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Impacts of ecological succession and climate warming on permafrost aggradation in drained lake basins of the Tuktoyaktuk Coastlands, Northwest Territories, Canada

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Abstract

Rapidly increasing air temperatures will alter permafrost conditions across the Arctic, but variation in soils, vegetation, snow conditions, and their effects on ground thermal regime complicate prediction across spatial and temporal scales. Processes that result in the emergence of new surfaces (lake drainage, channel migration, isostatic uplift, etc.) provide an opportunity to assess the factors influencing permafrost aggradation and terrain evolution under a warming climate. In this study we describe ground temperatures, vegetation, and snow and soil conditions at six drained lake basins (DLBs) that have exposed new terrain in the Tuktoyaktuk Coastlands in the last 20–100 years. We also use one-dimensional thermal modeling to assess the impact of ecological succession and future climate scenarios on permafrost conditions in historical and future DLBs. Our field observations show that deep snow pack and shallow organic layers at shrub-dominated DLBs promote increased thaw depth and ground temperatures compared to a sedge-dominated DLB and two ancient DLB reference sites. Modeling of past and future drainages shows that climate warming projected under RCP 8.5 will reduce rates of permafrost aggradation and thickness, and drive top-down thaw that could degrade permafrost in shrub-dominated DLBs by the end of the century. Permafrost at sedge-dominated sites was more resilient to warming under RCP 8.5, with the onset of top-down thaw delayed until about 2080. Together, this indicates that the effects of ecological succession on organic soil development and snow drifting will strongly influence the aggradation and resilience of permafrost in DLBs. Our analysis suggests that DLBs and other emergent landscapes will be the first permafrost-free environments to develop under a warming climate in the continuous permafrost zone.

KEYWORDS
climate change, drained lake, permafrost, tundra

1 | INTRODUCTION

Air temperatures in northern Canada have increased at more than double the average global rate.[1,2] Accelerated warming has increased permafrost temperatures and active-layer thickness at sites across the

The data used in this study will be made available through a Northwest Territories Geological Survey Open Data Report.

Arctic,[3–5] causing top-down permafrost thaw, ice wedge degradation,[6–8] increased thaw-driven slope failure,[9–11] and lake drainage.[12–14] In the subarctic, permafrost degradation is causing the collapse of permafrost peat plateaus, and the conversion of spruce woodlands into wetlands and fens.[15,16] Coupled Earth System Models project declines in the global extent of permafrost, but the rate and magnitude of this loss vary widely among models.[17–19] The degradation of permafrost is a transient process with a high degree of spatial variability, and our understanding of the response of permafrost to climate warming remains limited, particularly in areas of continuous permafrost where thermal history, ground ice, latent heat effects, and ecosystem feedbacks have an important influence on rates of degradation.[20] Predicting the impacts of climate warming on permafrost conditions is also made challenging by fine-scale variability in ground temperatures associated with variation in soil moisture, vegetation, and snow conditions.[4,21–23] A significant lag in the response of continuous permafrost to a changing climate should be anticipated.[20,24,25] Conversely, emergent surfaces that have not been preconditioned by a cold Holocene climate and the ecological conditions created by long-term succession (vegetation, organic soils, etc.) are likely to host thin and warm permafrost that will be more responsive to ongoing changes in climate.

The emergence or exposure of new terrain in cold regions where ecological succession, soil development, and the time since exposure result in permafrost conditions distinct from the surrounding landscape offers a unique opportunity to study the factors influencing permafrost aggradation, terrain evolution, and the resilience of permafrost to changing climate conditions. Lake drainage or drying,[26–29] channel migration,[30,31] and isostatic uplift[32,33] are widespread processes that all yield permafrost-free environments where exposure to a cold climate, coupled with biophysical feedbacks, drives permafrost aggradation.[34–36] Past research in these terrain types has shown that factors including substrate properties, snow cover, vegetation, and time since emergence interact with climate to dictate the rate of permafrost aggradation, permafrost thickness, ground-thermal regimes, and ground ice development.[31,32,37,38] Projected climate warming will alter ecological trajectories, rates of ground heat loss, and the process of permafrost aggradation, raising the possibility that emergent environments (in the continuous permafrost zone) will be amongst the first permafrost-free areas to develop in the near future.

To investigate how emergent land surfaces in permafrost regions will evolve under changing climate, in this study we focus on permafrost aggradation after rapid lake drainage, a process that occurs in many areas of continuous permafrost.[26,39–41] Water bodies within the continuous permafrost zone significantly moderate the effect of climate on ground temperatures.[42–44] In lakes and ponds where depth exceeds the maximum thickness of winter ice, lake bottoms are unfrozen.[28,42,45] The geometries of these "taliks" are related to the diameter of the waterbody, lake-bottom and adjacent permafrost temperatures, the thermal properties of earth materials, and the geothermal heat flux.[42] Based on data from Richards Island, Northwest Territories, Canada, where the mean annual temperature of permafrost ranges from −5 to −8°C, Burn[42] estimated that deep circular lakes with a radius of greater than 180 m are likely to be underlain a talik that extends beyond the depth of the adjacent permafrost.

Following drainage and exposure of the lake-bottom, cold climate conditions drive permafrost aggradation and the development of ground ice and periglacial landforms.[26,29,46] In the decades following their emergence, permafrost and biophysical conditions in drained lake basins (DLBs) are relatively dynamic and may affect the rates of permafrost aggradation.[27,35,47] Former lake basins tend to be poorly-drained, and on centennial to millennial timescales, these conditions facilitate organic soil accumulation and the development of polygonal peatlands and wetland complexes.[48,49] These features are stable components of the landscape and can comprise a significant portion of the terrestrial environment.[26,39,50–54] As such, rapidly-drained lake basins provide excellent systems to explore the factors influencing permafrost aggradation in a warming climate and the evolution and resilience of permafrost in emergent landscapes across continuous and extensive discontinuous permafrost zones. In this study, we describe the surface conditions (soil, vegetation, snow, and ground temperature) at six lake basins that drained between 20 and ~100 years prior to our fieldwork. We also use thermal modeling to explore the impact of ecological succession and future climate scenarios on permafrost aggradation and its resilience following lake drainage.

2 | METHODS

2.1 | Study area

This study focusses on the Tuktoyaktuk Coastlands, in the northern Northwest Territories, Canada (Figure 1). This upland area east of the Mackenzie Delta is underlain by Pleistocene sediments characterized by ice-rich glacial till and ice-contact deposits.[53] The landscape is predominantly hummocky uplands with mineral soils,[56] but organic deposits are also common in low low-lying areas commonly associated with lacustrine basins.[57,58] Vegetation in the southern part of this area is characterized by the transition from open spruce woodlands to a landscape dominated by tundra.[59] Moving northward across this region, upright shrub tundra dominated by green alder, dwarf birch, and willows gradually shifts to graminoid and dwarf shrub-dominated tundra.[60,61] The climate of the Tuktoyakuk Coastal Plain is cold and terrestrial surfaces are underlain by continuous permafrost, which can be up to 500 m thick.[62] Average annual air temperatures at Inuvik and Tuktoyaktuk are −8.2 and −10.1°C, and summers are short and cool, with average air temperatures of 12.2 and 8.8°C, respectively.[63] At undisturbed locations across the study area, the mean annual temperature at the top of the permafrost ranges from −2 to −7°C in response to variation in soil moisture, vegetation, and snow conditions.[4] A composite climate record from Inuvik and Aklavik (55 km west of Inuvik) demonstrates that average annual air temperatures have increased by 3.1°C since 1926.[64] This shift in climate has driven an increase in permafrost temperatures of ~2°C.[4]

