S. J., Romanovsky, V. E., and Osterkamp, T. E.: The role of snow cover in
16 the warming of arctic permafrost, Geophysical Research Letters, 30, 1721,
17 10.1029/2003GL017337, 2003.

18 Strauss, J., Schirrmeyer, L., Wetterich, S., Borchers, A., and Davydov, S. P.: Grain-size
19 properties and organic-carbon stock of Yedoma Ice Complex permafrost from the Kolyma
20 lowland, northeastern Siberia, Global Biogeochemical Cycles, 26, GB3003,
21 10.1029/2011GB004104, 2012.

22 Strauss, J., Schirrmeyer, L., Grosse, G., Wetterich, S., Ulrich, M., Herzschuh, U., and
23 Hubberten, H.-W.: The deep permafrost carbon pool of the Yedoma region in Siberia and
24 Alaska, Geophysical Research Letters, 40, 6165–6170, 10.1002/2013GL058088, 2013.

25 Strauss, J., Schirrmeyer, L., Mangelsdorf, K., Eichhorn, L., Wetterich, S., and Herzschuh, U.:
26 Organic matter quality of deep permafrost carbon - a study from Arctic Siberia, Biogeosciences,
27 12, 2227-2245), 10.5194/bg-12-2227-2015, 2015.

1 Taylor, K. E., Stouffer, R. J., and Meehl, G. A.: An Overview of CMIP5 and the Experiment
2 Design, *Bulletin of the American Meteorological Society*, 93, 485-498, 10.1175/BAMS-D-11-
3 00094.1, 2011.

4 Turetsky, M. R., Kane, E. S., Harden, J. W., Ottmar, R. D., Manies, K. L., Hoy, E., and
5 Kasischke, E. S.: Recent acceleration of biomass burning and carbon losses in Alaskan forests
6 and peatlands, *Nature Geoscience*, 4, 27-31, 2011.

7 van Huissteden, J., Berrittella, C., Parmentier, F. J. W., Mi, Y., Maximov, T. C., and Dolman, A.
8 J.: Methane emissions from permafrost thaw lakes limited by lake drainage, *Nature Clim.
9 Change*, 1, 119-123, 2011.

10 van Huissteden, J., and Dolman, A.: Soil carbon in the Arctic and the permafrost carbon
11 feedback, *Current Opinion in Environmental Sustainability*, 4, 545-551, 2012.

12 Velichko, A. A., Catto, N., Drenova, A. N., Klimanov, V. A., Kremenetski, K. V., and Nechaev,
13 V. P.: Climate changes in East Europe and Siberia at the Late glacial–holocene transition,
14 *Quaternary International*, 91, 75-99, 10.1016/S1040-6182(01)00104-5, 2002.

15 Walter Anthony, K. M., Anthony, P., Grosse, G., and Chanton, J.: Geologic methane seeps along
16 boundaries of Arctic permafrost thaw and melting glaciers, *Nature Geoscience*, 5, 419-426,
17 2012.

18 Walter Anthony, K. M., Zimov, S. A., Grosse, G., Jones, M. C., Anthony, P. M., Iii, F. S. C.,
19 Finlay, J. C., Mack, M. C., Davydov, S., Frenzel, P., and Froliking, S.: A shift of thermokarst
20 lakes from carbon sources to sinks during the Holocene epoch, *Nature*, 511, 452-456,
21 10.1038/nature13560, 2014.

22 Walter, B. P., and Heimann, M.: A process-based, climate-sensitive model to derive methane
23 emissions from natural wetlands: Application to five wetland sites, sensitivity to model
24 parameters, and climate, *Global Biogeochem. Cycles*, 14, 745-765, 2000.

25 Walter, K. M., Edwards, M. E., Grosse, G., Zimov, S. A., and Chapin, F. S.: Thermokarst lakes
26 as a source of atmospheric methane during the last deglaciation, *Science*, 318, 633-636, 2007a.

27 Walter, K. M., Smith, L. C., and Chapin, F. S.: Methane bubbling from northern lakes: present
28 and future contributions to the global methane budget, *Philosophical Transactions of the Royal
29 Society a-Mathematical Physical and Engineering Sciences*, 365, 1657-1676, 2007b.

