

Northward expansion of high-stature vegetation reveals net surface-cooling feedbacks in the majority of Canadian Boreal-Tundra ecozones

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Climate warming has driven widespread shrub and tree expansion across the Canadian boreal-tundra. This is widely expected to amplify surface warming. However, whether this expectation is supported by observations remains unresolved. Here we show, using satellite observations of vegetation change alongside climate and surface-energy data from 1986 to 2023, that about 70% of central and northern Canadian ecozones exhibit net surface cooling tendencies, driven by interactions between vegetation change and the environment that reduce the energy available to warm the land surface (-0.003 to $-0.009\text{ W m}^{-2}\text{ yr}^{-1}$). Cooling is strongest in landscapes transitioning toward mixedwood forests, broadleaf vegetation and treed wetlands, reducing warming by approximately $0.015\text{--}0.028^\circ\text{C yr}^{-1}$ relative to shrub- and conifer-dominated areas. These findings challenge the view that vegetation expansion at high latitudes uniformly drives warming, showing that its effects depend on vegetation type, moisture and environmental context. They further suggest that the continued spread of mixed vegetation communities may moderate surface warming previously associated with conifer-dominated landscapes.

Climate change represents an existential threat to human societies and ecosystems, with the IPCC^{1,2} warning that exceeding 1.5°C of warming may trigger irreversible Earth system feedbacks including permafrost collapse, ice sheet loss and large scale ecosystem shifts. Recent records confirm this acceleration, as Copernicus Climate Change Service³ reported that both 2023 and 2024 were the warmest years on record, with global mean temperatures surpassing the 1.5°C threshold across sustained multi-month periods. This outcome highlights the urgency of understanding boreal-Arctic system responses, as the Arctic is

warming nearly four times faster than the global average^{1,2}. Permafrost thaw in the Canadian Boreal-Tundra (hereafter CBT; Supplementary Fig. 1 and Supplementary Table 1) is among the most significant of these risks. Storing over 1500 Gt of carbon, its destabilization could trigger irreversible feedbacks, amplifying greenhouse gas emissions and undermining global climate targets^{4–7}. Consequently, the evaluation of mitigation strategies in the CBT requires consideration not only of their carbon sequestration potential but also of their effects on surface energy balance and permafrost stability^{8,9}.

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In this context, tree expansion within the CBT, including both natural vegetation expansion ('borealisation') and afforestation, has become an important focus in discussions of vegetation-climate interactions, given the region's role as a critical frontier for understanding high-latitude climate-ecosystem feedbacks. However, a number of recent studies that are conceptualized through fire-albedo constraints argue that increasing tree dominance in tundra regions north of the boreal treeline, whether through natural vegetation transitions or assisted ones toward taller, darker evergreen vegetation, can reduce surface albedo and modify ground and turbulent heat fluxes. These changes, in turn, may elevate wildfire risk, accelerate permafrost thaw, and increase carbon emissions^{9–15}.

While these studies, framed within the fire-albedo paradigm and emphasizing positive warming feedback mechanisms, have strongly influenced current interpretations of vegetation-climate feedbacks in the CBT, their synthesis simultaneously raises a critical question regarding whether the hypothesized warming responses are consistently supported by observations across the broader CBT. Importantly, many of these expectations can be traced to earlier coarse-resolution modeling studies which, due to simplified representations of surface and vegetation structure, predicted that forest expansion into tundra would enhance localized warming sufficient to accelerate permafrost thaw^{8,16,17}.

Interestingly, emerging empirical evidence from ecological data and remote sensing (both spectral, e.g., normalized difference vegetation index (NDVI)), and direct vegetation identification from satellite imagery now challenge this view. Over the past four decades, large portions of the CBT have undergone persistent greening, primarily through unmanaged expansion of shrubs, treed wetlands, and mixedwood systems^{18–25}. These transitions toward taller and more structurally complex vegetation modify surface energy fluxes and therefore can provide an empirical basis for understanding how vegetation change influences surface energy dynamics across the CBT. Within this framing, if transitions toward higher stature vegetation inherently and uniformly enhanced surface warming, widespread ongoing greening would be expected to produce detectable regional-scale energy imbalances, including persistent increases in radiative forcing, elevated ground heat fluxes, and thaw-related surface responses. However, such patterns have not been systematically confirmed across the CBT, demonstrating that the biogeophysical consequences of northward high-stature vegetation transitions may be more complex than the prevailing fire-albedo paradigm suggests.

While canopy densification is expected to lower albedo, especially in simulation-based frameworks, several empirical studies highlight concurrent vegetation-type structural mechanisms that can increase albedo (e.g., the dominance of deciduous and relatively high-reflectance vegetation) and enhance surface energy dissipation through processes, such as increased latent heat flux, thermal bridging, snow interception, and sub-canopy shading in high-latitude boreal-tundra ecosystems^{26–31}. Moreover, recent modeling studies have demonstrated that vegetation structure and composition (e.g., shrub vs. conifer, deciduous vs. evergreen) critically influence both the direction and magnitude of surface energy and carbon-cycle feedback^{11,32–35}. This perspective is further supported by Ofosu et al.³² and Dsouza et al.³⁶, who, using the Carbon Budget Model of the Canadian Forest Sector (CBM-CFS3) together with remote sensing and climate physics modeling at the western boreal margin, showed that species mixing and ecohydrological conditions determine the balance between albedo-related warming effects and carbon uptake associated with shrub and tree expansion.

Despite these prevailing contrasting modeling and empirical evidence perspectives, a clear observation-constrained assessment of how vegetation structural transitions influence surface energy partitioning across the CBT remains lacking. In particular, it is not yet resolved whether large-scale shifts toward higher stature vegetation

states produce consistent surface warming responses, as commonly hypothesized, or whether alternative energy balance adjustments emerge under heterogeneous real-world landscape conditions. Therefore, in this study, we present a multi-decadal observation-constrained assessment of how vegetation structural transitions influence surface energy dynamics across the CBT. By integrating satellite-derived vegetation structure information with meteorological and energy flux records spanning 1986 to 2023, we evaluate whether transitions toward higher stature vegetation states are consistently associated with surface warming or cooling responses, thereby providing an observation-based perspective on the biogeophysical consequences of treed landscape development in the CBT.

Results

NDVI trends and associated vegetation structural transitions

To quantify vegetation structural change across the CBT, we combined Landsat-derived NDVI trends (1986–2023; Supplementary Tables 2–3) with vegetation type transitions from independent land-cover datasets (VLCE2 and GLFC; Methods 4.3–4.6,^{37–40}). NDVI greening and browning trends represented continuous vegetation densification and decline, while land-cover transitions mapped from land-cover datasets captured the redistribution of vegetation types and associated area changes^{41–43}. Agreement between these independent approaches provided a consistency check on the mapped transitions. Widespread northward linear greening and localized exponential greening were observed across CBT ecozones (Supplementary Fig. 2 and Supplementary Tables 4–6^{18,22,44}).

Greening trends were consistent across both disturbed (fire and harvest-affected) and background CBT conditions (i.e., full landscape conditions without explicit disturbance masking), with the highest greening trends observed in the Taiga Shield East and the Taiga and Tundra Cordillera ecozones (0.0013–0.0015 NDVI yr⁻¹) over more than 100,000 km² (Fig. 1). Localized exponential greening is most evident in the Taiga Cordillera and Southern Arctic, reaching up to 69,933 km² in the Southern Arctic, indicating rapid vegetation expansion in the latter⁴⁵. Adopting both linear and exponential NDVI trends in this study proved relevant for capturing complementary vegetation dynamics (Supplementary Fig. 3), with linear trends representing steady change and exponential trends capturing localized rapid expansion that would otherwise be underestimated if a linear-only approach were adopted.

In this study, we quantified the extent of exceedance of vegetation density class thresholds to provide additional context for observed vegetation land cover changes. Using Monte Carlo simulations (Method 4.4), we derived the idealized NDVI trend thresholds required to trigger vegetation density class transitions between 1986 and 2023. This yielded a critical minimum threshold of 0.003 NDVI yr⁻¹ for at least one transition to occur along the trajectory from barren to sparse, moderate, and dense vegetated cover, and ± 0.0015 NDVI yr⁻¹ for infilling transitions within adjacent classes (classes 4–5 and 1–2; Supplementary Figs. 4–5 and Supplementary Table 3) over the 38-year study period. Thus, high rates of these threshold exceedances were observed across the greening CBT ecozones, including the Taiga Shield East, Taiga Cordillera, Boreal Shield, Tundra Cordillera, and Southern Arctic. These exceedances are consistent with land cover transitions typical of greening ecological succession (precision $\geq 50\%$; Table 1), corroborating the observed patterns of widespread vegetation densification and northward expansion across the CBT. Altogether, these changes affected approximately 17–50% of the region, or 1.63 million km². Browning signals, however, were limited, occurring in less than 4% of the CBT area (Supplementary Fig. 6).

Consistent with the agreement between NDVI trends and land cover transition changes described above, both indicating a northward redistribution and intensification of high statured vegetation relative to 1986 baseline conditions (Figs. 2, 3 and Supplementary Fig. 7), the