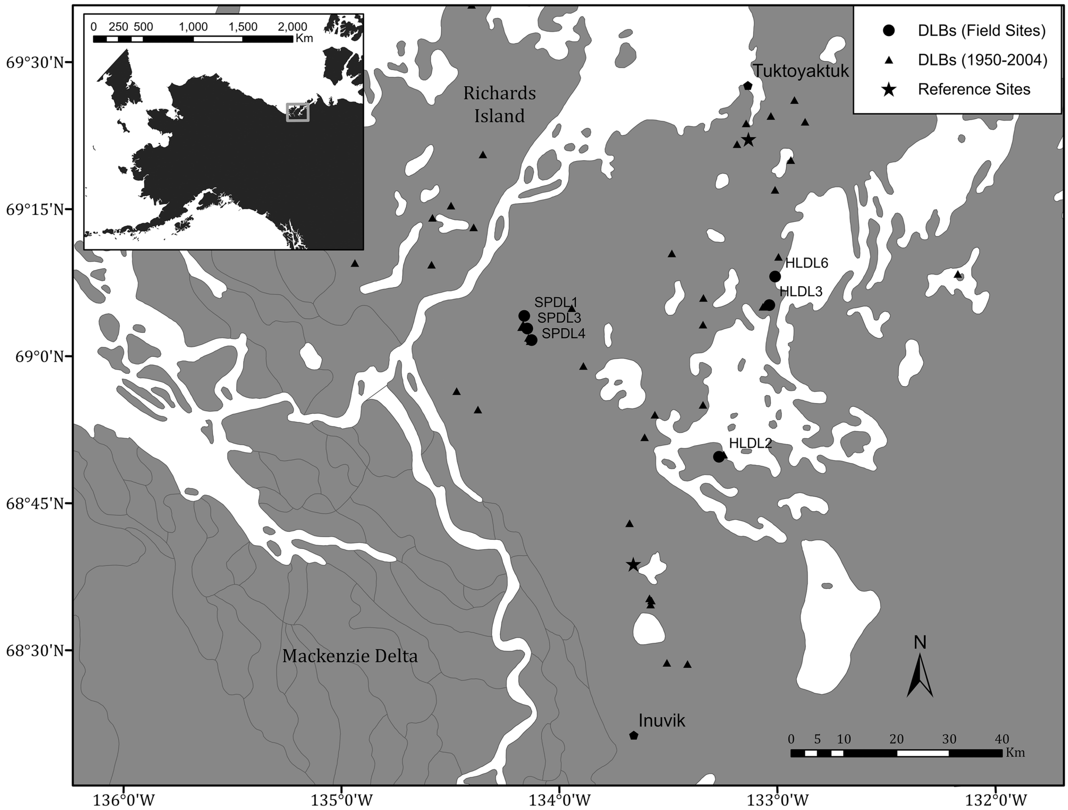

FIGURE 1 Map of the study area in the Tuktoyaktuk Coastlands showing drained lake basins (DLBs) and reference sites sampled in the field and lake basins that drained between 1950 and 2000 across the entire study area (see[55]). Inset map in the upper left corner shows the extent of the main map in northwestern North America

TABLE 1 Average mean annual ground temperature at 1-m depth (MAGT), thaw depth, soil moisture, organic soil thickness, late winter snow depth, and vegetation height in the six drained lake basins (DLBs) two reference sites. The first column shows the site name and the date of rapid lake drainage. The location of each DLB is also shown in Figure 1. Estimates of mean vegetation cover shown in bold type were calculated based on data from the 38 DLBs shown in Figure 1. HLDL and SPDL are abbreviations for Husky Lakes (Imaryuk) Drained Lake and Swimming Point Drained Lake, respectively

Drained Lake Basin (drainage date)	MAGT (°C)	Thaw depth (cm)	Soil moisture (%)	Organic thickness (cm)	Snow depth (cm)	Canopy height (cm)	Shrub cover (%)	Herb cover (%)	Sedge wetland cover (%)
HLDL3 (1995)	0.80	99.7	42.4	1.5	59.5	NA	70	25	5
SPDL4 (1950–70)	−0.11	85.6	64.1	3.4	88.9	36.5	60	25	15
HLDL2 (1950–70)	0.09	94.6	45.1	3.3	93.1	56.0	35	60	5
SPDL1 (1950–70)	−0.10	66.5	55.3	4.2	77.5	95.0	70	25	5
SPDL3 (1950–70)	0.44	92.2	47.4	2.2	155.8	40.7	70	25	5
HLDL6 (Pre 1930)	−2.75	45.6	52.1	21.5	43.8	23.7	15	30	55
Reference site				**Study area mean (*n* = 38)**			**48**	**37**	**15**
Tuktoyaktuk	−4.73	34.0	31.6	126.3	31.7	14.3			
Jimmy Lake	−4.42	40.5	28.2	>30	47.0	21.0			

2.2 | Site selection

In the Tuktoyatuk Coastlands, rapid lake drainage typically occurs when thermomechanical erosion degrades near-surface ground ice.[40] This results in complete or partial drainage, which typically occurs quickly (i.e., hours to days) and creates conditions conducive to the aggradation of permafrost.[35,55,65] We used historical air photos and field reconnaissance to select six drained lakes and two reference sites in the study area (Figure 1; Table 1). Drainage dates were estimated using georeferenced air photos from 1930, 1950, and 1972 and Landsat imagery (1973–2010) from the USGS archive (Figure 2). All of the study basins selected have outlet channels visible on aerial

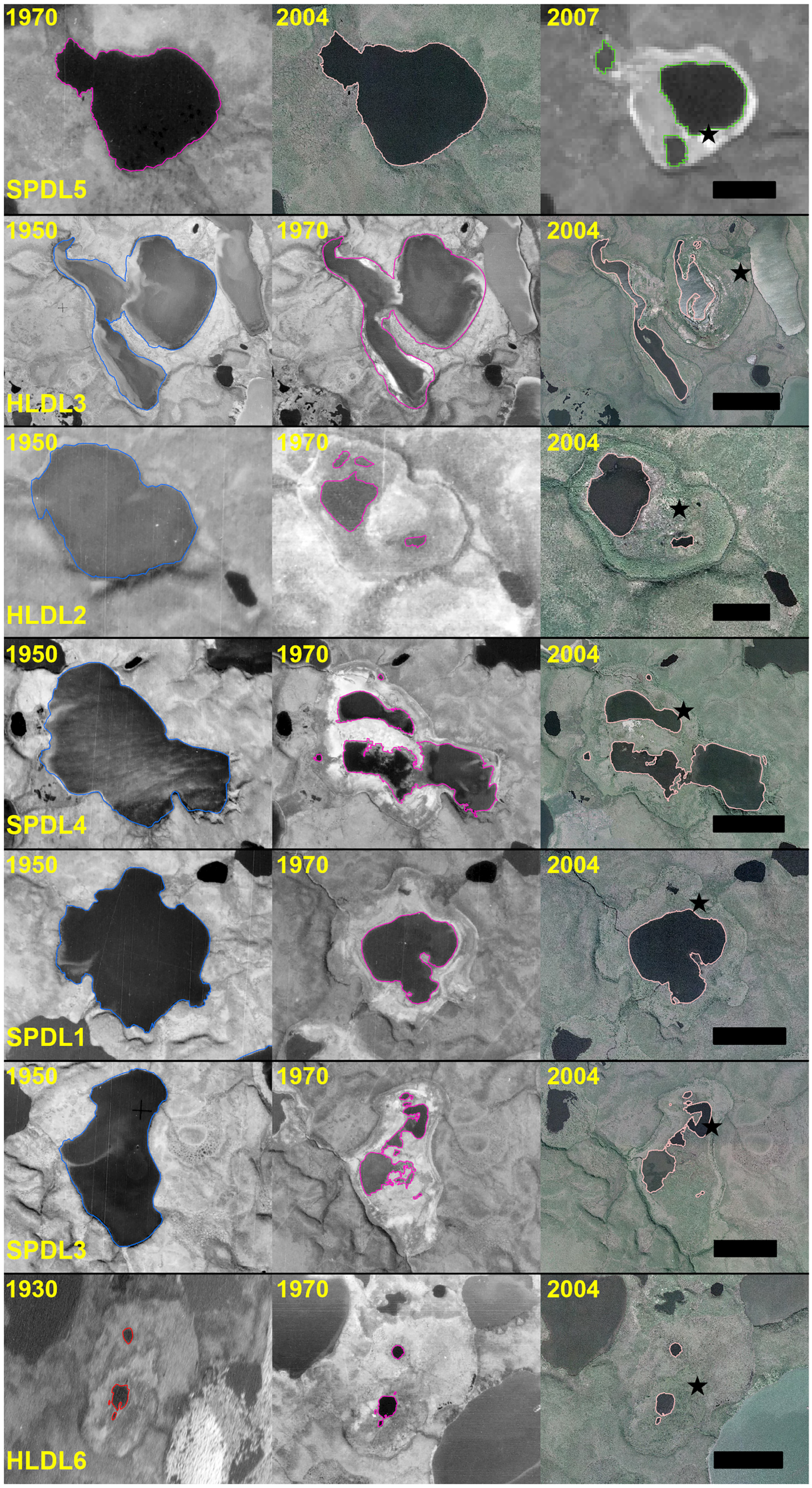

FIGURE 2 Airphotos of the six drained lakes showing the extent and approximate timing of drainage. The scale bar on each group of photos represents 400 m. Colored outlines show the extent of surface water at each time point. Thermistor locations are shown as stars on the most recent image in the column on the right [Colour figure can be viewed at wileyonlinelibrary.com]

imagery that are indicative of rapid drainage. Data from the ArcticDEM,[66] coupled with georeferenced air photos of the lakeshore prior to drainage, indicate that maximum lake depth ranged from 3 to 8 m. Prior to drainage, these lakes had minimum diameters ranging from 400 to 1500 m. Together, these observations indicate that all of the DLBs we studied were all underlain by a through-going talik.[28,42] These data also show that our study lakes did not have littoral terraces, which are common in some parts of the study area and are typically underlain by permafrost.[42] Two reference sites were selected in areas of high centered polygonal terrain ~9 km south of Tuktoyaktuk and ~32 km north of Inuvik, respectively (Figure 1). These peatland sites are located in ancient lake basins that probably drained sometime during the mid-Holocene.[48,49]

2.3 | Ground temperatures

To measure ground temperatures at our sites we drilled a shallow borehole (1 m) at each site and attached thermistors to PVC pipes, which were installed in the boreholes to position thermistors at depths of 0.1 and 1.0 m. We used Hobo Pro U23–003 data loggers with our thermistors, which had an accuracy and precision of ±0.21 and 0.02°C, respectively. Loggers were set to record measurements every 2 h from August 2013 to August 2018. In some instances, PVC pipes were subjected to wintertime frost heave of 5–15 cm and had to be adjusted to reposition the thermistors. In these instances, data (at the surface and 1-m depth) from the start of freezeback to the time of repositioning the following year were not included in our analyses.