1 Wisser, D., Marchenko, S., Talbot, J., Treat, C., and Frolking, S.: Soil temperature response to
2 21st century global warming: the role of and some implications for peat carbon in thawing
3 permafrost soils in North America, *Earth Syst. Dynam.*, 2, 121-138, 2011.

4 Yi, S., Wischnewski, K., Langer, M., Muster, S., and Boike, J.: Freeze/thaw processes in
5 complex permafrost landscapes of northern Siberia simulated using the TEM ecosystem model:
6 impact of thermokarst ponds and lakes, *Geosci. Model Dev.*, 7, 1671-1689, 10.5194/gmd-7-
7 1671-2014, 2014.

8 Yoshikawa, K., and Hinzman, L. D.: Shrinking thermokarst ponds and groundwater dynamics in
9 discontinuous permafrost near council, Alaska, *Permafrost and Periglacial Processes*, 14, 151-
10 160, 10.1002/ppp.451, 2003.

11 Zimov, S. A., Davydov, S. P., Zimova, G. M., Davydova, A. I., Schuur, E. A. G., Dutta, K., and
12 Chapin, F. S.: Permafrost carbon: Stock and decomposability of a globally significant carbon
13 pool, *Geophysical Research Letters*, 33, L20502, 10.1029/2006GL027484, 2006.

14 Zhang, T., Frauenfeld, O. W., and Barry, R. G.: Time series of active layer thickness in the
15 Russian Arctic, 1930-1990, Digital media, National Snow and Ice Data Center, Boulder, CO.
16 URL <http://nsidc.org/data/arcss160.html>, 2006

17

18

1 Table 1. Permafrost model parameters and uncertainties.
 2 Some parameters are soil pool specific (MS: mineral soils, ORG: organic soils, Y: Yedoma,
 3 RTK: refrozen thermokarst deposits (separated into surface and taberal sediments), some
 4 parameters depend on hydrologic conditions (AER: aerobic, WET: wetland anaerobic, TKL:
 5 thermokarst lake anaerobic), and some parameters depend on organic matter quality (FAST and
 6 SLOW).

Parameter	Unit	Default setting	Uncertainty range	References
Carbon inventory				
Mineral soils (MS) 0-3m (orthels & turbels)	Pg-C	540	±40%	Hugelius et. al (2014)
Organic soils (ORG) 0-3m (histels)	Pg-C	120	±40%	Hugelius et. al (2014)
Yedoma (Y) 0-15m	Pg-C	83	±75%	Strauss et al. (2013)
Refrozen thermokarst deposits				
RTK _{Surface} (0-5m)	Pg-C	128	±75%	Strauss et al. (2013)
RTK _{Taberal} (5-15m)		114	±75%	Walter-Anthony et al. (2014)
Fraction Fast Pool ^(a)	%	2.5	1-4	(Dutta et al. (2006);Burke et al. (2012);Schädel et al. (2014))
Fraction Slow Pool	%	45	30-60	(Sitch et al. (2003);Koven et al. (2011);Burke et al. (2012))
Carbon release				
Turnover time of aerobic slow pool at 5°C ^(b)	yrs	25	10-40	Sitch et al. (2003), Burke et al. (2012), Dutta et al. (2006)
Ratio of production CH ₄ :CO ₂ ^{aerobic}		1:50	±50%	Lee et al. (2012);Schuur et al. (2008);Segers (1998)
Ratio of production CH ₄ :CO ₂ ^{anaerobic} ^(c)		FAST 1:1 SLOW 1:7	±20% ±50%	Walter-Anthony et al. (2014) Lee et al. (2012)
Q ₁₀ sensitivity aerobic		2.5	1.5-3.5	Schädel et al. (2013) and