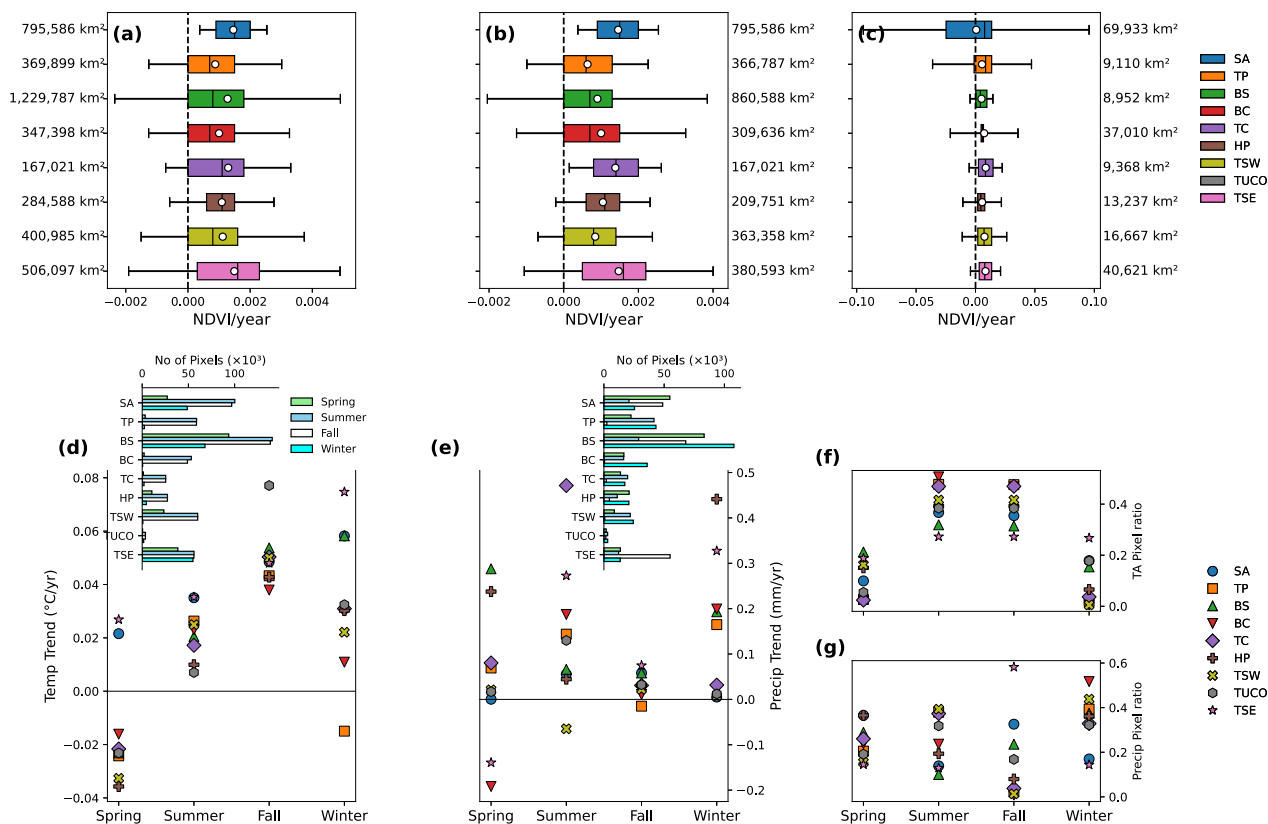


Fig. 1 | Ecozone-wise vegetation and climate trends from statistically significant pixels (1986–2023). **a** Linear NDVI trend summary for all statistically significant (both disturbed and undisturbed) pixels; **b** Linear NDVI trend summary for undisturbed pixels (disturbance excluded: fire and harvest); **c** exponential NDVI trend summary for undisturbed pixels. In (a–c), boxes show the interquartile range of pixel-wise slopes, the vertical line is the median, the white circle is the mean, whiskers denote mean $\pm$ 1 SD, and the dashed vertical line marks zero trend; numbers along the margins give the contributing area (km²). Colors denote

ecozones. **d** Seasonal near-surface air temperature (T2m) trends (°C yr⁻¹): The inset reports the number of statistically significant T2m pixels per season (in $\times 10^3$). **e** Seasonal precipitation trends (mm yr⁻¹) shown as points; inset reports the number of statistically significant precipitation pixels per season (in $\times 10^3$). **f, g** Seasonal fraction of statistically significant T2m and precipitation pixels per ecozone (fractions sum to 1 across seasons). SA Southern Arctic, TP Taiga Plain, BS Boreal Shield, BC Boreal Cordillera, TC Taiga Cordillera, HP Hudson Plains, TSW Taiga Shield West, TUCO Tundra Cordillera, TSE Taiga Shield East.

specific land cover transitions show that herbaceous vegetation declined by more than 90% in the Taiga Shield East and Hudson Plains, while bryoids declined by about 25% across ecozones. As an ecological successional process reflecting vegetation turnover rather than vegetation decline, shrub cover increased in the Taiga Shield by 62,000 km² (27%) and in the Southern Arctic by 14,000 km² (25%).

Latitudinally, the band between 50 and 66°N marked the strongest structural transitions, where shrubs gained about 30,000 km² as bryoid and herbaceous vegetation contracted, especially near 63–65°N. Mixedwood forests expanded most consistently, adding about 170,000 km² across the Boreal Shield and Taiga Cordillera and increasing by 6600 km² (332%) in the Boreal Cordillera. Coniferous forests contracted in the Boreal Shield by about 125,000 km² (15%) but expanded into the Taiga Plains and Taiga Shield East, offsetting losses at 42–52°N with gains above 60°N. Above 66°N gains were smaller but still positive for taller vegetation types, whereas below 56°N changes remained modest and patchy. In high-fire ecozones, such as the Boreal Shield, Taiga Shield West, and Boreal Cordillera, post-disturbance recovery increasingly favored coniferous and mixedwood stands. Generally, wetland conversion to treed wetlands increased across all ecozones. Taken together, these shifts indicate structural intensification of vegetation, with low-stature types, such as bryoids, herbs, and non-treed wetlands progressively being replaced by shrubs, treed wetlands, and taller forest types, most strongly across 55–66°N^{44,46}.

Climatic gradients in zones of vegetation change

Climatic gradients across the CBT indicate increasingly favorable growth conditions toward northern ecozones^{47,48} (Supplementary Figs. 8–9 and Supplementary Tables 7–8), consistent with the northward vegetation densification described above. Across 45–70°, temperatures increased at 0.026 °C yr⁻¹ (0.022–0.034; Fig. 4a). Greening in northern ecozones (Southern Arctic, Taiga Plains, and Tundra Cordillera) is primarily associated with temperature (r up to 0.3 in spring-summer), whereas southern regions show stronger precipitation sensitivity (Table 2 and Supplementary Tables 9–13). Latitudinal correlations show NDVI-temperature relationships shifting from negative in southern regions (spring $r = -0.23$ at 47–48°N) to positive across 51–60°N, peaking near 57–58°N in summer ($r = 0.33$ and Supplementary Table 11). NDVI-precipitation relationships show the opposite pattern, remaining positive in southern summers (up to $r = 0.42$) but negative across mid-latitudes ($r = -0.39$ to -0.27) before becoming weakly positive in the far north ($r \approx 0.09$ –0.12). These patterns indicate that greening, particularly north of the CBT, coincides with a relaxation of cold limitation, whereas vegetation responses in southern CBT ecozones reflect stronger moisture sensitivity^{24,25,40}.

Energy feedbacks observed across the ecozones

Natural transitions toward shrub- and tree-dominated landscapes are known to modify surface energy exchange across high-latitude environments^{49–53}. To assess how these vegetation structural

Table 1 | Precision of NDVI-landcover greening and browning intersection across CBT ecozones)

Land-cover Transition	NDVI Class	Zone	TP (pixels)	FP (pixels)	Precision (%)
Greening	Linear (Class 4-5)	Non-Arctic	169,194,418	148,439,433	53.27
		Southern Arctic	6,980,611	9,375,326	42.68
		Combined	176,175,029	157,814,759	52.75
	Exponential (Class 4-5)	Non-Arctic	71,485,495	12,775,371	84.84
		Southern Arctic	14,306,196	9,795,940	59.36
		Combined	85,791,691	22,571,311	79.17
Browning	Linear (Class 1-2)	Non-Arctic	13,578,594	6,741,550	66.82
		Southern Arctic	268,710	1,353,639	16.56
		Combined	13,847,304	8,095,189	63.11
	Exponential (Class 1-2)	Non-Arctic	11,017,918	4,777,410	69.75
		Southern Arctic	764,198	8,094,208	8.63
		Combined	11,782,116	12,871,618	47.79

TP true positives (pixels correctly classified as significant greening or browning, i.e. correct NDVI and landcover class intersection); FP false positives (pixels incorrectly classified). The Non-Arctic zone includes all ecozones together excluding the Southern Arctic ecozone. NDVI-slope thresholds for Classes 1–5 and landcover greening and browning classes are given in Supplementary Figs. 4–5 and Table 3.

transitions influenced ground-plant-atmosphere energy exchange, particularly through reductions in ground energy via outward heat export, we compared ecozone-specific growing-season (June–September; Supplementary Fig. 10), seasonal energy exchange trends (Table 3 and Figs. 5, 6) relative to their respective 1986 baselines (Supplementary Tables 14–15), and annual net trends from 1986 to 2023 (Supplementary Fig. 11).

Ecozones experiencing strong herbaceous declines ($> 90\%$; Hudson Plains and Taiga Shield East), mixedwood expansion ($\sim 170,000 \text{ km}^2$; Boreal Shield and Taiga Cordillera), and conifer contraction ($\sim 125,000 \text{ km}^2$; Boreal Shield) showed widespread spring cooling. This cooling was driven mainly by negative ground heat (G) trends (-0.008 to $-0.011 \text{ W m}^{-2} \text{ yr}^{-1}$) that offset positive sensible heat (H) (0.002 to 0.007) and latent heat (LE) (0.0025 to 0.0031) trends (Table 3; Fig. 6). In summer, less energy was transferred to the ground, which in concert with enhanced evapotranspiration (ET) moderated surface warming. Negative G trends (-0.010 to -0.011) and enhanced LE trends (-0.0016 to -0.0033) in the Hudson Plains and Taiga Shield East offset rising H trends (0.006 to $0.007 \text{ W m}^{-2} \text{ yr}^{-1}$), limiting net warming through both ground heat and evapotranspiration (ET) pathways. The Boreal Shield was the exception, with G and H trends turning positive (0.0034 and 0.0039), offsetting the negative LE trend (-0.0033) and tipping summer towards warming. By fall, across the low latitude ecozones, the combined positive G and H trends exceeded the negative LE trend, producing a net warming of 0.002 – $0.008 \text{ W m}^{-2} \text{ yr}^{-1}$, indicating the dominance of storage over ET cooling. However, this warming diminished in winter, with peaks reaching only the lower end of the range observed in fall. When integrated annually, Hudson Plains and Taiga Shield East showed net warming of 0.006 to $0.003 \text{ W m}^{-2} \text{ yr}^{-1}$, representing modest increases above their 1986–2023 baselines of 1.64 and 1.17 W m^{-2} . Warming fluxes covered just over half of each region. In contrast, the Boreal Shield remained effectively neutral relative to its 1.92 W m^{-2} mean baseline (Supplementary Table 16), despite a marginal spatial dominance of warming over cooling.

Furthermore, in the central latitude CBT ecozones, major vegetation transitions included shrub expansion ($11,600 \text{ km}^2$; Taiga Shield West), mixedwood expansion (6600 km^2 ; Boreal Cordillera), and conifer expansion in the Taiga Plains and Boreal Cordillera, with additional broadleaf increases. Despite forests remaining predominantly coniferous ($>60\%$), energy flux trends favored net cooling from spring through fall. This pattern was driven mainly by negative ground storage ($G \approx -0.004 \text{ W m}^{-2} \text{ yr}^{-1}$) and

evapotranspiration trends (LE up to $-0.004 \text{ W m}^{-2} \text{ yr}^{-1}$), which outweighed smaller positive sensible heat (H) trends (Table 3 and Fig. 6), indicating enhanced turbulent cooling and reduced subsurface energy retention. Although winter trends across the three ecozones generally shifted toward warming, the annual balance converged to net cooling; Boreal Cordillera and Taiga Shield West cooled by about $-0.003 \text{ W m}^{-2} \text{ yr}^{-1}$, and Taiga Plains by -0.009 . Cooling trends dominated more than half of each ecozone (57–69%), corresponding to reductions of 8.41, 9.41, and 24.5% from 1986 mean annual flux baselines of 1.32, 1.18, and 1.36 W m^{-2} (Supplementary Table 16). These results indicate that, unlike the low latitude ecozones, central latitude regions acted as persistent energy buffers, releasing ground heat toward the surface and transferring energy from the surface to the atmosphere across most of their landscapes.