2.4 | Site characterization

At each DLB or reference site we used transects, grids, and points to measure a suite of biotic and abiotic variables. We measured organic soil thickness at eight locations in each DLB using a small shovel to expose the upper soil horizons. Organic thickness at the Tuktoyaktuk reference site was measured using a gas powered coring system to drill to the base of the peat deposit at three locations. At each site we made 10 measurements of late season thaw depth by pushing a graduated steel probe to the depth of refusal. Volumetric soil moisture (%) was estimated using a handheld moisture probe (HH2 Moisture Meter with a ML2x sensor) from Delta-T Devices to take three readings at each site. Measurements of thaw depth and soil moisture were made from August 13 to 15, 2017, within a 2-m radius of each thermistor. On April 23, 2017 we measured late winter snow depth at all sites using a graduated avalanche probe. Snow measurements were made every 0.5 m along a 25-m transect centered on each thermistor ($n = 50$). We measured vegetation height at each site using a 30 × 30-m grid with sampling points every 2 m ($n = 144$). At each grid point we measured the maximum height of the vegetation and recorded the functional groups present (tall shrubs, dwarf shrubs, forbs, grasses, sedges, mosses, lichen). At the reference sites, all measurements were made within polygon centers. To characterize variability in vegetation structure across all of the lakes in the study area that drained between 1950 and 2000, we used plot-based field surveys and high-resolution drone surveys completed in August, 2018 to determine the cover types that could be reliably mapped using the Worldview images available in ArcPro (v2.6). These cover classes included upright shrub (*Salix* spp., *Alnus viridis*, and *Betula glandulosa*), herbaceous (*Arctagrostis* spp., *Calamagrostis* spp., *Epilobium* spp., moss, etc.), sedge (*Carex aquatilis* and *Eriophorum* spp.), and bare ground. Subsequently, we used Worldview images to estimate their percentage cover in 38 DLBs across the study area, including the six DLBs we visited in the field (Table 1). To compare biotic and abiotic parameters we used one-way ANOVAs and Tukey-adjusted pairwise comparisons to test for statistically significant differences among sites.[67]

2.5 | Permafrost modeling

2.5.1 | Overview

We used historical and projected climate data and known soil properties to model the aggradation of permafrost in a newly exposed DLB using the Northern Ecosystem Soil Temperature (NEST) model.[68] NEST has been used to model permafrost conditions at Arctic and subarctic sites across a range of spatial scales.[25,69,70] It is a one-dimensional process-based permafrost model that considers the effects of climate, vegetation, snow, and soil conditions on ground thermal dynamics based on energy and water transfer through the soil–vegetation–atmosphere system. Ground temperature is calculated by solving the one-dimensional heat conduction equation. The upper boundary condition is determined based on the energy balance of the ground surface or snow surface if snow exists. The lower boundary condition is defined by geothermal heat flux at a depth of 120 m. NEST also integrates dynamics in snow depth, snow density, soil moisture, phase change of soil water, and their effects on ground temperature.[68]

Our approach to modeling utilized historical and projected climate data and included three ecological succession scenarios (dense shrubland, open shrubland, and sedge meadow). These scenarios reflected the range of ecological conditions in DLBs across the study area, and allowed us to explore the resilience of aggrading permafrost to future climate. First, we verified that the NEST model could capture ground thermal dynamics at DLBs by calibrating the model for the six sites we studied in the field. This process used measurements of organic layer thickness, vegetation structure, ground temperatures, and snow conditions made in the field (Table 1) and is described in more detail in the supplementary methods section and Table S1. The correspondence between modeled and observed near-surface ground temperatures that resulted from this process are shown in Figures S3 and S4.

Following model calibration with historical climate data, we simulated three typical postdrainage ecological succession scenarios: (a) dense shrubland, (b) open shrubland, and (c) sedge meadow. In each scenario, snow and vegetation conditions and organic soil development were parameterized to represent decadal-scale changes that accompany ecological succession. In the dense and open shrubland scenarios, tall vegetation increased snow trapping, and organic soil development was slow (Table 2). In the sedge meadow scenario, snow pack was comparatively lower and wet soils promoted rapid organic soil development. These parameters are described in more detail in the next section. All three scenarios simulated lake drainage in 1950 and were run until 2100 using climate projections under Representative Concentration Pathways (RCP) 4.5 and 8.5 (Figure S1). To assess the influence of drainage date in the context of a changing climate, we ran the model for the dense shrubland and sedge meadow scenarios using five drainage years (1900, 1960, 1995, 2030, and 2060). In all simulations we ran the model from the time of drainage until 2100 under two climate change projections, RCP 4.5 and 8.5. To assess the influence of ground ice and its latent heat effects on the rates of permafrost aggradation and degradation, we also simulated the effect of

TABLE 2 Parameters used in model scenarios of permafrost development in drained lake basins. In all drained lake basins the snow drifting parameter was set to decrease linearly from 0.9 to 0.61 between 1951 and 1960. Subsequently, the snow drifting parameter was decreased at a linear rate so it reached the value shown in the table in 2005. After 2005, the snow drifting parameter was held constant until 2100. The rate of organic soil development was held constant for the entire model run (1950–2100). LAI is the abbreviation for leaf area index

Model scenario	Snow drifting parameter[a]	Maximum summer LAI[a]	Maximum plant height (m)[a]	Rate of organic layer development (cm/yr)
Dense shrubland	0.0	2	1.0	0.06
Open shrubland	−0.5	1	1.0	0.06
Sedge meadow	−1.0	1	0.2	0.2

[a]Values from 2005 to 2100.

excess ice in the upper soil column (1.5–4 m) by increasing the porosity of the soil to reflect 0, 10, 20, and 40% excess ice content.

2.5.2 | Initial conditions in the DLB

Based on the size and predrainage depth of our study lakes, we developed model scenarios, which assumed that lakes had through-taliks prior to drainage. The postdrainage bathymetry of the DLBs we studied indicates that lake depths would have exceeded the maximum lake ice thicknesses measured in this region from 2000 to 2017.[71] The diameters of the DLBs we studied were greater (400–1500 m) than the threshold for through-going taliks (360 m) estimated by Burn[42] for lakes in the study area. The initial ground temperature was determined using the approach described by Mackay[72] and Burn.[42] In this calculation we assumed that lakes are circular with a diameter of 800 m, are influenced by geothermal heat flux of 0.05 W/m^2,[73] and have mean annual lake bottom and surrounding ground temperatures are 3 and −7°C, respectively.[43] The estimated initial ground temperature decreased from 3.0 to 2.1°C from the top of the talik to a depth of 120 m. In this analysis we focused on near-surface conditions (0–60 m) at the center of a DLB, and neglected the potential role of lateral heat transfer.

2.5.3 | Snow drifting, vegetation, and organic soil development

To integrate the effects of snow drifting due to wind, NEST includes a snow drifting parameter to modify the input of snowfall.[74] Negative values of this parameter indicate that a site receives wind-transported snow from its surroundings. For example, a value of −1.0 indicates that drifting snow input effectively doubles the snowfall from precipitation. In this study we calibrated the snow drifting parameter to reflect observations made by Mackay[35] and field observations we made in 2017 (Table 2). Based on observations at Illisarvik (an artificially drained lake about 80 km northwest of our study area[35]), we set the snow drifting parameter in 1950 (the first year following drainage) at 0.9 (90% of the snow blows away) for all three scenarios.[35] Between 1951 and 1960, we decreased the snow drifting parameter from 0.9 to 0.6 to emulate an increase in snow trapping by early successional vegetation.[35] Subsequently, we decreased the snow drifting parameter at a linear rate, which corresponded to increasing snow capture as a result of the vegetation that developed in each scenario from 1960 to 2005 (Table 2).