references therein				
Q ₁₀ sensitivity anaerobic		3.0	2-6	Walter and Heimann (2000)
CH ₄ oxidation rate	%	TKL 15 WET 40	10-20 20-60	See Burke et al. (2012) and references therein
Permafrost thaw				
Thaw rate (MS, AER) for warm and cold permafrost ^(d)	cm/yr/K	1.0 0.1	±50% ±50%	Frauenfeld et al. (2004), Hayes et al. (2014), Schaphoff et al. (2013)
Scale factor thermal diffusivity WET:AER ^(e)		1/3	±30%	see ^(e)
Scale factor thermal diffusivity TKL:AER ^(e)		9.3	±30%	Kessler et al. (2012)
Wetland description				
Wetland extent ^(f) (pre-industrial)	%	MS 2 ORG 60 Y, RTK 40	±50% ±10% ±10%	GLWD, Lehner and Döll (2004) Burke et al. (2012)
maximum increase in wetland extent ^(g) (above pre-industrial)	%	MS 30 ORG,Y,RTK 10	±50% ±50%	Gao et al. (2013)
Thermokarst description				
Newly formed thermokarst lake fraction F^{TKLmax}	% (coverage per latitude)	MS 8 ORG 16 Y 40 RTK 25	±25% ±25% ±25% ±25%	see supplementary material
High latitude temperature anomaly dT^{TKLmax} at F^{TKLmax} ^(h)	°C	5	4-6	see supplementary material

^(a) For Yedoma deposits, we assume a doubled labile fraction (5±3%) as sedimentation of organic material was rather fast and had favoured the burial of fresh organic carbon with little decomposition in the past (Strauss et al., 2012). In contrast, we assume a reduced labile fraction in taberul sediments of 1% as these deposits had been thawed over long timescales in the past and are therefore depleted in high quality organic matter (Walter et al., 2007b;Kessler et al., 2012).

^(b) We assume the turnover time of the fast pool to be one year.

^(c) We discard very small ratios of CH₄:CO₂^{anaerobic} inferred from incubation experiments as it is likely that these ratios are strongly affected by a large CO₂ pulse during the initial phase of the incubation.

1 ^(d) Indicated thaw rates are exemplary for warm and cold permafrost (corresponding to a MAGT of just
2 below 0°C and -10°C). They were calculated based on equation (1) (supplementary material) by
3 assuming that above-zero temperatures prevail during four months per year and that thaw is driven by a
4 surface temperature warming anomaly of 1°C.

5 ^(e) We prescribe aggregated thermal diffusivities for soils under aerobic conditions and use scale factors
6 to determine modified thermal diffusivities under anaerobic conditions. Based on observational evidence
7 (Romanovsky et al., 2010), we assume reduced thaw rates for the wetland pools as water-saturated soils
8 require an increased latent heat input for thaw of ice-filled pore volumes. For the thermokarst soil carbon
9 pools, we tuned scaling factors to reproduce long-term behaviour of talik propagation as simulated by
10 Kessler et al. (2012).

11 ^(f) Based on the GLWD database, Burke et al. (2012) estimate an area coverage of 9% for wetlands and
12 3% for lakes for all permafrost regions. Based on calculated permafrost deposit extents (Hugelius et al.,
13 2014), we estimate an area weighting of 80%:15%:2.5%:2.5% for the permafrost extents of our four soil
14 pools (MS:ORG:Y:RTK). This results in a total weighted initial wetland extent of about 13%.

15 ^(g) The potential for increases in wetland extent in mineral soils is considered larger than for the other soil
16 pools because the initial assumed wetland fraction in mineral soils is rather small.