At the high latitude CBT ecozones, where vegetation transitions comprised shrub expansion ($14,000 \text{ km}^2$; Southern Arctic), mixedwood expansion within the Taiga Cordillera, and increasing dominance of taller vegetation north of 66°N (e.g., Tundra Cordillera), seasonal LE trends indicated strengthening ET cooling (-0.00513 to $-0.0009 \text{ W m}^{-2} \text{ yr}^{-1}$), except for a slight winter ET reduction in the Southern Arctic ($0.00014 \text{ W m}^{-2} \text{ yr}^{-1}$). Mean H was generally positive (0.00139 to $0.00588 \text{ W m}^{-2} \text{ yr}^{-1}$), except during winter in the Southern Arctic (-0.00161) and Taiga Cordillera (-0.00010). Ground storage (G) was the dominant control. In the Southern Arctic, positive spring and fall storage (0.0009 – 0.0025) reinforced H and produced seasonal warming despite LE cooling, whereas in both Cordilleras strongly negative G (-0.009) intensified LE losses and offset H gains, sustaining net cooling (Table 3 and Fig. 6). Even in summer, when all ecozones experienced storage losses, shallow G decline in the Southern Arctic (-0.001) allowed positive H to dominate, whereas larger Cordilleran drawdowns maintained cooling. Annually, spatial aggregation (Fig. 6) showed convergence to net cooling of $-0.006 \text{ W m}^{-2} \text{ yr}^{-1}$, corresponding to reductions of 30, 22, and 20.4% from baselines of 0.74, 1.03, and 1.09 W m^{-2} . Cooling fluxes dominated roughly two-thirds of each ecozone. This contrasts with the central ecozones, where flux trends show modest cooling relative to 1986 baselines, and with the low-latitude ecozones, where two-thirds warmed.

Consistent with these energy partitioning patterns, ecozonal albedo responses (Supplementary Tables 17, 18) provide additional radiative context for surface-energy tendencies across ecozones. To distinguish vegetation-driven responses from those modulated by seasonal snow, and to isolate the radiative contribution of vegetation structure from snow-vegetation interactions, we calculated mean

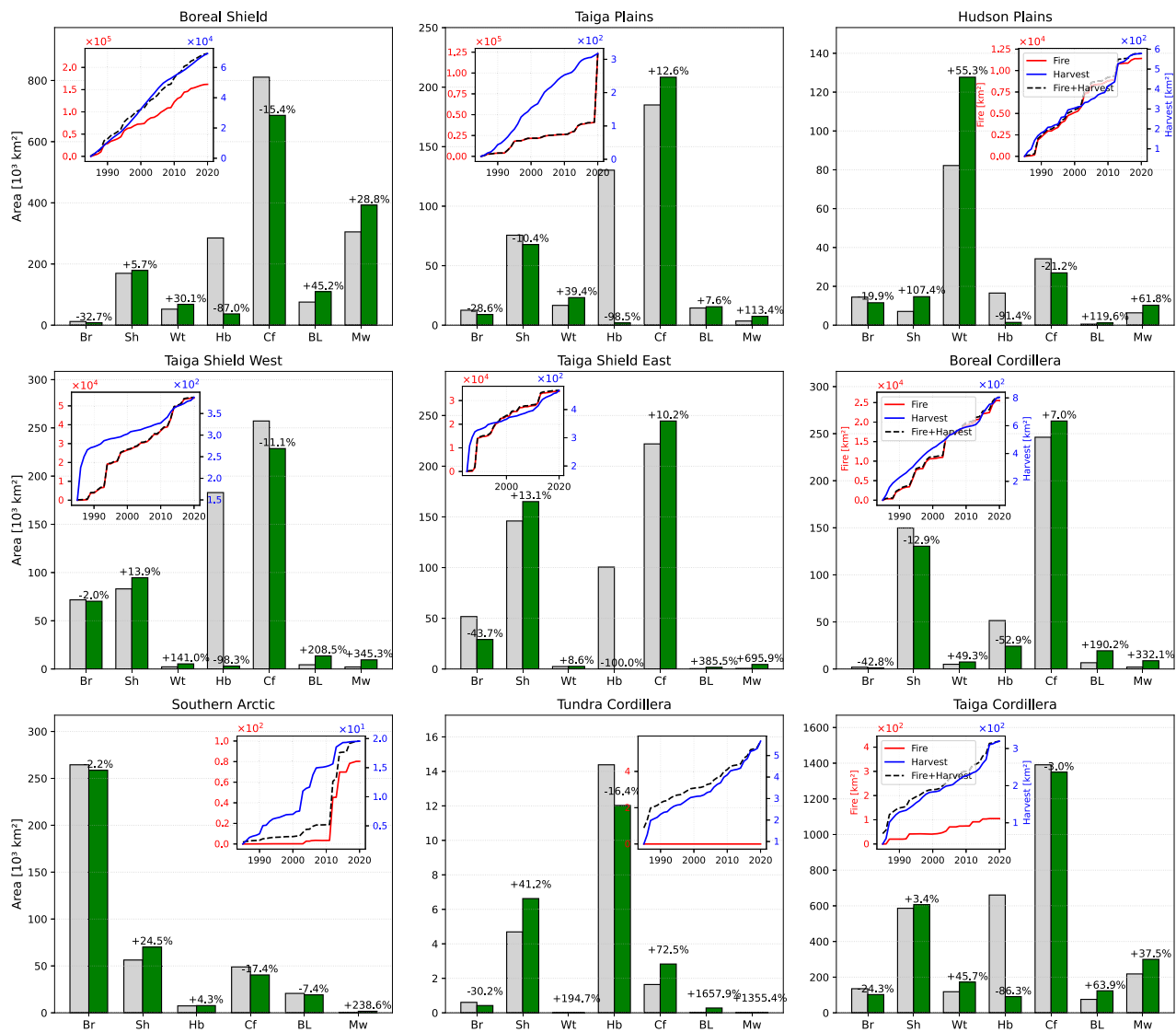


Fig. 2 | Land cover changes relative to the 1986 baseline and cumulative fire and harvest area from 1986 to 2022. Insets show cumulative disturbance time series, where the red line represents cumulative fire area and the blue line represents cumulative harvest area (km^2). The black dashed line represents the cumulative

total (fire + harvest) and is plotted on the same primary (left) y-axis as fire, while harvest is plotted on the secondary (right) y-axis. Br Bryoid, Sh Shrub, Wt Wetland treed, Hb Herbs, Cf Conifers, BL Broadleaf, and Mw Mixedwood.

percent decadal albedo changes under snow-free summer and annual conditions (Supplementary Table 17) and compared them with responses across the full seasonal and annual cycle under combined vegetation-snow conditions (Supplementary Table 18). Under snow-free conditions, summer and annual albedo responses are predominantly positive (generally up to 0.8% per decade), indicating that vegetation structural transitions alone do not consistently reduce surface reflectivity. When seasonal snow is included, winter responses remain largely positive, whereas spring responses become more variable, with both positive and negative changes across ecozones (e.g., negative spring albedo in the Taiga Shield East, Taiga Plains, and Boreal Cordillera). Overall, annual albedo responses remain predominantly positive across ecozones under both snow-free and combined vegetation-snow conditions except at the Boreal Shield and Boreal Cordillera. Together, these results show that albedo responses to vegetation structural transitions are not uniformly expressed as surface darkening, but are strongly modulated by snow-vegetation interactions, with the greatest divergence occurring during the spring transition period.

In summary, across the CBT ecozone, our attribution analyses show a consistent vegetation-energy coupling. Ecozone-level Kendall's τ results (Supplementary Table 19) reveal that vegetation gains are strongly associated with increasingly negative net energy flux trends in nearly all ecozones, indicating consistent surface cooling tendencies. Independent statistical comparisons between greening and stable vegetation areas reinforce this pattern: in all five sampling rounds, greening zones exhibit significantly more negative (cooling) net flux trends, supported simultaneously by Welch's t -tests and Mann-Whitney U tests (Supplementary Table 20). Taken together, these findings demonstrate that ongoing CBT vegetation gains are not only responding to climate warming but are actively mediating surface-energy pathways, promoting outward flux tendencies that can help protect permafrost.

Discussion

Vegetation transitions across the CBT during the past four decades indicate increasing vegetation expansion and reorganization across northern landscapes. Increasing temperature and precipitation have

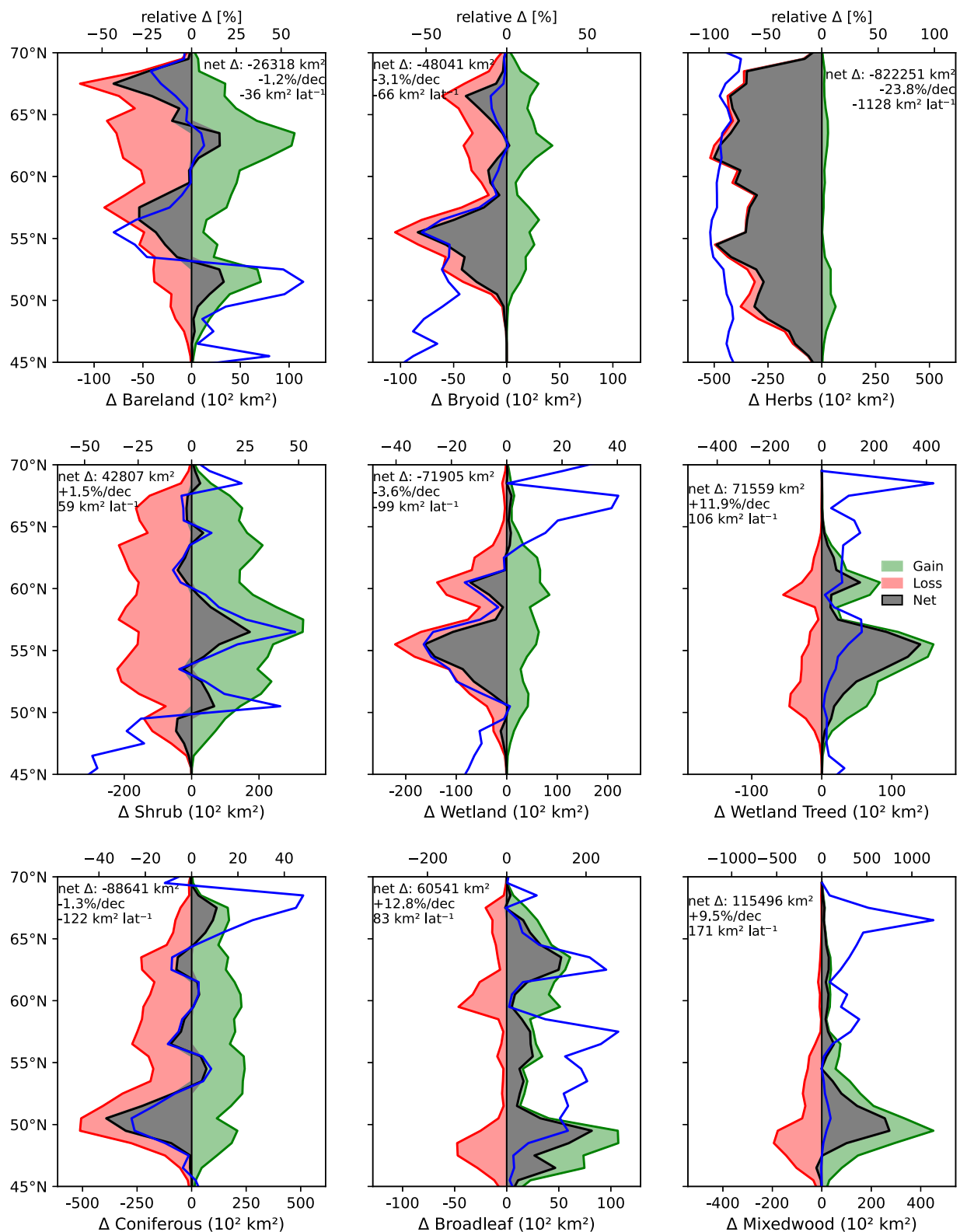


Fig. 3 | Vegetation latitudinal changes. The green, red, and gray shaded areas represent vegetation area gain, loss, and net change, respectively, between 1986 and 2022. These are plotted along the lower x-axis in square kilometers. The blue

line, plotted along the upper x-axis, shows the percentage of observed vegetation change relative to the 1986 baseline. The summary values shown above the plot represent totals aggregated across all latitude bands.