In the dense shrubland scenario we decreased the snow drifting parameter from 0.6 to −1.0 from 1960 to 2005. In the open shrubland scenario we decreased the snow drifting parameter from 0.6 to −0.5 from 1960 to 2005. In the sedge meadow scenario we decreased the snow drifting parameter from 0.6 to 0 from 1960 to 2005. In all scenarios we assumed that plant height and summer leaf area index (LAI) increased linearly in the first two decades following drainage until they reached two-thirds of the maximum values shown in Table 2. After 1971, we reduced the rate of increase in these parameters so that they reached their maximum values in 2005. After 2005, we assumed that vegetation remained stable for the rest of the simulation and held the snow drifting parameter constant (Table 2). Modeled snow density is a function of both time and depth of snow. It is a dynamic process related to fresh snow density, compaction, and metamorphosis.[68] The model explicitly considers the redistribution of snow due to wind, which results in large differences in snow depth and the ground thermal conditions in tundra landscape. The model does not consider the effect of wind on snow density. In both shrub scenarios we set the organic accumulation rate at 0.06 cm/year. In the sedge meadow scenario we set the organic accumulation at 0.2 cm/year. These values were obtained from organic accumulation rates estimated in the six DLBs we visited in the field (organic thickness/estimated time since drainage). The assumed rate of organic matter accumulation in our shrub-dominated scenarios (0.06 cm/year) is slightly higher than rates reported for drained lakes on the Alaska North Slope, which ranged from 0.009 to 0.047 cm/year.[26,75] The rate of organic soil development we used in the sedge wetland scenario (0.2 cm/year) is similar to rates of organic matter accumulation reported for productive peatlands.[76,77] Soils in the region are typically fine-grained tills and glaciofluvial deposits with some sandier sediments. In all models we assumed that organic soil was underlain by sandy clay loam (sand 60%, silt 13%, clay 27%) which extended to 40 m and was underlain by bedrock.

2.5.4 | Climate data

Climate data used to drive the model were generated by Met1km, a long-term 1-km-resolution daily meteorological dataset,[78] except between 2014 and 2019, during which air temperature and precipitation were estimated based on observations at the Inuvik climate station (Supplemental Methods). Met1km integrates five coarser gridded meteorological datasets, including two future emission scenarios (RCP 4.5 and 8.5). These datasets were downscaled to 1-km resolution using the rebaselining method[79] based on 1-km-resolution monthly averages from WorldClim2.[80] The data generated include daily minimum and maximum air temperature, precipitation, vapor pressure, solar radiation, downward longwave radiation, and wind speed from 1901 to 2100. The time-series meteorological data that we used are based on the grid cell that includes HLDL3, a DLB in the middle of our study area. Under RCP 8.5 the average annual temperature increases by 1.0°C per decade (based on the slope of the linear trend) between 2020 and 2100, which is similar to the observed rate of 0.8°C per decade from 1970 to 2020 (Figure S1). Under RCP 4.5 the average rate of temperature increase from 2020 to 2100 is 0.3°C per decade (Figure S1). Total annual precipitation is highly variable under both RCP 4.5 and 8.5, but shows a gradual increase over time with rates of 2.8 and 13.2 mm per decade between 2020 and 2100, respectively (Figure S1). To contextualize average monthly temperatures projected by climate change scenarios RCP 4.5 and 8.5 with the broad-scale climate gradient in the Northwest Territories, they are also plotted against climate normals (1981–2010) for Tuktoyaktuk, Inuvik, and Normal Wells (Figure S2).

3 | RESULTS

3.1 | Field measurements

The DLBs that we sampled were physically, ecologically, and thermally distinct from the surrounding tundra. The vegetation in the basins, which drained between 1950 and 1995, was dominated by dwarf birch and willows and had a canopy that was significantly taller than the reference site (Figure 3). Mapping conducted using high-resolution satellite imagery showed that all of the DLBs in the study area that formed between 1950 and 2000 were dominated by a mix of upright shrub and herbaceous vegetation and had only a sparse cover of sedge wetland (Table 1). The basin that drained prior to 1930 was dominated by sedges and short willows. The canopy height at this site was significantly lower than at most other drained basins, but was similar to the reference site (Figure 3a). Vegetation in the ancient lake basins that we used as reference sites consisted of dwarf shrub and tussock tundra underlain by thick peat deposits.[81]

All five DLBs that drained between 1950 and 1995 had a thinner organic layer than the reference site. These five DLBs had similar organic soil thickness (1.5–4.2 cm), which were significantly less than the lake that drained prior to 1930 (21.5 cm; Figure 3b). Late winter snow depth was also greater at drained lakes compared to the reference site. Lakes that drained between 1950 and 1995 had a snow depth that was 28–124 cm greater than at the reference site, whereas mean snow depth at the basin that drained prior to 1930 was only 12 cm greater than at the reference site (Figure 3c).

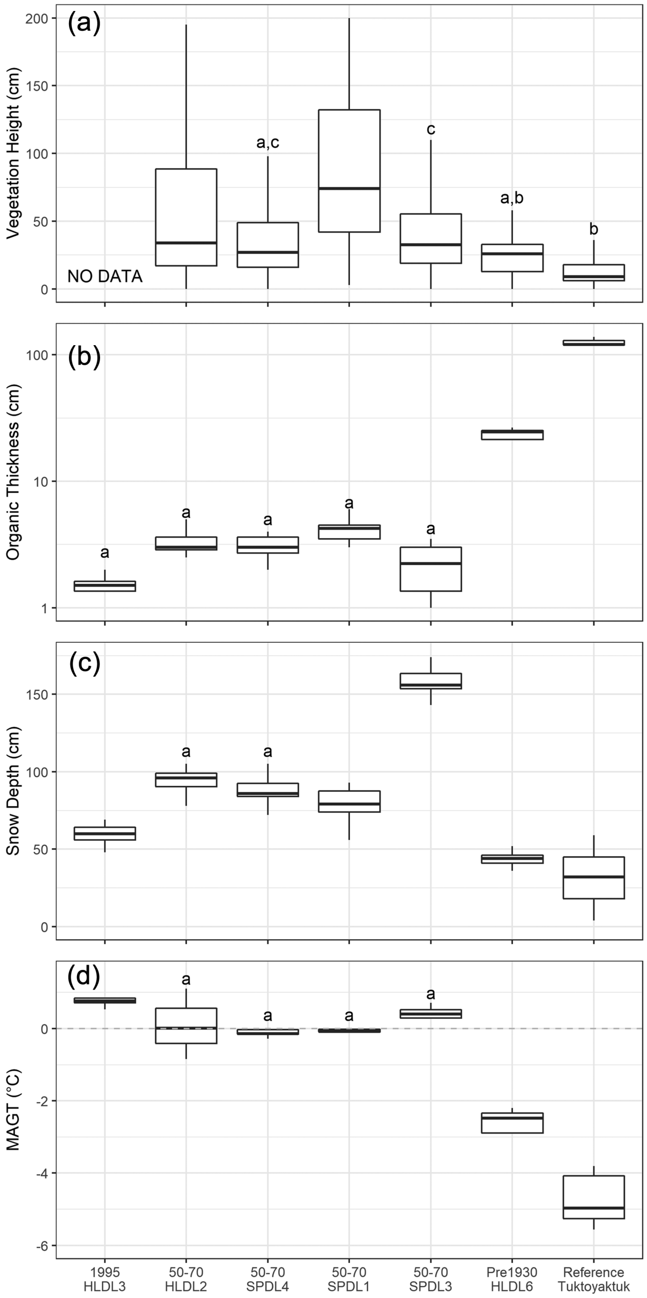

FIGURE 3 Physical conditions including: (a) vegetation height, (b) organic soil thickness, (c) late winter snow depth, and (d) mean annual ground temperature (1 m) at six drained lakes and one reference site in the Tuktoyaktuk coastlands. The solid black line inside each box represents the median value, the ends of the boxes represent the 25th and 75th percentiles, and the whiskers show the 10th and 90th percentiles. Bars marked with the same letter are not significantly different from each other and unlabeled bars and bars marked with different letters are significantly different from each other