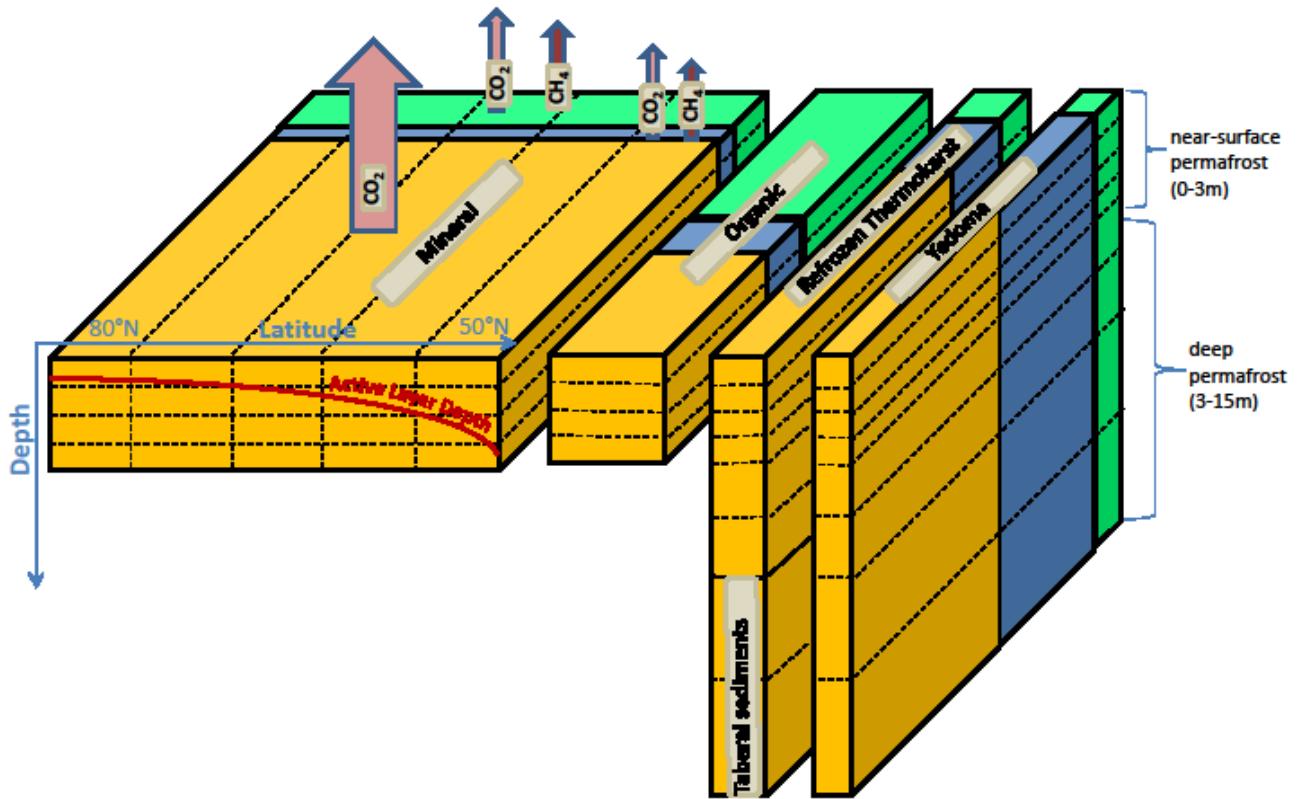
17 ^(h) Early Holocene warming by a few degrees Celsius in northern hemisphere land areas (Kaufman et al.,
18 2004; Velichko et al., 2002; Marcott et al., 2013) resulted in rapid and intensive thermokarst activity (Walter
19 et al., 2007a; Brosius et al., 2012).

1 Table 2.
 2 Cumulated carbon fluxes and increase in global average surface temperature through newly
 3 thawed permafrost in the years 2050, 2100, 2200 and 2300. Median and 68% ranges (in brackets)
 4 were calculated from an ensemble of 500 model runs which account for parameter uncertainty.

	2050	2100	2200	2300
RCP2.6				
cumulated CO ₂ [Pg-C]	17 (8 29)	36 (20 58)	56 (35 89)	64 (40 98)
cumulated CH ₄ [Tg-CH ₄]	173 (85 354)	446 (218 921)	818 (410 1753)	1035 (539 2236)
dT (PF) [°C]	0.03 (0.01 0.05)	0.06 (0.03 0.10)	0.10 (0.06 0.15)	0.11 (0.06 0.18)
RCP4.5				
cumulated CO ₂ [Pg-C]	18 (9 32)	54 (28 92)	118 (75 180)	155 (104 216)
cumulated CH ₄ [Tg-CH ₄]	227 (109 466)	1126 (538 2356)	3117 (1657 5969)	4705 (2592 8449)
dT (PF) [°C]	0.03 (0.01 0.05)	0.08 (0.05 0.14)	0.16 (0.10 0.25)	0.19 (0.13 0.29)
RCP6.0				
cumulated CO ₂ [Pg-C]	18 (8 30)	60 (29 101)	156 (103 224)	193 (134 270)
cumulated CH ₄ [Tg-CH ₄]	201 (97 407)	1270 (663 2440)	3104 (1818 5372)	4615 (2664 7778)
dT (PF) [°C]	0.03 (0.01 0.05)	0.08 (0.04 0.13)	0.18 (0.11 0.29)	0.24 (0.16 0.39)
RCP8.5				
cumulated CO ₂ [Pg-C]	20 (9 36)	87 (42 141)	194 (136 270)	228 (157 313)
cumulated CH ₄ [Tg-CH ₄]	333 (154 665)	1474 (836 2614)	3592 (2141 6093)	5877 (3644 9989)
dT (PF) [°C]	0.03 (0.02 0.05)	0.09 (0.05 0.14)	0.14 (0.10 0.21)	0.16 (0.11 0.23)