expanded the environmental envelope for vegetation growth, producing widespread greening signals that coincide with structural land cover transitions across ecozones (Fig. 4). Here, by linking long-term NDVI trajectories with independent land cover observations and

surface energy flux analyses, this study provides a pan-CBT assessment connecting vegetation structural change and spatial range shifts with surface energy redistribution across ecozones, extending earlier regional boreal-arctic vegetation expansion studies^{25,40,54–59}. These

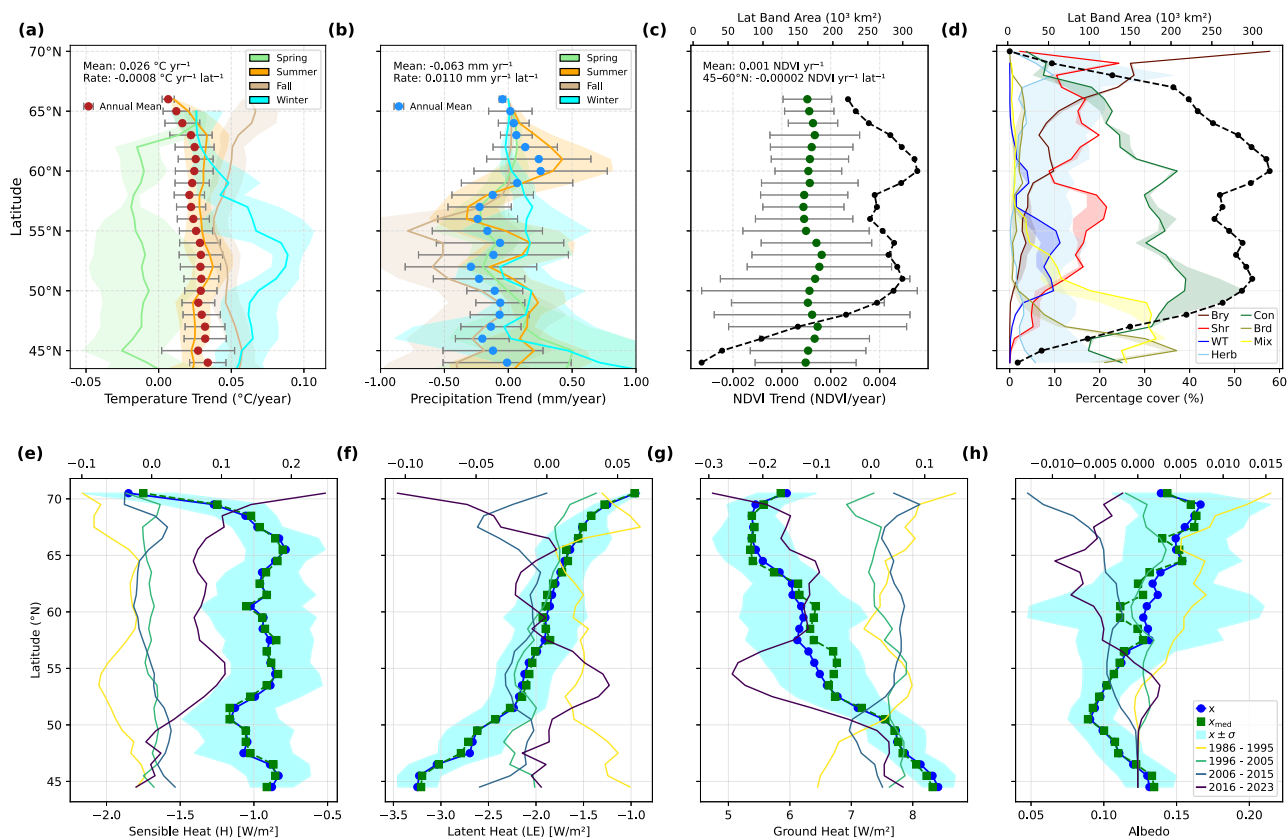


Fig. 4 | Observed trends in climate, vegetation, and energy fluxes across the boreal-tundra zone (1986–2023). **a–c** Zonal trends in temperature, precipitation, and NDVI across latitude (45–70° N); circles show annual means $\pm 1\sigma$; colored lines denote seasonal means. **d** Fractional vegetation cover transitions among major plant functional types (PFTs): Bryoids (Bry), Shrubs (Shr), Wetland Treed (WT), Herbs (Herb), Coniferous (Conn), Broadleaf (Brd), and Mixedwood (Mix). Shaded envelopes mark the 1986 baseline distribution; solid lines denote the 2022 state. The black dotted line on the secondary (top) axis gives latitudinal band area (10³ km²). **e–h** Latitudinal gradients of sensible heat flux, latent heat flux, ground

heat flux (which we assumed also as energy imbalance (G), i.e., the residual required to close the energy budget), and forecast albedo. In these panels, the dark-blue curve with X markers shows per-latitude mean values; green squares (X_{med}) indicate the median values by latitude; the cyan band shows mean $\pm 1\sigma$ by latitude. They are plotted on the down x-axis. However, decadal deviations are plotted on top x-axis as colored lines indicate: 1986–1995 (purple), 1996–2005 (yellow), 2006–2015 (blue), and 2016–2023 (teal). All trends are normalized per year; where indicated, slopes are expressed per degree latitude.

Table 2 | Spearman correlations between NDVI trends (1986–2023) and seasonal temperature/precipitation trends across Canadian ecozones

Ecozone	Temperature			Precipitation		
	Spring	Summer	Fall	Spring	Summer	Fall
Southern Arctic	0.28 ^{**} (252.0)	0.19 ^{**} (735.2)	−0.03 ^{**} (828.1)	0.10 ^{**} (454.2)	0.16 ^{**} (170.6)	−0.36 ^{**} (329.8)
Taiga Plains	0.17 ^{**} (32.0)	−0.22 ^{**} (496.8)	0.19 ^{**} (538.6)	0.09 ^{**} (192.1)	0.21 ^{**} (371.1)	0.19 ^{**} (60.3)
Boreal Shield	0.07 ^{**} (1174.2)	0.03 ^{**} (1483.1)	−0.09 ^{**} (1760.0)	−0.13 ^{**} (1069.0)	0.08 ^{**} (353.2)	−0.01 ^{**} (979.8)
Boreal Cordillera	0.01 (22.3)	−0.09 ^{**} (514.8)	0.12 ^{**} (471.7)	−0.11 ^{**} (159.0)	−0.04 ^{**} (154.6)	0.00 (167.9)
Taiga Cordillera	0.05 (11.1)	0.02 [*] (194.5)	−0.03 ^{**} (218.9)	0.07 ^{**} (119.2)	0.02 [*] (169.2)	−0.06 ^{**} (36.4)
Hudson Plains	0.09 ^{**} (122.2)	−0.02 [*] (174.1)	−0.07 ^{**} (330.9)	−0.06 ^{**} (257.3)	0.06 ^{**} (135.1)	−0.03 ^{**} (182.5)
Taiga Shield West	−0.05 ^{**} (248.0)	−0.08 ^{**} (497.7)	0.06 ^{**} (603.2)	0.00 (85.2)	0.14 ^{**} (217.2)	−0.02 [*] (131.8)
Tundra Cordillera	0.19 ^{**} (4.1)	−0.03 (9.1)	0.10 ^{**} (27.8)	0.01 (14.5)	−0.12 ^{**} (24.2)	0.01 (14.7)
Taiga Shield East	0.17 ^{**} (444.4)	0.22 ^{**} (622.1)	0.04 ^{**} (647.4)	−0.18 ^{**} (156.0)	−0.04 ^{**} (142.5)	−0.18 ^{**} (478.6)

Values are Spearman correlation coefficients (r , unitless, rounded to two decimal places) between NDVI trends (1986–2023) and seasonal temperature or precipitation trends. Values in parentheses are land surface area (10³ km²) of valid pixels used in each correlation, rounded to one decimal place. ^{*} $p < 0.01$, ^{**} $p < 0.05$. Pixel areas were estimated using a geodesic approach, converting raster resolution (0.041667°) to true surface area on a spherical Earth (radius = 6371 km), accounting for latitudinal variation in pixel size.

vegetation structural and spatial range transitions across the CBT remain spatially heterogeneous, varying differentially across latitude, including pronounced changes north of 50°N due to interacting thermal and hydrological constraints (Tables 2 and Supplementary Tables 9–11). In the Southern Arctic and ecozone portions between 60 and 64°N, warming continues to relax thermal limitations and support

shrub expansion^{25,60}, as evidenced by NDVI trends correlating positively with growing season and spring temperature ($r = 0.20$; spring $r = 0.28$; Supplementary Tables 9–10), whereas in the Taiga Plains warming-induced moisture stress appears to constrain growth (NDVI-temperature and precipitation correlations of -0.26 and 0.21 , respectively)⁶¹. The Taiga Shield, in contrast, represents a transitional

Table 3 | Ecozone summary of annual energy trends (warming, cooling, net) and vegetation fractional-cover change rates (1986–2023)

Ecozone	Eco Area(10 ⁵ km ²)	Warm area(%)	Cool area(%)	Σ Warming (W m ⁻² yr ⁻¹)	Σ Cooling (W m ⁻² yr ⁻¹)	Net Σ (W m ⁻² yr ⁻¹)	Ecozone NET(W m ⁻² yr ⁻¹)	Δ Vegetation(10 ⁻³ fcyr ⁻¹)		
								Δ Tree	Δ Shrub	Δ LowVeg
SA	9.49	27.6	69.8	47.04	-158.67	-111.63	-0.006 ± 0.012	-0.26	0.40	-3.96
TP	5.54	30.3	68.5	29.46	-108.93	-79.44	-0.009 ± 0.015	1.75	-0.39	-6.62
BS	18.93	50.6	47.8	168.18	-140.94	27.24	0.000 ± 0.015	0.19	0.14	-3.70
BC	5.58	41.2	56.7	32.10	-55.14	-23.04	-0.003 ± 0.012	1.95	-0.96	-1.40
TC	2.31	32.8	66.7	11.40	-44.49	-33.09	-0.006 ± 0.012	1.97	0.61	-3.09
HP	3.50	64.3	34.5	48.33	-22.32	26.01	0.006 ± 0.018	3.40	0.60	-1.43
TSW	6.19	41.0	59.0	46.53	-62.91	-16.38	-0.003 ± 0.012	-0.33	0.43	-6.73
TUCO	0.29	31.1	67.0	1.41	-4.74	-3.33	-0.006 ± 0.012	1.42	1.85	-2.43
TSE	7.02	54.4	44.5	86.85	-71.25	15.60	0.003 ± 0.018	1.34	0.91	-5.88