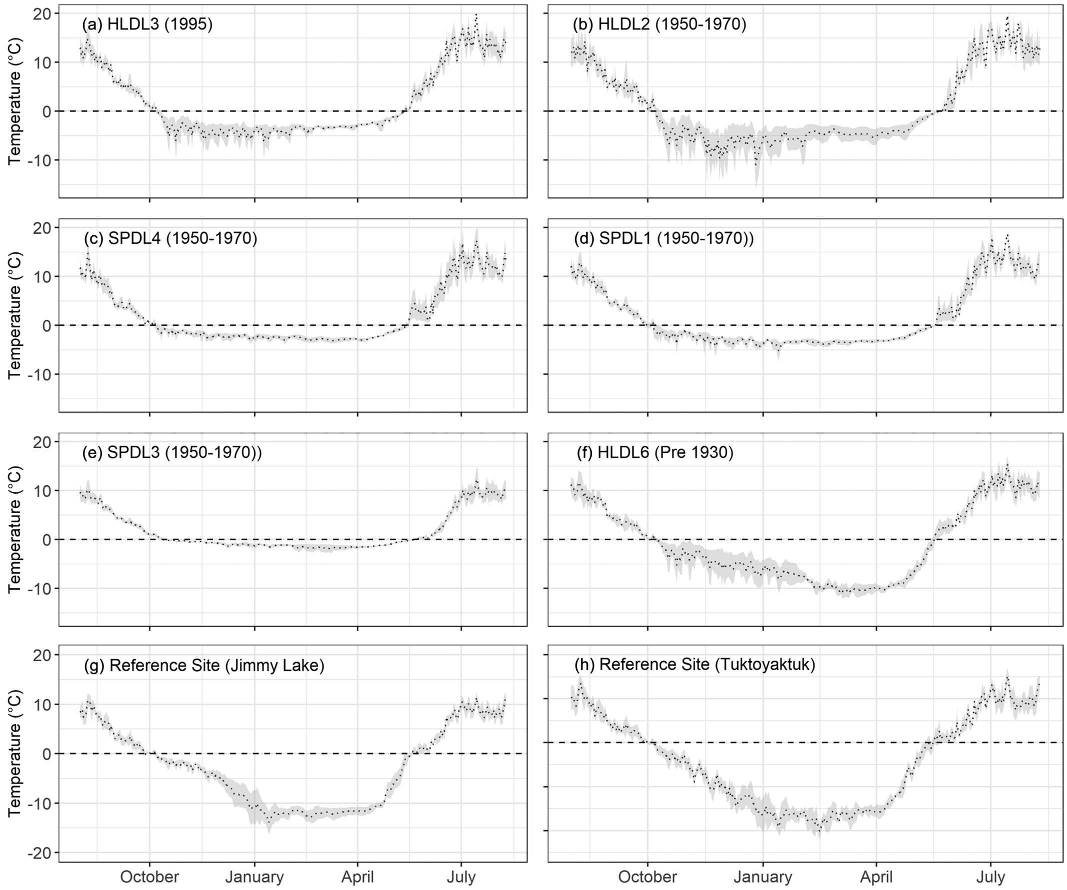

FIGURE 4 Temperature at 0.1 m below the ground surface at six drained lakes and two reference sites in the Tuktoyaktuk Coastlands. The dotted line shows the mean daily temperature from 2013 to 2017 and the shaded area represents the 95% confidence interval of the daily mean. HLDL and SPDL are abbreviations for Husky Lakes Drained Lake and Swimming Point Drained Lake, respectively. Note the changes in the *y*-axis scale between (a)–(f) and (g–h)

Differences in ground surface temperatures and at 1-m depth reflected contrasts in biophysical characteristics of the sites, with all drained basins exhibiting higher ground temperatures than the reference sites. Ground-surface temperatures (0.1 m below the surface) in lakes that drained between 1950 and 1995 were elevated compared to both reference sites and the sedge-dominated DLB (HLDL6), and remained at the zero curtain throughout the winter (Figure 4). Ground-surface temperatures at the sedge-dominated DLB were similar to our southern reference site near Jimmy Lake (Figure 4). Near-surface ground temperatures in the summer were similar among all six sites (Figure 4). Mean annual ground temperature (MAGT) at 1-m depth in recently drained basins ranged between −0.11 and 0.8°C, and these basins were all significantly warmer than the oldest drained lake (HLDL6) with MAGT of −2.75°C, which was closer to the MAGT at the reference site (Figure 4d). Average daily temperatures at 1 m below the ground surface (Figure 5) indicate that active layer thickness was greater than 1 m at all of the shrub-dominated lakes we sampled. Active-layer freezeback at these sites was variable, and completed between December 8 and March 12, or not at all (Figure 5). The sedge-dominated lake, which drained prior to 1930 (HLDL6), was the only basin with temperatures consistently below 0°C at 1-m depth (Figure 5).

3.2 | Permafrost modeling

In the thermal modeling scenarios shown in Figures 6 and 7, permafrost began to establish in lake bottoms immediately after drainage in 1950, but subsequent conditions varied in response to interactions between climate warming and the impacts of vegetation succession on snow depth and organic soil development (Figure 6). In the dense and open shrubland scenarios (deep snow and slow organic development), permafrost aggraded to the maximum depths of 12.8 and 14.3 m 42 and 63 years following drainage, respectively (Figure 6a,b). In the sedge meadow scenario (shallow snow and rapid organic soil development), modeled permafrost depth increased to 20.3 m by 2100 under climate scenario RCP 4.5. Due to a similar climate under RCP 8.5 (until about 2060), modeled permafrost depth was 20.2 m by 2100 under this scenario (Figure 6c). Modeled permafrost aggradation in the first three decades after the drainage was comparable with

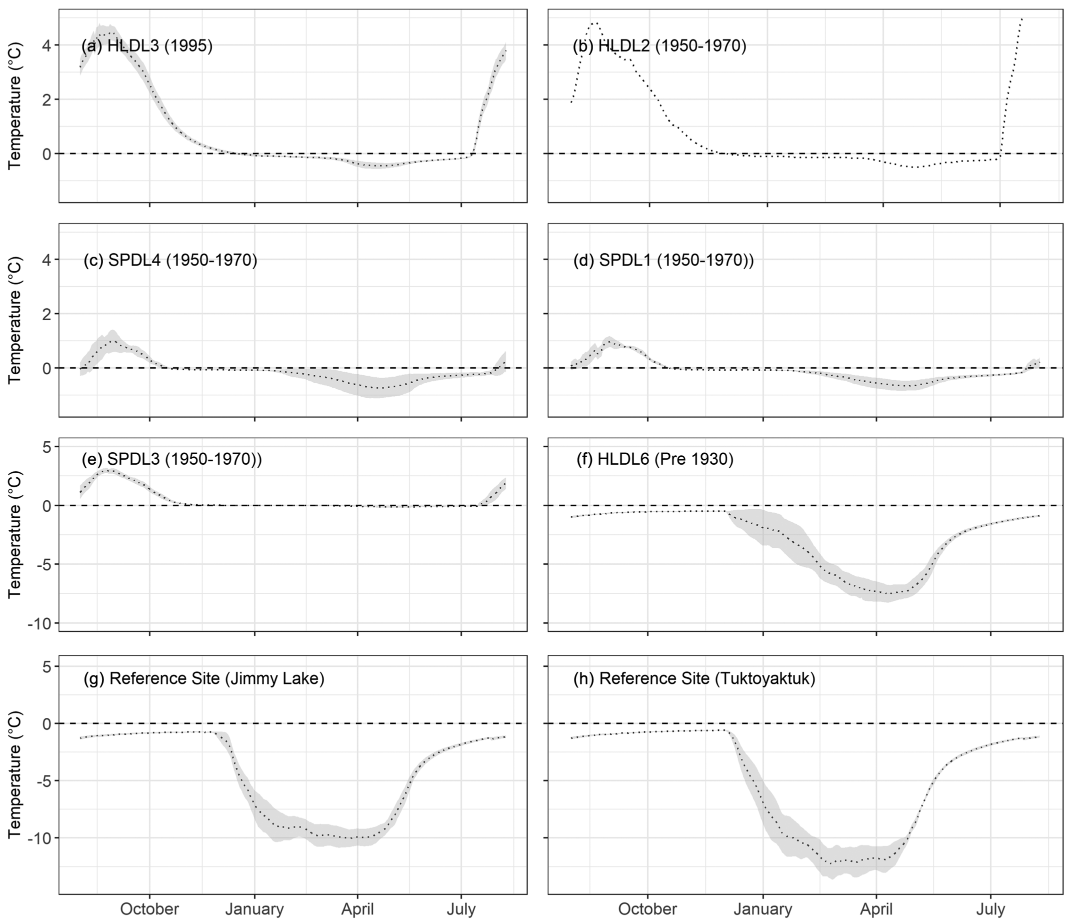

FIGURE 5 Temperature at 1 m below the ground surface at six drained lakes and two reference sites in the Tuktoyaktuk coastlands. The dotted line shows the mean daily temperature from 2013 to 2017 and the shaded area represents the 95% confidence interval of the daily mean. HLDL and SPDL are abbreviations for Husky Lakes Drained Lake and Swimming Point Drained Lake, respectively. Note the changes in the *y*-axis scale between (a)–(d), (e–f), and (g–h)

observations reported for Illisarvik, where a lake was experimentally drained in 1978.[35] The freezing front at Illisarvik was 5.7, 13.5, 13.8, and 14.3 m in 1981, 1998, 2001, and 2005, respectively, corresponding to 3, 20, 23, and 27 years after drainage in 1978.[82] The modeled ranges of permafrost depths in our three scenarios were 4.6–5.4, 11.9–12.3, 11.9–13.4, and 12.0–13.7 m in 3, 20, 23, and 27 years after the drainage, respectively, with the sedge meadow scenario developing greater permafrost thicknesses than shrubland scenarios.