5
 6
 7
 8

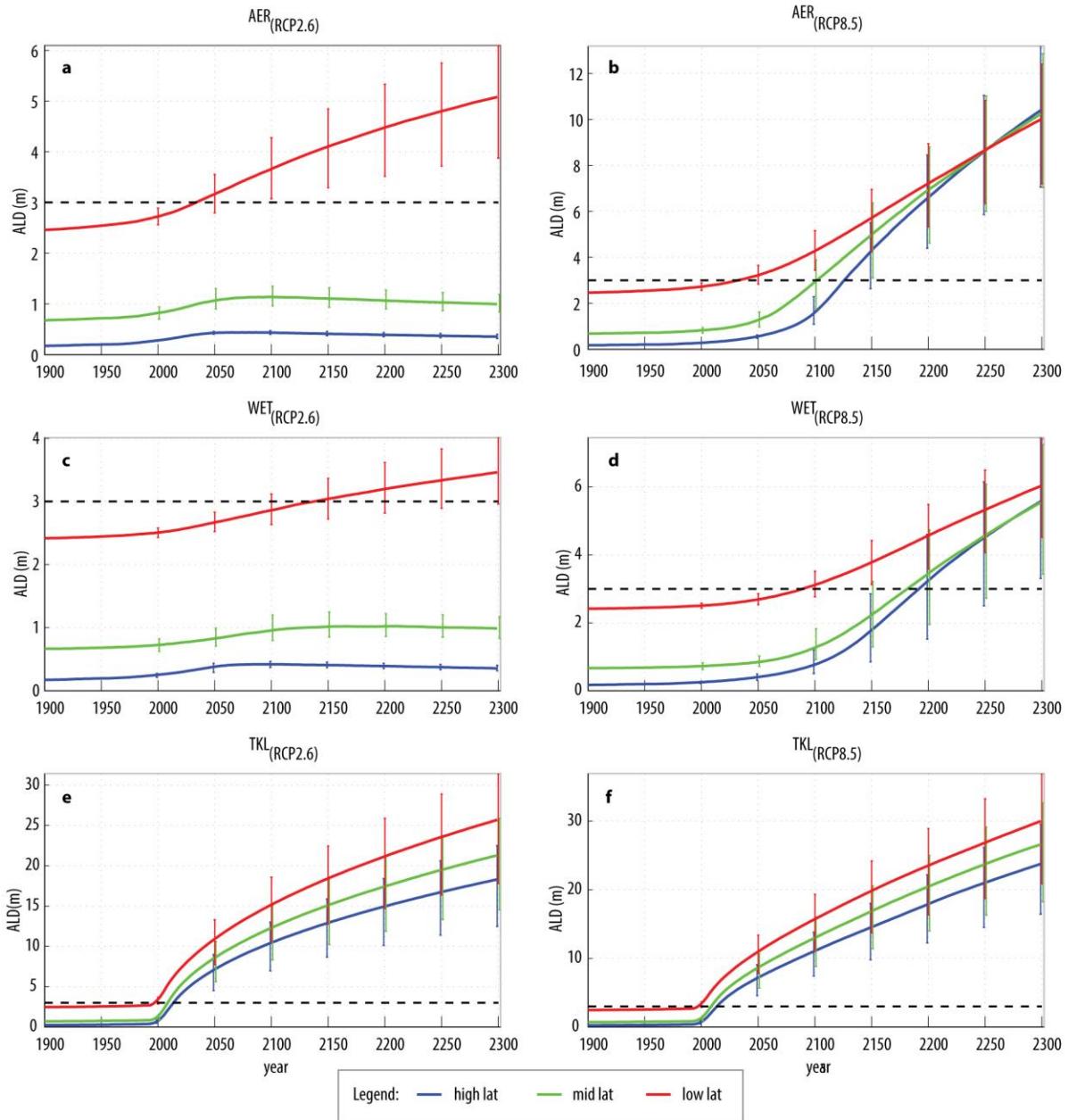
1



2

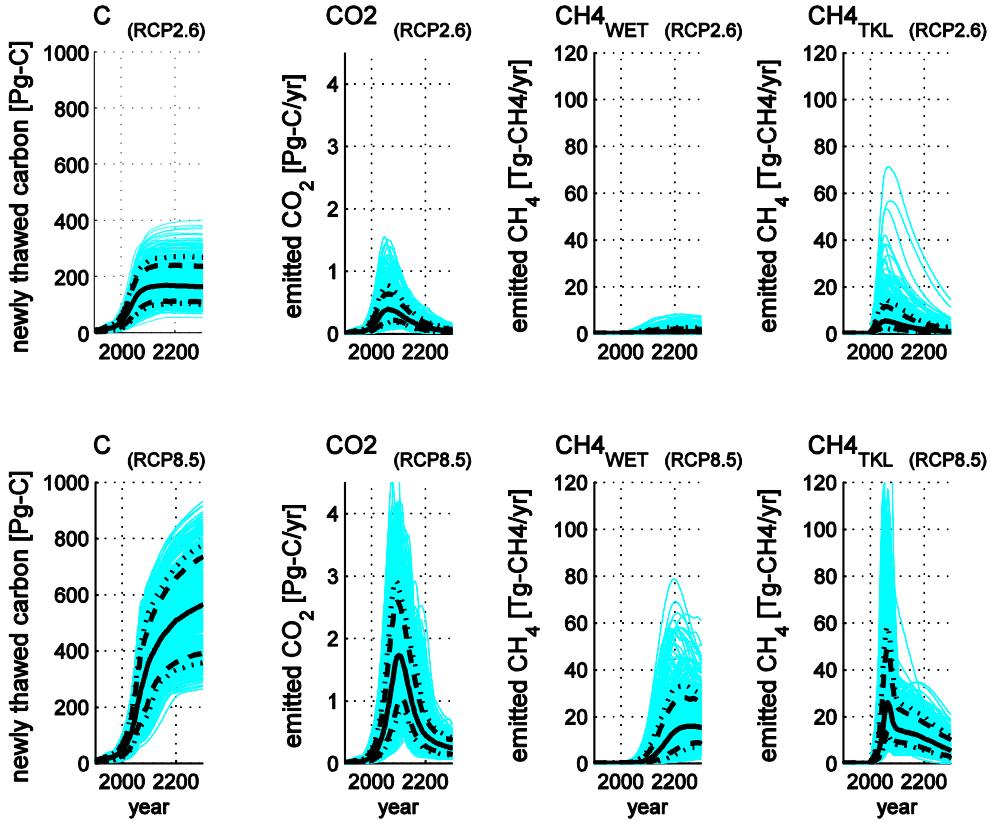
3 Figure 1. Schematic subdivision of permafrost soil carbon stocks into the four main pools
 4 (mineral soils, organic soils, refrozen thermokarst deposits (including taberl), and Yedoma
 5 deposits) and into aerobic (dark yellow) and anaerobic (blue: thermokarst lake, green: wetland)
 6 fractions. Individual boxes indicate the vertical extent and overall soil carbon quantity, as well as
 7 the aerobic and anaerobic fractions (not fully to scale). The dashed lines illustrate the model
 8 resolution into latitudinal bands (only shown for the mineral soil carbon pool) and vertical layers.
 9 Exemplarily, for the mineral soil carbon pool the North-South gradient of active layer depth (red
 10 line) and soil carbon release as CO₂ and CH₄ are also shown (broad arrows). Not shown is the
 11 additional differentiation into a fast and slow pool component.