Notes: Warm/Cool area (%) are the fractions of each ecozone where the pixel-wise net-trend grid is > 0 (net warming) or < 0 (net cooling). The underlying seasonal flux trends (Spring, Summer, Fall, Winter) of LE, H, and G were first summed at each statistically significant pixel and then multiplied by 3 so that the resulting pixel-wise net-trend flux represents the trend in the full 12-month annual cumulative flux. Σ Warming and Σ Cooling (W m⁻² yr⁻¹) are the spatial totals of the annual warming and cooling contributions over valid pixels; Net Σ is their signed sum (<0 = net cooling; > 0 = net warming). Ecozone NET (W m⁻² yr⁻¹) is the mean annual net trend over valid pixels, scaled by the valid-area fraction so that it represents the entire ecozone. Eco Area is displayed in 10⁵ km² for compactness (use the true km² value for conversions). Vegetation share-change rates (Δ Tree, Δ Shrub, Δ LowVeg) are reported in $\times 10^{-3}$ fc yr⁻¹, where fc denotes the fraction (0–1) of the ecozone occupied by that class. Class definitions: Tree = Coniferous + Broadleaf + Mixedwood + Wetland Treed; LowVeg = Bryoid + Herbs. For each class, the share-change rate is Δ Tree/Shrub/LowVeg = $\frac{\text{P}_{\text{share}} - \text{P}_{\text{base}}}{36}$ (reported as 10⁻³ fc yr⁻¹). The individual vegetation share-change rate (fc yr⁻¹) is convertible to an area-change rate by $\Delta A = (\Delta \text{Tree, Shrub or Lowveg fc/yr}) \times A_{\text{ecozone}}$ (km² yr⁻¹). Ecozone SA Southern Arctic, TP Taiga Plain, BS Boreal Shield, BC Boreal Cordillera, TC Taiga Cordillera, HP Hudson Plains, TSW Taiga Shield West, TCO Tundra Cordillera, TSE Taiga Shield East.

hydroclimatic regime in which relatively small changes in temperature and precipitation can shift ecosystems between greening and decline^{24,25,62}.

Together, these patterns, including exponential greening that outpaces linear responses (e.g., Southern Arctic, Boreal Cordillera, and Taiga Shield) and net forest gains across ecozones toward mixedwood and wetland-treed systems (Fig. 2 and Supplementary Tables 4–6)^{18,23,24,40,63–66}, indicate that the CBT is currently in a disequilibrium vegetation state under climate warming that is tending, with delayed responses to climate forcing^{22,23,29,53,67}, toward intensified high-stature vegetation range shifts, redistribution, infilling, and a more mixed tree-dominated state. This interpretation contrasts with Rotbarth et al.²³, who reported limited tree expansion at the boreal tundra boundary and attributed it to demographic and environmental constraints. In the present analysis, these constraints manifest not as stasis but as turnover, where cases of tree decline coincide with increases in shrub (24.5%) and mixedwood (239%) systems, consistent with shrubification dynamics⁵⁰ that indicate active vegetation reorganization rather than absence of change.

As taller vegetation increasingly becomes a dominant feature across the CBT through ongoing natural vegetation transitions, these transitions provide a basis for examining how shifts in vegetation structure are associated with coupled changes in ground-plant-atmosphere energy flux partitioning under current climatic conditions. These observations may also provide broader insight into how future vegetation transitions and northward expansion pathways can influence surface energy redistribution across northern environments^{49,51,53}. It is known, and as also found in our study, that borealisation in the CBT toward shrub- and tree-dominated states, and the associated vegetation-mediated energy fluxes, are accompanied by distinct shifts in surface energy exchange that strongly influence ecological feedbacks in the permafrost-underlain CBT^{68–71}. A central question is whether, and which, vegetation structural changes can generate cooling through energy redistribution that offsets the warming effects commonly associated with reduced reflectivity and increased sensible heat flux in taller canopies.

In this study, mixedwood, broadleaf, and wetland-treed expansions are consistently associated with reduced warming, with annual temperature increases damped to approximately 0.015 °C–0.028 °C yr⁻¹ relative to stronger warming in shrub- and conifer-dominated

ecozones. Even during the summer thaw period, when higher temperatures would typically enhance ground heat storage, several ecozones exhibited reduced warming trends of approximately 0.01 °C–0.02 °C yr⁻¹ relative to annual means, and in the majority of ecozones increasing albedo rather than the expected decline^{72,73}. This apparent deviation from the expected albedo decline can be understood in the context of snow-vegetation interactions, but also reflects contrasting radiative responses when vegetation structural effects are evaluated independently from, and in combination with, seasonal snow conditions^{72,74,75}. Comparing snow-free conditions, which isolate vegetation structural effects, with full seasonal conditions shows that vegetation change alone does not consistently produce surface darkening, while the inclusion of snow introduces strong seasonal variability, particularly during winter and the spring transition period, when snowmelt likely exposes underlying vegetation and surface heterogeneity, leading to both positive and negative albedo responses across ecozones^{8,76}. Ecozones characterized by mixedwood expansion and Arctic and Shield shrub-dominated transitions generally retain positive snow-free albedo responses, whereas more negative or variable responses occur where conifer-related transitions remain important, consistent with the dependence of albedo on vegetation structure and snow-canopy interactions^{74,77}.

These patterns suggest that albedo responses to vegetation expansion in the CBT are influenced by both snow-vegetation interactions and ecozone-specific vegetation transition pathways, indicating that vegetation structure modifies surface energy exchange through mechanisms that extend beyond simple canopy darkening. For example, denser and moister canopies can scatter incoming radiation and enhance ET, redistributing energy through latent and turbulent flux pathways rather than ground heat storage^{76,78}. The observed combination of increasing albedo, enhanced latent and ground heat flux export (LE, H, G; Figs. 5, 6, Table 3, and Supplementary Table 16), and reduced summer warming, especially in mixedwood, broadleaf, and wetland-treed systems in this study, supports this mechanism. More broadly, our results show that, in the tree- and shrub-vegetated CBT, surface energy exchange is insufficient when interpreted solely through two-dimensional albedo contrasts, but is more realistically understood when accounting for three-dimensional vegetation structure and its influence on radiation scattering and flux partitioning⁷⁸.

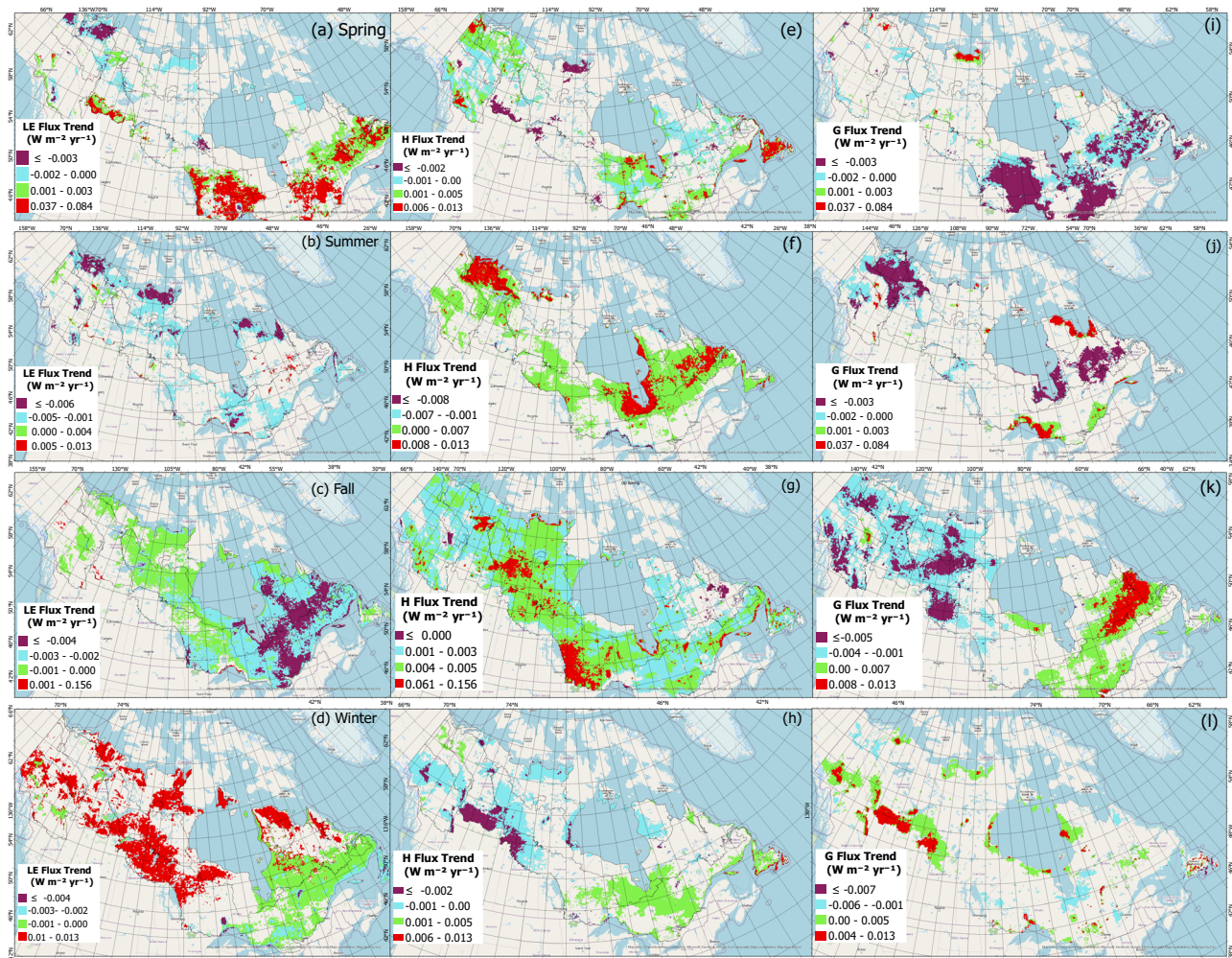


Fig. 5 | Seasonal maps of energy-flux trends across the Canadian Boreal-Tundra (1986–2023). Panels are arranged by season (rows: Spring, Summer, Fall, Winter); across a row the same season's statistically significant pixel-wise trends are shown for LE, H, and G: Spring (a,e,i); Summer (b,f,j); Fall (c,g,k); Winter (d,h,l). Column-

wise the same flux is compared across seasons: LE (a–d), H (e–h), G (i–l). All values are in $\text{W m}^{-2} \text{yr}^{-1}$, and black outlines show ecozone boundaries. Background base-map in each panel: OpenStreetMap Style (© OpenStreetMap contributors; Microsoft, Esri Community Maps contributors; rendered via Esri in ArcGIS Pro).