In the dense shrub scenario (deep snow and slow organic soil development), maximum active layer thicknesses ranged from 1.2 to 1.3 m in the first four decades after drainage. Top-down permafrost thaw associated with a warming climate began in this scenario around 2000 and, by 2097, permafrost was completely degraded under RCP 8.5, and only persisted below 8.1 m under RCP 4.5 (Figure 6a). Modeling showed that top-down thaw proceeded slowly at first, but then increased rapidly when near-surface taliks began to form (Figure 6). This is similar to the modeled permafrost degradation pattern with climate warming in the Hudson Bay Lowlands.[83] In the open shrubland scenario, the depth to the top of the permafrost ranged from 1.1 to 1.4 m in the first 50 years following drainage. In this scenario, top-down permafrost degradation associated with a warming climate began around 2020. Under RCP 4.5 permafrost persisted at 8-m depth below a near-surface talik at the end of the simulation, but thawing progressed to a depth of more than 10.5 m under RCP 8.5. The depth to the permafrost table in the sedge meadow scenario gradually decreased from 1.5 to 0.8 m over the first 50 years after drainage with increases in vegetation cover and organic soil development. Active layer thickness remained relatively stable until the end of simulation under RCP 4.5, but began to increase steadily starting in 2080 under RCP 8.5 (Figure 6c).

The influence of biophysical conditions on ground thermal sensitivity to climate warming was also evident in ground temperatures at 1 m at different periods during each ecological succession scenario (Figure 7). In the dense shrub scenario, maximum summer temperature at 1-m depth increased by ~2°C and 5.5°C from 2010–2090 under RCP 4.5 and 8.5, respectively (Figure 7a,b). In the open shrubland scenario, maximum summer temperature at 1-m depth increased by 3 and 6.5°C from 2010–2090 under RCP 4.5 and 8.5, respectively (Figure 7c,d). In the dense shrubland scenarios, temperatures throughout winter remained isothermal by 2010, indicating that the

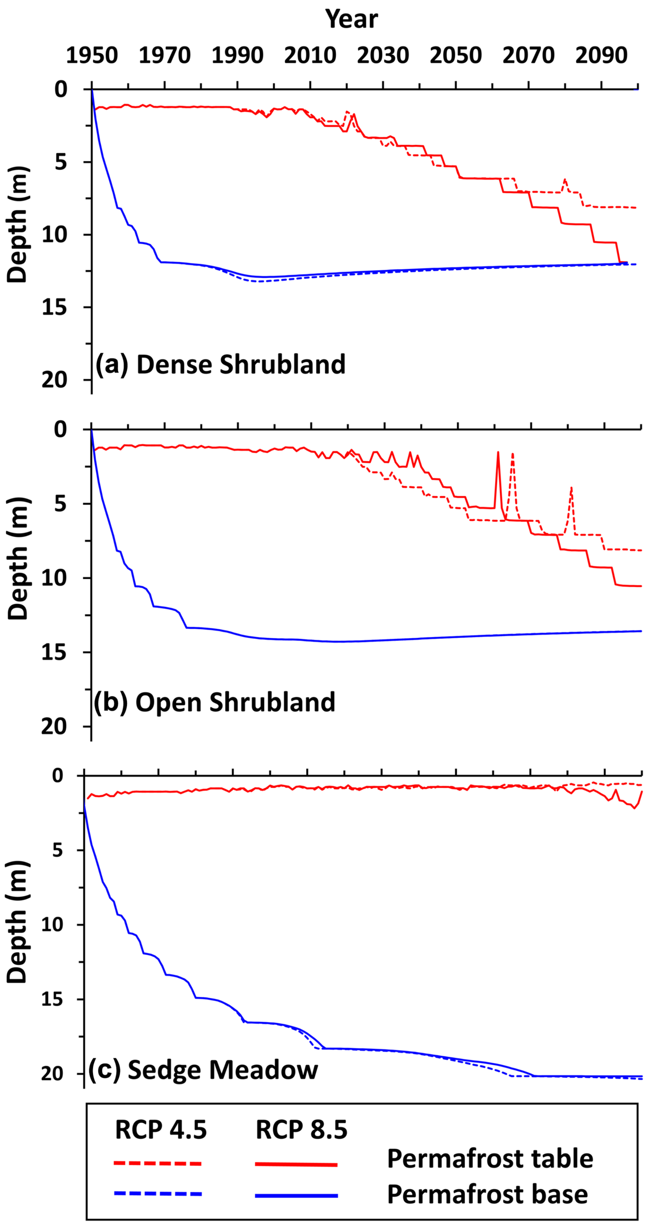

FIGURE 6 Modeled permafrost conditions (1950–2100) in three drainage scenarios: (a) dense shrubland, (b) open shrubland, and (c) sedge meadow. The red lines show the depth to the top of permafrost and the blue lines show the depth to the base of permafrost. Model runs using climate data from RCP 4.5 (2020–2100) are shown as dashed lines and model runs using RCP 8.5 data are shown as solid lines [Colour figure can be viewed at wileyonlinelibrary.com]

active layer was not completely refreezing (Figure 7a,b), and corresponding to the onset of the top-down thaw evident in Figure 6. Incomplete freezeback of the active layer in the open shrubland scenarios occurred after 2010 (RCP 4.5) and 2030 (RCP 8.5), resulting in top-down permafrost thaw and near-surface talik development (Figure 7c,d). In the sedge meadow scenario, a gradual increase in minimum winter temperature at 1 m depth was evident over time in both RCP 4.5 and 8.5, but annual winter cooling of near-surface permafrost persisted at all time steps except 2090 under RCP 8.5 (Figure 7e,f). In the sedge meadow scenario, the temperatures at 1-m depth increased with time and reached the highest levels in the 2090s under RCP 8.5 with the commencement of top-down thaw.

The impact of climate warming on permafrost aggradation and degradation was also evident in model runs with progressively later drainage dates (Figure 8). In the dense shrub scenario, increasingly later drainage dates decreased the rate of permafrost aggradation and sped the onset and rate of top-down thaw. The maximum thickness of the permafrost in the dense shrub scenario was reduced from 16.5 m (drainage in 1900) to 8.2 and 6.1 m (drainage in 2060) under the climate change projections of RCP 4.5 and 8.5, respectively. Under RCP 8.5, permafrost that aggraded following drainages after 1960 thawed completely by 2100 (Figure 8b). Under RCP 4.5, later drainage decreased the rate of permafrost aggradation, and sped top-down thaw, but the complete thaw of permafrost by 2100 did not occur for all the drainage dates (Figure 8a). In the sedge meadow scenario, drainage in later years slowed the rate of permafrost aggradation, decreased maximum depth of permafrost, and increased active layer thickness (Figure 8c–f). In this scenario the thickness of permafrost increased continuously before 2100 in all the cases except drainage in 2060 under RCP 8.5 (Figure 8e,f). Under RCP 4.5, active layer thickness in the sedge meadow scenario decreased gradually over time regardless of drainage date due to organic soil development. However, under RCP 8.5 later drainage dates resulted in top-down thaw, which occurred more rapidly with increasingly later drainage dates (Figure 8c,d). Model experiments that considered the development of excess ice for the dense and open shrubland scenarios decreased the rate of permafrost aggradation and reduced maximum permafrost depth by up to 1.3 m due to the latent heat requirement of ground ice formation and changes in the thermal properties of materials between 1.5 and 4 m depth (Figure S5). After 3 decades of warming, the presence of excess ice in these model experiments reduced top-down permafrost thaw by up to 2.3 m for a given year and delayed the degradation of permafrost by 10–15 years (Figure S5).