12

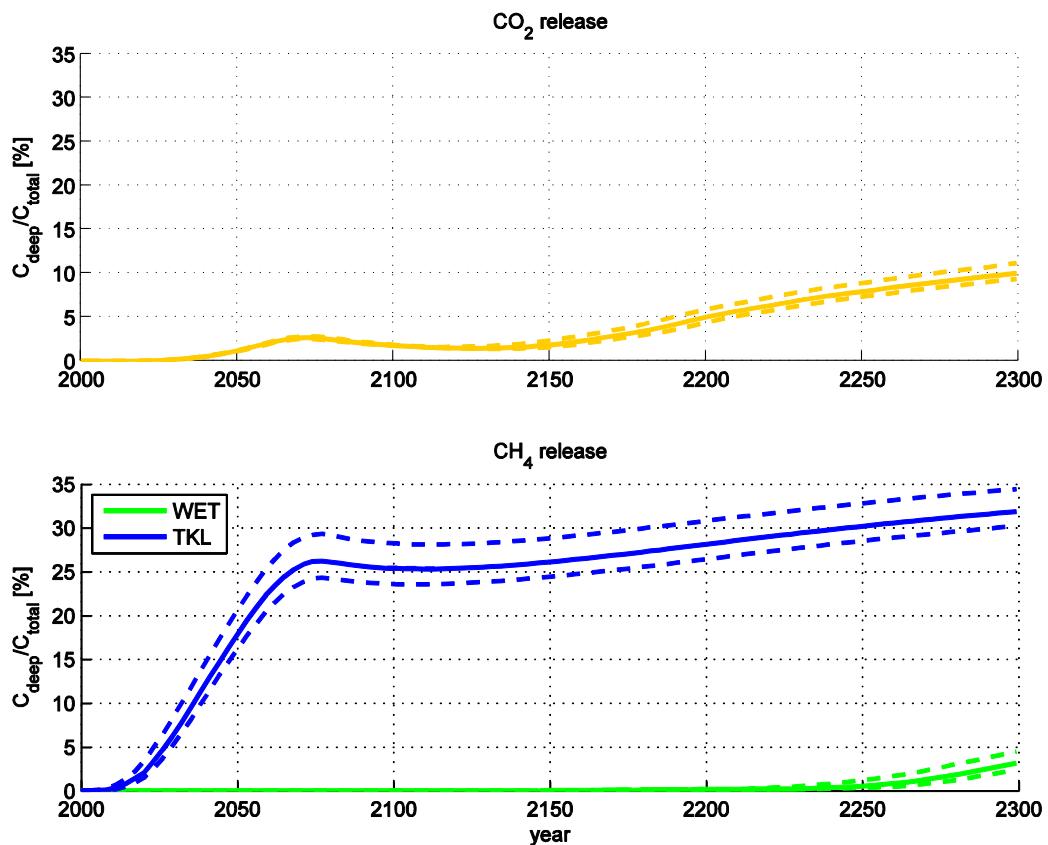


1
2 Figure 2. Simulated changes in active layer depths ALD for mineral soils under moderate
3 (RCP2.6) and extensive (RCP8.5) warming (left and right panels). Shown is the deepening of the
4 active layer from the year 1900 until 2300 for a north-south gradient of different initial
5 permafrost temperatures (blue: $\text{MAGTt0}=-10^\circ\text{C}$, green: $\text{MAGTt0}=-5^\circ\text{C}$, red: $\text{MAGTt0}=-0.5^\circ\text{C}$)
6 and for different hydrologic conditions (a,b: aerobic, c,d: wetland, e,f: thermokarst lake). We
7 assume that newly formed lakes reach the critical depth which prevents winter refreeze by the
8 year 2000. Vertical bars illustrate the model spread inferred from an ensemble of 500 runs (68%