In line with the above energy dissipation mechanism postulation, the observed vegetation change-induced coupling between flux trends and annual surface temperature showed a supporting and expected pattern across most ecozones. Across ecozones, outward (cooling) flux tendencies are associated with reduced warming tendencies, whereas inward (warming) tendencies correspond to stronger warming (Table 4). Broadleaf and mixedwood expansion strengthens LE and G fluxes, enhancing evaporative and conductive energy export^{10,35,79}, while shrub- and conifer-dominated systems amplify sensible heat flux and reinforce warming^{75,80,81}. The Boreal Shield illustrates how balanced canopy composition can damp warming despite inward baseline fluxes ($1.94 \pm 0.07 \text{ W m}^{-2}$; Tables 3, 4, and Supplementary Table 16), in contrast to Taiga Shield East, where conifer persistence sustains stronger warming. Similarly, the Hudson Plains and Cordilleras maintain cooling tendencies through wetland-treed and open mixedwood mosaics with high reflectivity and strong latent coupling^{82–84}. These findings confirm that vegetation architecture, more than latitude or background forcing alone, drives spatial heterogeneity in warming across the CBT^{49,85,86}. Mixedwood, broadleaf, and wetland-treed systems promote outward energy redistribution and higher reflectivity, whereas shrubification and conifer densification intensify turbulent coupling and warming^{5,45,73}. In addition, modern thermal observations from the

U.S. Army Corps of Engineers' Cold Regions Research and Engineering Laboratory show that vegetated permafrost terrains exhibit reduced ground-heat flux and enhanced surface energy dissipation compared with unvegetated tundra⁸⁷. By extension, the vegetation-mediated energy redistribution underlying these observed cooling tendencies in our study may contribute to limiting permafrost thaw and its release of soil carbon⁶.

The cooling tendencies identified here reflect outward surface energy export and reduced ground heat uptake, contributing to reduced surface warming, yet are not directly equatable to net climate forcing benefit, particularly in wetland-dominated systems transitioning to treed landscapes, where disturbance-driven methane emissions not accounted for in this study may limit the magnitude of the climatic benefit of vegetation change for climate warming. Nevertheless, our results provide a structural and process-based basis for understanding how vegetation pathways influence surface energy balance and thermal responses in northern environments. In this context, the biogeophysical outcomes of both natural and managed shrub and tree expansion pathways in the CBT are likely to vary significantly depending on the resulting vegetation composition. Where mixedwood, broadleaf, and wetland-treed systems dominate, surface cooling tendencies have been observed, whereas conifer-dominated expansion is more consistently associated with warming responses.

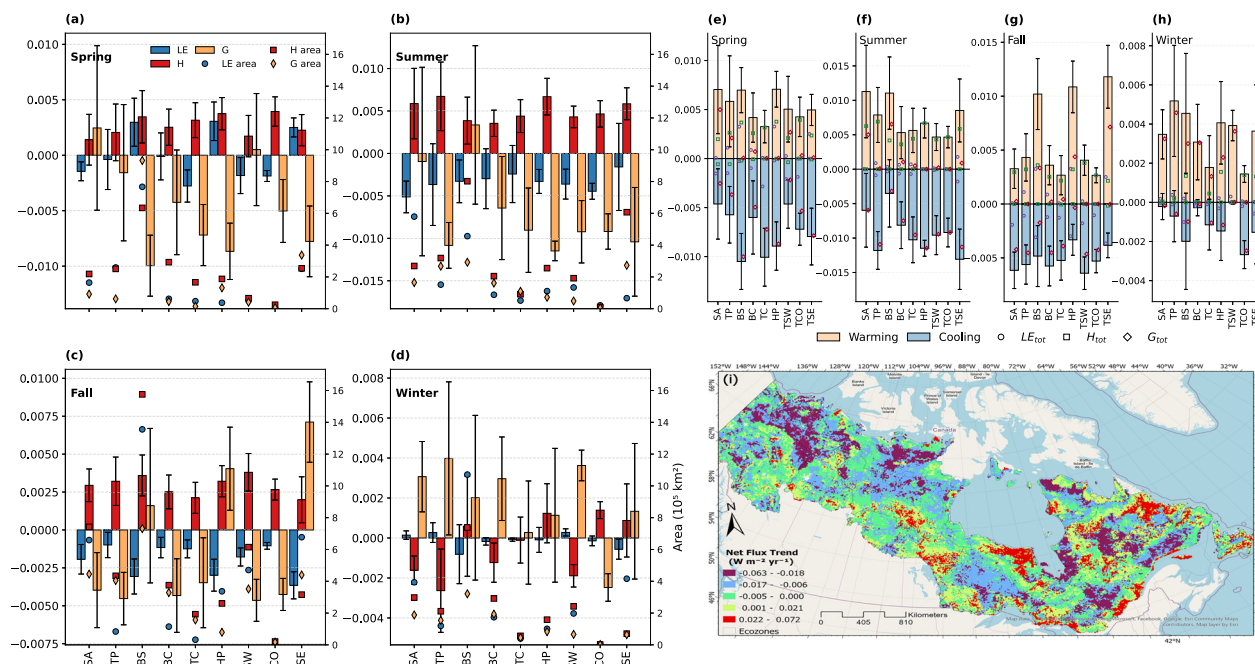


Fig. 6 | Mean Seasonal energy-flux trends summary by ecozone and spatial pattern of net trend flux (1986–2023). **a–d** Ecozone-mean, pixel-wise linear trends in latent heat (LE), sensible heat (H), and the ground term flux (G) for Spring, Summer, Fall, Winter initially presented spatial in Fig. 5. Colored bars show mean trends across pixels ($W m^{-2} yr^{-1}$); black whiskers are $\pm 1\sigma$. Point markers plotted on the right y-axis give the valid area used per flux trend component (in $10^5 km^2$; circles = LE, squares = H, diamonds = G). **e–h** Total magnitude of mean seasonal warming (positive trend flux) and cooling (negative trend flux) trends per ecozone: Each

semi-transparent bars show per ecozone for a given season the sum of absolute component magnitudes for warming (light yellow) and cooling (light blue) flux trends, with error bars from quadrature of component SDs; hollow markers show the contribution of each component (LE, H, G). **i** Pixel-wise map of the net trend flux ($W m^{-2} yr^{-1}$) with ecozone outlines; Pixel-wise map of the net energy-flux trend ($W m^{-2} yr^{-1}$) with ecozone outlines; full map in Supplementary Figs. 10, 11. Background basemap in (i): OpenStreetMap Style (© OpenStreetMap contributors, Microsoft, Esri Community Maps contributors; rendered via Esri in ArcGIS Pro).

Table 4 | Spearman correlation (ρ) between raster metrics and temperature by ecozone, with baseline and linear trend metrics for near-surface air temperature (T2m) and land-surface (skin) temperature

Ecozone	Cooling (ρ)		Warming (ρ)		T2m: mean & trend		Skin: mean & trend	
	T2m	Skin	T2m	Skin	$\bar{T}$ (°C)	Tr (°C/yr)	$\bar{T}_s$ (°C)	Tr_s (°C/yr)
BC	-0.20 [-0.22, -0.19]	-0.22 [-0.23, -0.21]	0.22 [0.21, 0.23]	0.24 [0.22, 0.24]	-3.99	0.03 (0.26)	-5.45	0.02 (0.22)
BS	-0.16 [-0.16, -0.15]	-0.16 [-0.17, -0.16]	0.24 [0.23, 0.25]	0.25 [0.25, 0.26]	0.93	0.03 (0.31)	0.79	0.03 (0.31)
HP	-0.62 [-0.63, -0.61]	-0.63 [-0.63, -0.62]	0.61 [0.59, 0.62]	0.61 [0.60, 0.62]	-2.11	0.05 (0.46)	-2.08	0.05 (0.45)
SA	0.07 [0.06, 0.08]	0.07 [0.07, 0.08]	0.07 [0.06, 0.08]	0.05 [0.04, 0.06]	-8.69	0.07 (0.66)	-9.79	0.06 (0.64)
TC	-0.38 [-0.40, -0.37]	-0.55 [-0.56, -0.54]	-0.02 [0.14, 0.17]	0.32 [0.30, 0.34]	-7.31	0.04 (0.38)	-8.78	0.03 (0.33)
TCO	-0.50 [-0.54, -0.46]	-0.54 [-0.59, -0.51]	0.53 [0.49, 0.57]	0.61 [0.56, 0.64]	-8.69	0.06 (0.56)	-10.01	0.06 (0.57)
TP	0.69 [0.68, 0.69]	0.69 [0.69, 0.70]	-0.53 [-0.54, -0.52]	-0.52 [-0.53, -0.51]	-4.51	0.04 (0.40)	-4.88	0.04 (0.38)
TSE	-0.25 [-0.26, -0.24]	-0.23 [-0.25, -0.22]	0.20 [0.19, 0.21]	0.17 [0.16, 0.18]	-3.71	0.06 (0.61)	-3.89	0.06 (0.61)
TSW	-0.03 [-0.04, -0.02]	-0.03 [-0.04, -0.02]	0.46 [0.45, 0.47]	0.45 [0.44, 0.46]	-6.13	0.04 (0.38)	-6.39	0.04 (0.36)

Note: Cooling (ρ) and Warming (ρ): Spearman rank correlations between the static cooling/warming raster intensities and multi-year mean temperatures, computed across co-registered pixels within each ecozone with values in square brackets showing their respective 95% Monte Carlo resampling confidence intervals. All correlation coefficients were statistically significant under the non-parametric permutation test. $\bar{T}$ and $\bar{T}_s$ is the mean of ecozone-mean annual temperatures (2m) and skin temperature over 1986–2023. Tr and Tr_s shows the linear slope of temperature (2m) and skin temperature (1986–2023, in $^{\circ}C yr^{-1}$ with the per-decade equivalent in parentheses ($^{\circ}C dec^{-1}$). Ecozone: SA Southern Arctic, TP Taiga Plain, BS Boreal Shield, BC Boreal Cordillera, TC Taiga Cordillera, HP Hudson Plains, TSW Taiga Shield West, TCO Tundra Cordillera, TSE Taiga Shield East.

It is important to note, however, that this study does not resolve the full climate implications of vegetation structural change across the CBT. In particular, the coupled energy and greenhouse gas consequences, especially methane dynamics, where carbon uptake and emissions may diverge, must be better understood before the observed surface cooling tendencies can be extended to net climate conclusions. Integrating flux tower, chamber, airborne, and remote sensing observations, together with field validation in key transition zones, will be essential to constrain these interactions and assess their broader applicability. Within this study framework, our results show

that vegetation structure governs surface energy partitioning and influences reduced surface warming, with implications for permafrost stabilization, indicating that the climatic relevance of afforestation pathways in the CBT may depend on ecozone conditions and vegetation type.