4 | DISCUSSION

Our analysis of field data shows that the effects of ecological succession on organic soil development and snow drifting will strongly influence the aggradation and resilience of permafrost in DLBs in the continuous permafrost zone. Our modeling study shows significant top-down thaw 50–70 years following lake drainage in scenarios with deep snow pack, shrub cover, and slow organic soil development. Near-surface (1 m) ground temperatures in five drained basins with dense shrub cover, shallow organic layers, and thick snow also remained nearly isothermal during the onset of winter over 5 years of observation (Figure 5). In the model scenario with sedge-dominated vegetation, low snow cover, and rapid organic soil development, permafrost aggradation continued and active layer thickness decreased throughout the entire model run under RCP 4.5, due to the accumulation of organic soil (Figure 6a,b). Under RCP 8.5 modest increases in

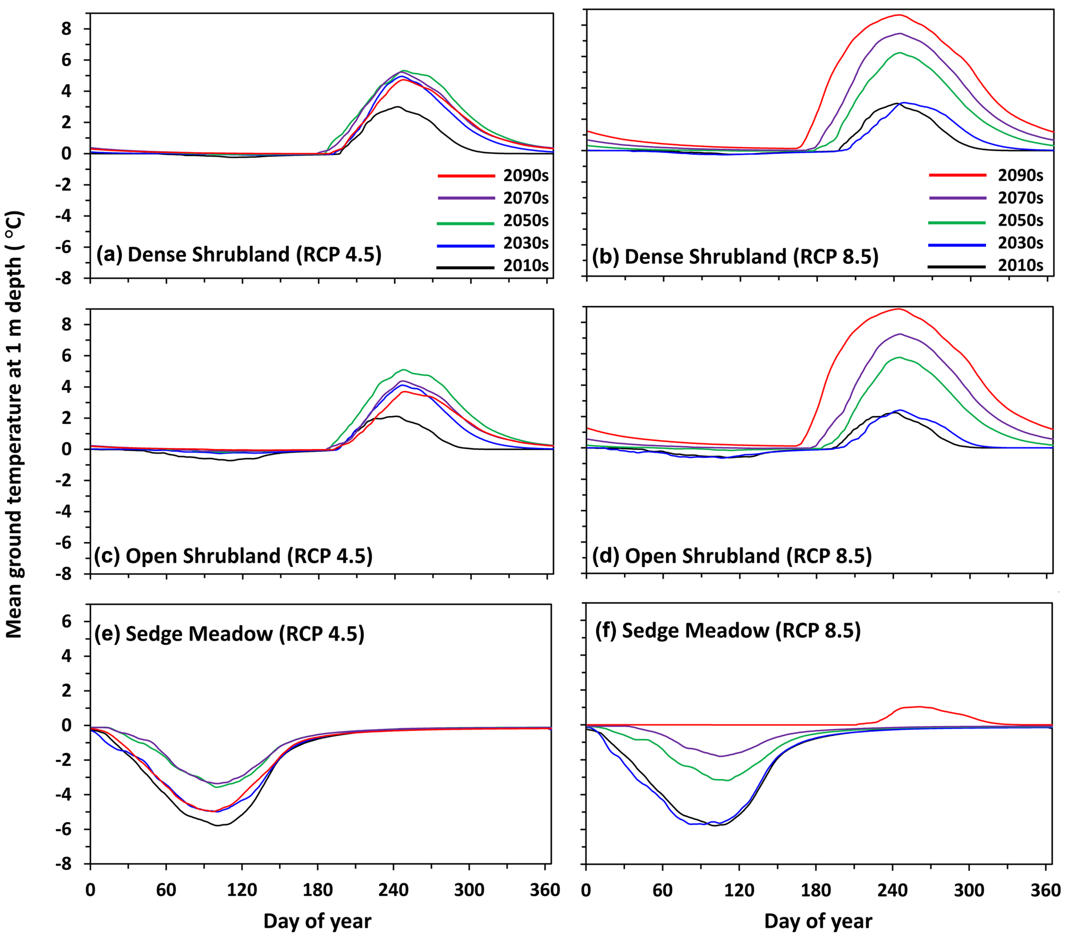

FIGURE 7 Modeled evolution of ground temperatures at 1-m depth under two climate projections in the three drainage scenarios initiated in 1950. Plots show daily mean temperatures by decade. Individual plots show the modeled conditions in the dense shrubland scenario (a–b), the open shrubland scenario (c–d), and sedge meadow scenario (e–f). The column on the left shows the models that use RCP 4.5 and columns on the right shows the models that use RCP 8.5 [Colour figure can be viewed at wileyonlinelibrary.com]

active layer thickness occurred in the last two decades of the sedge meadow scenario, 130 years following drainage (Figure 6c). The sedge-dominated site (thick organic soils and shallow snowpack) was also the only DLB we sampled where permafrost was encountered within the top 1 m (Figure 5).

Differences in temperature and active layer thickness in the DLBs in our field study were driven by higher ground heat flux in summer through soils with shallow organic horizons, and higher thermal conductivity, and reduced heat loss during winter caused by deep snow trapped by dense upright shrubs growing in DLBs. Observations at Illisarvik, an experimentally drained lake in the study region also show that snow captured by upright shrubs influences the ground thermal regime and increases permafrost temperature relative to the adjacent tundra.[29,47] At the oldest drained basin we sampled (HLDL6), which was dominated by sedges, a thick organic layer reduced heat flux to frozen soils, and shallow snow cover promoted rapid freezeback and cooling in the winter. Our model simulating succession to sedge meadow also showed that permafrost aggrading under these conditions was more resilient to climate warming than in shrub-dominated DLBs.

Higher ground temperatures and modeled permafrost degradation in shrub-dominated DLBs compared with the lower temperatures and shallower active layer thicknesses in sedge-dominated DLBs shows that the ecological modification of climate-driven permafrost can increase or decrease permafrost resilience within the continuous permafrost zone. Stabilized retrogressive thaw slumps with deep snow pack promoted by dense stands of upright shrubs also show persistent increases in thaw depth and ground temperature,[84,85] which increase the potential for disturbed slopes to destabilize in the future.[84] Observations of increased ground temperature and thaw depth on drilling mud sumps with shallow organic layers, dense upright shrub cover, and deep snow pack provide another example of how these processes and feedbacks can affect permafrost conditions.[86] Snow trapping by shrubs on slumps, drained lakes, and sumps probably has a larger impact on ground temperature than snow drifting caused by tundra shrub proliferation at undisturbed sites because well-developed organic layers at undisturbed sites reduce heat conduction in summer.[87,88] The topographic depression in slumps and drained lakes, combined with vegetation exceeding 2 m in height, also more effectively traps wind-transported snow compared