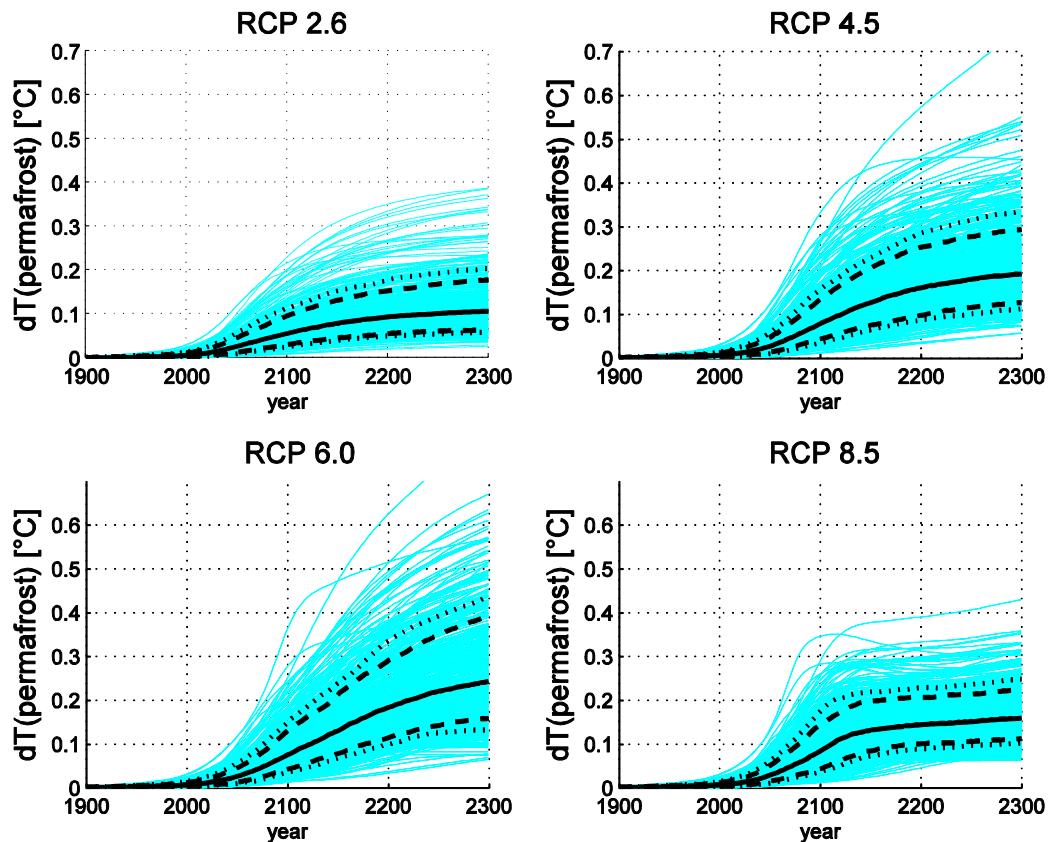
1 range). The horizontal dashed lines denote the near-surface permafrost boundary (3m). Note the
 2 different y-axes scales.



3
 4 Figure 3. Simulated increase in newly thawed permafrost carbon C and resulting rates of annual
 5 CO₂ and CH₄ release under moderate (upper panels) and extensive (lower panels) global
 6 warming for the years 1900 to 2300. CH₄ release is shown separately for fluxes from wetlands
 7 (WET) and newly formed thermokarst lakes (TKL). Blue lines show ensemble simulation results
 8 based on 500 model runs which account for parameter uncertainty. Black lines show statistical
 9 quantiles (solid line: median, dashed lines: 68% range, dotted lines: 80% range). Shown are
 10 contributions aggregated over all individual pools, summed over all latitudes and depths layers.
 11



1
2 Figure 4. Contribution of deep permafrost carbon deposits to total carbon fluxes under aerobic
3 (upper panel) and anaerobic (lower panel) conditions. Shown is the contribution of cumulated
4 CO₂ and CH₄ fluxes from deep deposits (3 to 15m) to total circumarctic carbon release (0 to
5 15m) under strong warming (RCP8.5). Solid lines represent median values, dashed lines 68%
6 ranges. The contribution of deep deposits to wetland-affected CH₄ release (green) and to
7 thermokarst-affected CH₄ release (blue) is shown separately.
8



1

2 Figure 5. Increase in global average surface air temperature through newly thawed permafrost
 3 carbon under various anthropogenic warming scenarios (RCP2.6 to RCP8.5). Blue lines show
 4 ensemble simulation results based on 500 model runs which account for parameter uncertainty.
 5 Black lines show statistical quantiles (solid line: median, dashed lines: 68% range, dotted lines:
 6 80% range). Shown is the temperature feedback as a consequence of CO_2 and CH_4 release from
 7 all individual pools.

8

9