Methods

Description of study area

The CBT analysis extent covers approximately 589 M hectares (or ~ 6.54 billion pixels), across which about 6.54 billion pixels were analyzed. This

region covers nine Canadian ecozones, including the Taiga Shield (TS), Taiga Shield West (TSW), Boreal Shield (BS), Boreal Cordillera (BC), Taiga Cordillera (TC), Taiga Plains (TP), Tundra Cordillera (TUCO), Hudson Plains (HP), and Southern Arctic (SA)^{88–93}. Full disturbance regimes, management histories, and floristic conditions across these regions is in the supplementary materials (Section 2) and by Payette et al.⁹⁴, Payette and Delwaide⁸⁹, St-Onge and Grandin⁹¹, Hermosilla et al.⁸⁸, Rowe⁹⁵, and Timoney et al.⁹⁶.

Annual peak growing season NDVI extraction

To ensure computational tractability on Google Earth Engine, we partitioned the study area into fifty 890×725 km tiles (Supplementary Fig. 1a). For each tile, we retrieved all Tier 1, Level 2 Landsat 5, 7, 8, and 9 scenes from 1986 to 2023, restricting the analysis to the peak growing season (July 1–September 30; DOY 218–285). These temporal windows were to avoid pre-1986 data gaps and capture the yearly peak leaf-area growth across the CBT^{24,55,97}. Within each scene collection, we retained the blue, red, and near-infrared (NIR) bands, along with the opacity band for Landsat 5/7 and the aerosol band for Landsat 8/9. Pixel-wise NDVI was computed using Equation (1) for each observation and appended as an additional band. Each Landsat acquisition provides one 30 m pixel per grid cell, yielding n candidate NDVI values per year ($4 \leq n \leq 13$ and Supplementary Table 2). A combined water body and built-up surface mask, constructed using the GLFC⁹⁸, Canadian landcover⁹⁴, and GHBS datasets⁹⁹, was then applied to remove building and water pixels from the NDVI grids. The best available pixel (BAP) algorithm (Supplementary Section 4) then selects from the masked NDVI grids the best NDVI pixel from the n observations per grid cell for each year, producing one complete peak-season NDVI composite raster per year and a continuous annual time series (1986–2023) for the study area.

$$\text{NDVI}_{i,k} = \frac{\text{NIR}_{i,k} - \text{RED}_{i,k}}{\text{NIR}_{i,k} + \text{RED}_{i,k}}, \quad k=1, \dots, n_t, \quad (1)$$

where i is the 30 m grid cell; $t \in [1986, 2023]$ denotes the year; n_t is the number of peak-season Landsat acquisitions in year t (Supplementary Table 2); and $k \in \{1, \dots, n_t\}$ indexes those acquisitions.

Peak growing season NDVI ensemble trend fitting

From the yearly peak NDVI time series stack derived in subsection 4.2, we fitted sequentially three models: nonparametric linear (Mann-Kendall with Theil-Sen slope), exponential (Equation (4)–(3)), and logistic (7) to derive the pixelwise NDVI trends across the study area. For linear trend the Mann-Kendall test ($|Z| > 0.84$) was used to identify pixels with significant monotonic trends, after which the Theil-Sen estimator was used to derive the corresponding slopes as:

$$\beta_{\text{TS},i} = \text{median} \left\{ \frac{\text{NDVI}_i(t_m) - \text{NDVI}_i(t_n)}{t_m - t_n} \right\}, \quad t_m > t_n, \quad t_m, t_n \in [1986, 2023]. \quad (2)$$

yielding a linear prediction of total change:

$$\Delta_{\text{lin},i} = \beta_{\text{TS},i} (2023 - 1986) \quad (3)$$

To be able to fit the exponential model in the Google Earth Engine platform, which is not yet equipped with direct non-linear model fitting functionalities, we performed the log-linear regression that underpins our exponential fit (Eq. (4)) with $R^2 \geq 0.6$ and $\text{RMSE} \leq 0.002$ acceptance level and for which baseline NDVI is non-negative. At this stage, any pixel that initially failed the linear test but met these exponential criteria was carried forward. Next, we back transform the log-space slope r into a per-year multiplicative NDVI change rate r' (Eq. (5))

relative to baseline 1986 NDVI value NDVI_0 (Eq. (6))

$$\ln(\text{NDVI}_i(t)) = \ln(\text{NDVI}_0) + r(t - t_0) \quad (4)$$

$$r' = e^r - 1 \quad (5)$$

$$\text{NDVI}_i(t) = \text{NDVI}_0 (1 + r')^{t-t_0} \quad (6)$$

where $t_0 = 1986$ and NDVI_0 is the 1986 NDVI value and r' being the predicted rate of change of NDVI from 1986 to 2023.

To address the known underestimation of long-term change from linear fits (parametric and non-parametric), we computed each pixel's observed ΔNDVI as the difference between early (1986–1989) and late (2020–2023) means. We then compared the total change predicted by the Theil-Sen and the r' parameterized exponential model (Eq. (6)) against this observed ΔNDVI , retaining the model that best matched without systematic under- or overestimation. Pixels not adequately described by either model were fitted with a logistic curve (Eq. (7)), enforcing $R^2 \geq 0.6$ and $\text{RMSE} \leq 0.002$. However, the logistic model was dropped since fewer than 100 pixels were captured. The end product of the NDVI trend ensemble fitting process consists of model class grids (linear, exponential, and non-significant) and corresponding slope grids, where each pixel is assigned a statistically significant trend magnitude (Theil-Sen slope or r') based on its best-fitting model, with logistic fits excluded due to negligible spatial representation. At the end of the NDVI trend fitting, we used the NFDB, NBAC fire, and Canada harvest datasets as composite masks to delineate disturbed (fire and harvest) and non-disturbed conditions, and computed ecozone-level mean linear NDVI slopes for statistically significant pixels, separately for all pixels and for disturbed pixels only.

At the end of the NDVI trend fitting, we used the NFDB, NBAC fire, and Canada harvest datasets as composite masks to delineate disturbed (fire and harvest) and non-disturbed conditions. These masks were then used to compute ecozone-level mean linear NDVI slopes for statistically significant pixels, separately for all pixels (disturbed and non-disturbed) and for disturbed pixels only, enabling comparison of trend magnitudes across disturbance conditions.

$$\text{NDVI}_i(t) = \frac{L_i}{1 + e^{-k_i(t_i - t_{\max})}} \quad (7)$$

where L_i is the pixel maximum-minimum observed ΔNDVI ; k_i is the logistic ΔNDVI rate constant fit via logit-linear regression; t_i is year (1986–2023); $t_{\max}$ is the year of maximum NDVI change.

NDVI linear and exponential trend slope reclassification

To utilize the observed NDVI trend slopes as an independent validation proxy to check consistency of landcover transition (subsection 4.6), we reclassified the NDVI slope grids to match landcover change or stable areas derived from landcover datasets. Grid-based binning is a standard approach for identifying threshold values across spatial grids. To detect areas where NDVI trends indicate forward (greening), backward (browning), or stable conditions and enable overlay with the 2022 landcover transition raster, we needed to define slope thresholds for NDVI reclassification. In the absence of clearly defined thresholds in the literature, we specified NDVI range thresholds based on theoretical NDVI behavior for four vegetation density classes in our region: barren (−0.01–0.10), sparse (0.10–0.30), moderate (0.30–0.60), and dense (0.60–0.80)^{100–102}. Based on NDVI behavior, we assumed that the barren threshold corresponds to bare ground and wetlands, sparse to lichens, herbs, and shrublands, moderate to patchy evergreen and deciduous vegetation, and dense to fully stocked single or mixed evergreen stands marking the end of succession. We further assumed that in a vegetative successional (greening) or non-recovery

(browning) system, the NDVI end-state must lie within either barren or dense thresholds according to its trajectory.

Using these ranges for each class below the dense category, we ran 5000 Monte Carlo simulations to derive a single minimum NDVI slope threshold required for class transition over 38 years (1986–2022) or stability across vegetation density classes consistent with observed landcover transitions. Results showed that an absolute median NDVI slope of at least $0.003 \text{ NDVI yr}^{-1}$ (Supplementary Fig. 2) is required for a full transition between classes. We therefore adopted half of this value as the critical threshold: $0.0015 \text{ NDVI yr}^{-1}$ for linear trends and $0.00132 \text{ NDVI yr}^{-1}$ for exponential trends, representing an intra-class “infilling” threshold: large enough to indicate intra-class dynamics but insufficient for a full class shift. With these thresholds, we formulated five NDVI trend classes corresponding to significant browning and moderate browning (Classes 1–2), statistically stable (Class 3), and moderate and significant greening (Classes 4–5; Supplementary Table 3), which we used to reclassify the NDVI trend slope rasters. The two (i.e., linear and exponential) resulting five-class rasters were then used for cross-consistency checking and validation of landcover transition classes as described in subsection 4.6.

Landcover transition analysis

For landcover transition analysis, we used Canadian annual landcover datasets (1984–2022) from the Virtual Land Cover Engine (VLCE2)⁹⁴ and the global landcover dataset (1985–2022; GLC_FCS30D, hereafter GLFC V3)⁹⁸. VLCE2 classes include water, snow/ice, rock/rubble, exposed/barren, bryoids, shrubs, wetland (treed and non-treed), herbs, coniferous, broadleaf, and mixedwood. For our analysis, we extracted the Southern Arctic extent of GLFC V3 and harmonized its sub-categories (waterbody, snow/ice, shrubland, bareland, herbs, deciduous forest, evergreen forest, and mixedleaf forest) into corresponding VLCE2 categories to form a unified CBT annual landcover dataset after masking unclassified, water, and built-surface pixels. To establish a stable baseline and reduce spurious year-to-year noise, we applied a pixelwise majority filter to the VLCE2 portion of the 1984–1986 unified landcover dataset and defined the dominant class per pixel as the 1986 baseline. For the GLFC portion, we used the 1985 epoch as the baseline, as it was derived from a multi-year composite that minimized classification errors. Using the harmonized dataset with stabilized baselines, we implemented a pixel-wise FROM–TO (i.e., from baseline class in 1986 to final class in 2022) landcover transition matrix (Supplementary Fig. 5), producing a raster of coded 2022 pixel-wise transition values (full implementation in Supplementary Section 3). From this transition raster, we tabulated geodesic landcover class area changes (1986–2022) and change types (greening, browning, stable, and lateral vegetation shift; Supplementary Fig. 5). For the form changes type, we defined a “meaningful ecological change” as a pixel whose 2022 class matches its 2020 or 2021 state, signifying transition completion, and that has followed a consistent meaningful ecological successional trajectory since its first change within the 37-year series.