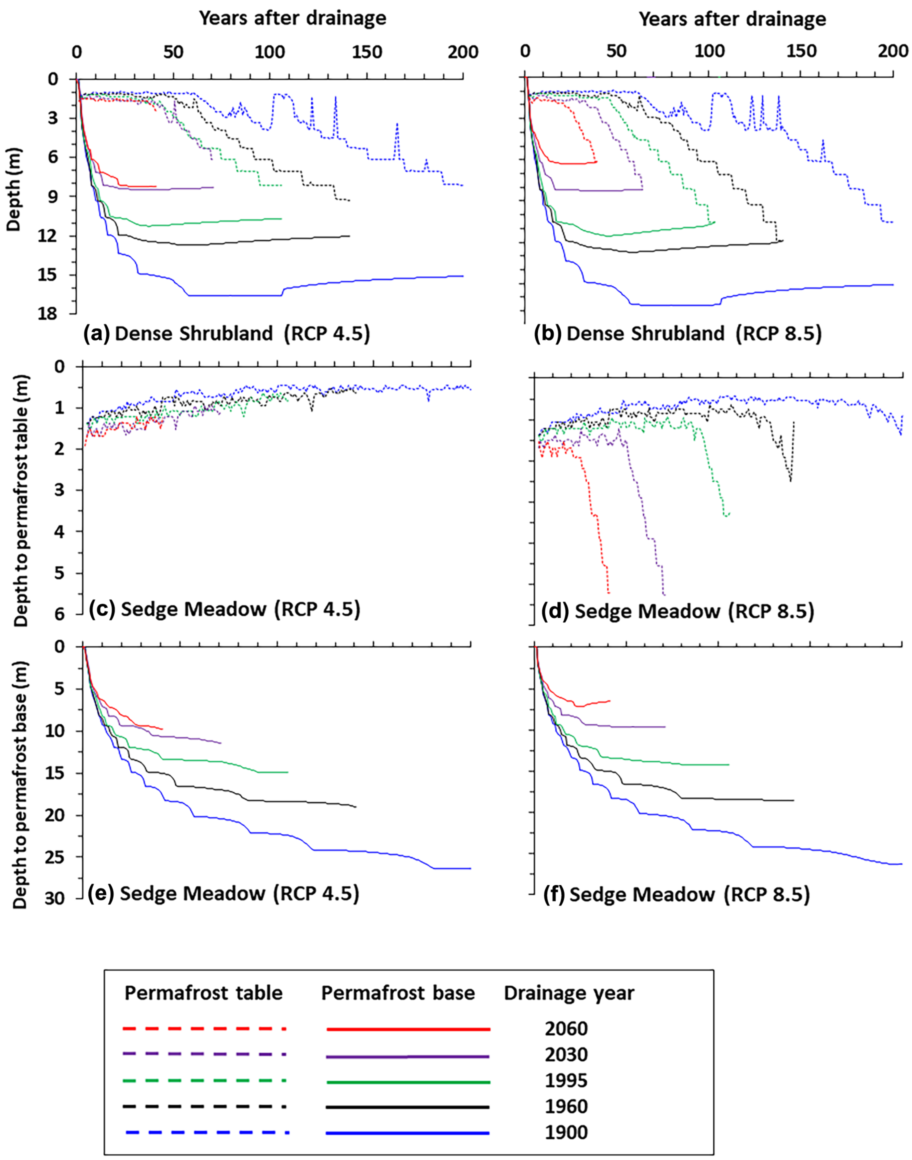

FIGURE 8 Modeled effects of drainage date on permafrost conditions following drainage in model runs using the climate time series shown in Figure S1. (a–b) Dense shrubland and (c–f) sedge meadow scenarios under RCP 4.5 and 8.5. The dashed lines show the depth to the top of permafrost, and the solid lines show the depth to the base of permafrost. Different drainage years are shown in different colors. Each curve shows the model results from the drainage year to 2100 [Colour figure can be viewed at wileyonlinelibrary.com]

with shorter shrub patches (<1 m) on the surrounding tundra.[88] Observations of permafrost thaw, terrain subsidence, and talik development following 30 years of deep snow caused by the construction of a snow fence in the Peel Plateau Region[89] suggest that areas with warm permafrost are likely to be most sensitive to ecological changes that promote snow accumulation or sustain shallow organics over mineral soils.

Latent heat effects associated with excess ice are likely to slow permafrost degradation in DLBs, but our modeling suggests that these effects will be small in young DLBs where excess ground ice is low. We did not sample the ground ice content at our study lakes, but low excess ice contents of 8% in the top 0.5 m of permafrost, 32 years following the drainage of Lake Illisarvik[38] suggests that near-surface ground ice in the top few meters of permafrost may also be low at our sites. Model experiments simulating excess ice contents ranging from 10 to 40% suggest that latent heat effects can reduce the rate of permafrost aggradation, and delay the progression of top-down thaw by several decades. High spatial variability in excess ice at Illisarvik also suggests that both aggradation of permafrost and surface heave, as well as top-down thaw will yield a mosaic of terrain, ecological, and thermal conditions within individual basins. Higher ground ice content and thicker organic layers,[26,51,74,89] in ancient DLBs suggest that these sites will be more resilient to climate warming than the recent DLBs we studied. However, excess ice levels exceeding 50% at sites on the Alaska North Slope[75] suggest that these environments will also be impacted by significant changes in topography and hydrology as near-surface permafrost thaws. Fieldwork to characterize the amount and types of ground ice in DLBs of varying ages in different geologic settings is needed to better characterize the nature and magnitude of changes that will result from climate-driven thaw.

Thermal modeling shows that thick organic soils and shallow snow pack can protect permafrost in DLBs from thaw, but it is unclear if ecological succession over the next century will produce these conditions in the Tuktoyaktuk Coastlands. Research in the Old Crow Flats shows that vegetation development following lake drainage depends on soil moisture, where wet soils promote the development of *C. aquatilis*-dominated communities with thick organic soil layers, and mesic soils facilitate the establishment of willow thickets with thinner organic layers.[27] In our study area, all five of the recent drainages (1950–1995) we sampled had mesic soils with thin organic layers and were dominated by upright shrubs that promoted snow accumulation (Table 1). Mapping across a larger number of DLBs in the Tuktoyaktuk Coastlands shows that vegetation at the basin scale is a heterogeneous mix of upright shrub thickets interspersed with areas of grasses and herbs, and small areas of hydrophilic vegetation (Table 1).[47] On centennial time scales these shrub- and grass-dominated plant communities will be replaced by later seral vegetation.[91–93] However, the dominance of relatively mesic conditions 40–60 years after drainage suggests that a transition to dwarf shrub tundra with relatively shallow organic layers is more likely than peatland development.[94,95] Several recent studies also indicate that increasing summer and winter temperature, and a longer thaw season are likely to promote mineralization and reduce rates of organic soil development in mesic tundra.[96–98]

The ubiquity of peatlands on ancient drained lakes in the western Arctic[48,49,57] suggests that long-term ecological succession can yield the low-stature vegetation and shallower snow that is conducive to permafrost aggradation, but it is unclear if this transition will occur under warming climate conditions and over what time scale. The presence of willow and birch macrofossils at the base of peat deposits in an ancient DLB in the Old Crow Flats[99] indicates that early successional shrub communities can be replaced by lower stature vegetation on centennial to millennial time scales. In the Tuktoyaktuk Coastlands, we sampled one sedge-dominated DLB with thick organic soil layers, which drained 100–200 years ago (Table 1) (see also[26,27,51]). However, the rarity of these communities on younger sites, raises the possibility that the warmer climate, which has predominated since the 1950s, favors the development and persistence of shrub communities following drainage.[100,101] This is also consistent with a large body of evidence demonstrating that increasing regional air temperatures have driven shrub expansion in undisturbed tundra across the western Arctic.[88,102–104] More extensive sampling across a range of drained lake ages, coupled with ongoing monitoring, could help to resolve these uncertainties. Ultimately, models projecting localized permafrost degradation indicate that long-term succession at these sites is likely to differ from changes that occurred in response to the cold climate that has predominated for the last 4,000 years.[105]

Taken together, our observations suggest that permafrost will continue to aggrade into emergent surfaces across the Tuktoyaktuk Coastlands despite even the most severe projected climate warming. However, as regional temperatures rise, developing permafrost will become increasingly susceptible to degradation caused by ecological feedbacks. This conclusion is consistent with the observation of short-term permafrost formation following the experimental drainage of a lake in the extensive discontinuous permafrost of the Pechora Lowland, Russia.[106] At this site the degradation of recently developed permafrost began a decade following drainage and was promoted by increased snowpack associated with willow proliferation. Regardless of the long-term outcome of succession, our modeling suggests that short-term rates of organic matter accumulation typical in DLBs will be too slow to protect developing permafrost from degradation in response to anticipated climate warming. Although ground temperatures in DLBs are likely to vary in response to fine-scale variation in soil moisture, vegetation structure, snow depth, and organic layer development, our data show that recently drained lakes will be amongst the first permafrost-free environments to develop in a warming Arctic. Our observation that most lakes in the Tuktoyaktuk Coastlands that drained in the last 60 years are dominated by upright shrubs indicates that this localized permafrost degradation will be regionally widespread (Figure 1). Although these areas make up a small proportion of the entire study area, the loss of warm and thin permafrost at these sites will have a significant impact on ecological processes including carbon flux,[107] subsidence caused by lateral thaw,[84] and population dynamics of beavers and moose in tundra ecosystems.[108,109]

Our field observations and modeling indicate that drained lakes will be among the first permafrost-free environments to develop under a warming climate in the continuous permafrost zone. Our observations highlight that other emergent landscapes (alluvial environments, floodplains, and coastal areas affected by isostatic uplift) will also be sensitive to permafrost degradation with climate warming, and are likely to emerge as the first permafrost-free environments to develop in the continuous permafrost zone. Long-term monitoring of DLBs and other emergent surfaces will provide critical insights into variability in the changes that can be expected as continuous permafrost adjusts to a warming climate. The thermal sensitivity of permafrost in DLBs also raises the possibility that other common periglacial landforms that typically characterize these environments (pingos, ice-wedge polygons) may not develop under a warming climate. It is also possible that the features that characterize discontinuous permafrost (lithalsas, peat plateaus palsas, etc.)[110,111] may become more common in the emergent environments of the Tuktoyaktuk Coastlands. Sustaining long-term monitoring, and conducting additional field and modeling studies will inform how these environments will evolve under warming climate conditions. Our results also imply that emergent environments further south within warm continuous permafrost, or extensive discontinuous permafrost terrain are likely to evolve as permafrost-free areas under warming climate conditions. This shift in the trajectory of permafrost aggradation/formation highlights the broader implications of this study, and underscores the importance of integrating field observation, sustained monitoring, and modeling to understand the dynamics of permafrost environments.

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CONFLICT OF INTEREST

The authors have no conflicts of interest to declare.

DATA AVAILABILITY STATEMENT

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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