NDVI trends and land cover transitions consistency

Using the NDVI trend binning framework defined in subsection 4.4 (and supplementary Table 3), we assessed consistency between NDVI trend classes (Classes 1–2 for browning, 3 for stable, and 4–5 for greening) and land cover transition raster classes derived from subsection 4.5. To achieve this, we intersected randomly drawn pixel samples from the two (linear and exponential) reclassified NDVI trend rasters (Classes 1–2 and 4–5) with the coded 2022 pixel-wise landcover transition raster (subsection 4.5). Then we computed the precision level as:

$$\text{Precision}_c = \frac{TP_c}{TP_c + FP_c} \times 100 \quad (8)$$

where c denotes the transition class (greening or browning); TP_c is the number of pixels correctly identified as class c in both the NDVI slope classification and the observed land cover coded transitions for greening or browning; and FP_c is the number of pixels predicted as class c in the NDVI slope classification that did not correspond to observed land cover coded transitions value for greening or browning. With this approach, we evaluated how a simple NDVI-based benchmarking approach with high-resolution ground truth data helps not only to identify regions of vegetation/landcover changes but also to predict the timing of landcover transitions to newer classes within the time series, in contexts where the historical vegetation or landcover type is known.

Vegetation change and climate trend relationship

To associate the observed vegetation type changes with climatic (temperature and precipitation) conditions, we performed latitudinal gradient and ecozone-level association seasonal and annual analyses. First, we retrieved spring (March–May), summer (June–August), fall (September–November), winter (December–February), and growing season (June–September) monthly minimum and maximum temperatures ($^{\circ}\text{C}$) and precipitation (mm/month) between 1986 and 2023 from the CRU TS 4.09 dataset¹⁰³, which had been spatially downscaled from its native 0.5° resolution to approximately 2.5 arc-minutes (km^2) using WorldClim v2¹⁰⁴. Monthly average temperature was calculated as the mean of minimum and maximum values, after which statistically significant seasonal linear trends ($P \leq 0.05$) were derived spatially, pixel-wise, for both climatic variables. We then intersected, pixel-wise, ecologically and latitudinally, the resulting temperature and precipitation statistically significant seasonal trend rasters one by one with the NDVI-derived statistically significant trend raster (Section 4.3) to estimate their correlations (Spearman, $P \leq 0.05$), which we presented in Tables 2, Fig. 4a, b, and Supplementary Table 4.

Vegetation change and energy flux trend relationship

From assessing whether observed NDVI and landcover trends are driven by climatic change, we next examined how landcover vegetation changes influence surface energy fluxes. For this, we used ERA5-Land monthly data (1986–2023) from the GEE platform to extract the seasonal (Spring to Winter as mentioned above) latent heat flux (LE), sensible heat flux (H), net solar radiation (R_{solar}), net thermal radiation (R_{thermal}), and forecast albedo datasets, across the ecozone and latitude range of $45\text{--}70^{\circ}\text{N}$. All fluxes, expressed in joules per square meter (J m^{-2}), were converted to watts per square meter (W m^{-2}) by dividing by the exact number of seconds in each respective month at the native 11.132 km spatial resolution. The ground heat flux (G) necessary to close the energy budget pixel-wise was computed by Eq. (9):

$$G (\text{W m}^{-2}) = (R_{\text{solar}} + R_{\text{thermal}}) - (LE + H) \quad (9)$$

where G is the ground heat flux, R_{solar} is the net solar radiation, R_{thermal} is the net thermal radiation, and LE and H are the latent and sensible heat turbulent fluxes, respectively. We then fitted a Mann-Kendall trend test at an absolute Z-score threshold of 1.64 to identify, individually per season, the three component fluxes (LE , H , G) pixels with a significant monotonic trend, and then used the Theil-Sen estimator to derive their corresponding slope values across the CBT ecozones.

Additionally, apart from the trend pixel-flux estimation, we also, for each 1° latitude band, derived the mean, median, standard deviation, and interquartile range (25 and 75th percentiles) of LE , H , G , and albedo of only the growing season (June to September). We also computed the decadal mean deviation from the overall long-term mean for each variable, defined by Equation (10) to help add context on latitudinal gradient vegetation growth influences and relationships

with observed NDVI and landcover changes:

$$\Delta_{\text{decade}} = \bar{X}_{\text{decade}} - \bar{X}_{\text{overall}} \quad (10)$$

where $\bar{X}_{\text{decade}}$ is the average value within a specific decade and $\bar{X}_{\text{overall}}$ is the mean across the full growing season period of 1986–2023.

Ecozonal energy flux trend partitioning. We partitioned the seasonally significant LE , H , and G flux trends into cooling and warming by adopting a sign-based assumption. Negative trend values were classified as cooling, indicating a reduction in surface energy retention, either due to an actual loss of energy (flux decreasing below zero) or a weakening of an initially positive flux. Positive trend values were classified as warming, reflecting increasing energy retention at the surface, either through enhanced flux magnitude or a reduction in energy loss relative to the 1986 baseline. The partitioning is formulated mathematically as:

$$F_{\text{trend}} = \begin{cases} \text{Cooling,} & f_{\text{trend}} < 0 \\ \text{Neutral,} & f_{\text{trend}} = 0 \quad (f_{\text{trend}} \in \{LE_{\text{trend}}, H_{\text{trend}}, G_{\text{trend}}\}), \\ \text{Warming,} & f_{\text{trend}} > 0 \end{cases} \quad (11)$$

We then aggregated the seasonally derived cooling and warming grids into composite rasters for each category, and computed the net energy trend as the signed sum of the two:

$$F_{\text{cooling}} = LE_{\text{cool}} + H_{\text{cool}} + G_{\text{cool}}, \quad (12)$$

$$F_{\text{warming}} = LE_{\text{warm}} + H_{\text{warm}} + G_{\text{warm}}, \quad (13)$$

$$F_{\text{net}} = F_{\text{cooling}} + F_{\text{warming}} = LE_{\text{trend}} + H_{\text{trend}} + G_{\text{trend}}. \quad (14)$$

Attribution of partitioned flux-temperature correlations to vegetation change. We previously applied a series of analytical steps to link climate warming as a primary driver of the vegetation changes reported in this study. First, in subsection 4.6 (results in subsection ??, Table 1), we matched NDVI change with vegetation-type transitions, establishing NDVI trends as a reliable proxy for vegetation gain and loss. Next, in subsection ??, we correlated NDVI trends with long-term temperature and precipitation trends (results in subsection ?? Table 2 and Supplementary Table 10–11), implicating climate warming as a primary driver of the observed vegetation changes. To quantify the ecozone-level association between positive NDVI trends (greening) and net energy flux trends, we computed Kendall's τ using unique, non-overlapping 30-m raster cells from four independent sampling-correlation rounds, considering pixels where NDVI trends exceeded 0 or 0.0015. By default, correlations based on NDVI trends >0 were retained unless higher NDVI thresholds produced stronger and statistically significant negative associations supported by sufficient sample size. These analyses revealed a clear negative monotonic association between positive NDVI trends and net energy flux trends (Supplementary Table 19), indicating surface-cooling tendencies.

We further compared whether stable vegetation areas (no change) differed from greening areas in the magnitude of their net energy flux trends. To do this, we compared net energy flux trends ($\text{W m}^{-2} \text{yr}^{-1}$) between stable and greening vegetation zones using four sampling-comparison rounds. Each round represents an independent random draw of new, non-overlapping 30-m cells from the full study area, stratified into stable (no vegetation transition) and greening (vegetation transition; Supplementary Fig. 5) zones. Net energy flux trend values were compared using Welch's t -test and the non-parametric Mann-Whitney U test, with sample sizes varying across

rounds due to the randomized, non-replacement spatial sampling design. Across all rounds, greening areas consistently exhibited more negative (outward) net energy flux trends than stable areas, with statistically significant differences (Supplementary Table 20).

Finally, to attribute how cooling and warming energy-flux trend partitions relate to ecozone surface (skin and near surface) temperature dynamics (results in subsection 11 and Table 4), we computed Spearman rank correlations ($\rho_{\text{skin}}^{\text{cool}}, \rho_{\text{2Tm}}^{\text{cool}}, \rho_{\text{skin}}^{\text{warm}}, \rho_{\text{2Tm}}^{\text{warm}}$) between static cooling and warming energy-flux trends and multi-year (1986–2023) mean near-surface air temperature (T_{2m}) and land-surface skin temperature (T_s) for each ecozone. Correlation significance was assessed using a non-parametric Monte Carlo permutation test (99 permutations, yielding a minimum resolvable p -value ≤ 0.01). Uncertainty in the correlation coefficients was quantified using a non-parametric bootstrap-style approach based on repeated subsampling (m-out-of-n bootstrap), using 99 resamples per ecozone. Including both T_{2m} and T_s was necessary because they represent complementary aspects of the climate system: T_{2m} reflects near-surface atmospheric conditions, while T_s represents the radiative skin of the land surface that responds instantaneously to energy fluxes.

Data availability

The datasets generated and analyzed during this study are available in Zenodo under <https://doi.org/10.5281/zenodo.20656242>. The data used in this study are publicly available from the following sources. Landsat 5, 7, 8, and 9 Collection 2 Level-2 surface reflectance data were retrieved from Google Earth Engine (GEE) via `LANDSAT/LT05/C02/T1_L2`, `LANDSAT/LE07/C02/T1_L2`, `LANDSAT/LC08/C02/T1_L2`, and `LANDSAT/LC09/C02/T1_L2`. Ecozone polygons were obtained from Agriculture and Agri-Food Canada (https://sis.agr.gc.ca/cansis/nsdb/ecostrat/gis_data.html). National Fire Database (NFDB) data were obtained from the Canadian Wildland Fire Information System Data Mart (<https://cwfis.cfs.nrcan.gc.ca/datamart>). ERA5-Land monthly data were accessed through GEE (`ECMWF/ERA5_LAND/MONTHLY`). Down-scaled CRU TS v4.09 monthly climate fields were obtained from WorldClim (<https://www.worldclim.org/data/monthlywth.html>). Global Land Cover Fine Classification (GLC-FCS30D) products were accessed through GEE via `projects/sat-io/open-datasets/GLC-FCS30D/annual` and `projects/sat-io/open-datasets/GLC-FCS30D/five-years-map`. The Canada annual land-cover (VLCE v2) dataset was accessed via `projects/sat-io/open-datasets/CA_FOREST_LC_VLCE2`. Albedo data were obtained from MODIS MCD43A3 v061 (`MODIS/061/MCD43A3`) and associated quality information from MCD43A2 v061 (`MODIS/061/MCD43A2`).

Code availability

The code generated and used during this study is available in the Borealis Dataverse repository under <https://doi.org/10.5683/SP4/GHGQXM> and through GitHub at <https://github.com/danielamaogu/Boreal-Tundra-Vegetation-Increase-Effects-on-Surface-Cooling-and-Warming-Tendencies>.

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Author contributions

D.C.A. conducted the research, literature review, wrote the article, plotted results, and created illustrations. P.M. was involved in ideation, conceptualization, setting objectives, methodological approaches, providing computational resources, editing, and review of the manuscript. K.B.D., E.O., and J.P. helped with data visualization and manuscript editing. R.B., P.M., L.U.A., J.M.-C., and Y.L. were involved in funding acquisition, manuscript review, and overall supervision. All authors contributed to the discussion and conceptualization of the study.

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Competing interests

The authors declare no competing interests.

Additional information

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