RESEARCH PAPER

A REGIONAL-SCALE ESTIMATE OF THE SOIL ORGANIC CARBON ISOTOPIC COMPOSITION (δ13C) AND ITS ENVIRONMENTAL DRIVERS: CASE STUDY OF THE BAIKAL REGION

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ABSTRACT. Modern plants and surface soil $\delta^{13}C$ values from 95 sites in the Baikal region were obtained for the first time and were used to establish relationships with regional environmental factors. Studied sites were distributed along the elevation gradient from 403 to 2315 m, which defined a strong landscape and climatic gradients encompassing mountain tundra, subalpine grasslands, mountain taiga, subtaiga, and steppe. $\delta^{13}C$ values of soil organic matter (SOM) varied from –29.50 to –22.98‰. This result showed that the stable C isotopic composition of the surface soils was mainly determined by $\delta^{13}C$ values of C3 plants (vary from –33.0 to –24.5‰) and C isotope fractionation during stabilization of plant-derived C into SOM. The $\delta^{13}C$ values of modern plants and surface soils were negatively correlated with mean annual and growing season precipitation ($p<0.05$), confirming that precipitation is the primary factor determining SOM's stable C isotopic composition in the Baikal region. A distinct increase in the $\delta^{13}C$ values with decreasing mean annual and growing season precipitation was found with a slope of –0.42‰/100 mm and –0.97‰/100 mm, respectively. Temperature had no significant effect on the spatial distribution of SOM $\delta^{13}C$ values at the regional scale but played an important role in the severe environments of mountain tundra (the coldest and wettest) and steppes (the warmest and driest). Such conditions strongly impacted SOM $\delta^{13}C$ values by influencing plant species composition and soil microbiological activity. As a result, the organic matter of these soils is characterized by the highest $\delta^{13}C$ values. The SOM of taiga soils formed under a favorable combination of temperature and precipitation was characterized by the lowest $\delta^{13}C$ values.

KEYWORDS: carbon, $\delta^{13}C$, C3 plants, climatic conditions, mountain areas, Southeastern Siberia

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INTRODUCTION

Understanding the spatio-temporal dynamics of soil organic carbon (SOC) and its driving factors is critical to assessing the feedback between the terrestrial C cycle and climate (Kudeyarov et al. 2007). This issue is especially relevant for mountain areas most affected by climatic changes (Pepin and Lundquist 2008). The overall impact of individual environmental factors on SOC dynamics is well-known. At the same time, analysis of the SOC dynamics in mountain regions is complicated by the high heterogeneity of bioclimatic conditions.

The Baikal region, where the fate of organic C in soils has been relatively poorly understood, is no exception. This circumstance brings to the fore the need to assess the main regional factors that determine soil C dynamics in various bioclimatic conditions. The problem is also relevant because the area under consideration is one of the planetary regions most prone to global warming (Shimaraev et al. 2002; Mackay et al. 2016). These climatic changes are bound to affect the C balance in regional ecosystems (Mackay et al. 2016).

Analysis of the stable C isotopic composition is an important approach to studying SOC dynamics (Dawson et al. 2002). Despite the significant potential of isotope studies, there are few examples of learning the $\delta^{13}C$ of SOM in the Baikal region and adjacent territories of Southeastern Siberia. Such studies cover individual soil types (Menyailo and Hungate 2006; Tsybenov et al. 2016; Golubtsov 2020; Tsybenov et al. 2022; Golubtsov et al. 2023a) as well as altitudinal profiles in depressions of the Baikal rift zone (Andreeva et al. 2013; Golubtsov et al. 2022; Andreeva et al. 2022; Golubtsov et al. 2023b). Most of the territory's soil remains unstudied regarding the geochemistry of stable carbon isotopes.

Filling this gap is also important in the context of the lack of knowledge about the soils in the vast inland regions of Northern Eurasia, which contain almost a fifth of the world's total soil carbon reserves (Kudeyarov et al. 2007; Schepaschenko et al. 2013). Most syntheses describing the global patterns of $\delta^{13}C$ value distribution in the SOM and organic matter-producing plants (Kohn 2010; Diefendorf et al. 2010; Rao et al. 2017; Cornwell et al. 2018) almost do not take into account data on the indicated region.

This study aimed to assess the spatial variability in SOC content and its stable C isotopic composition to identify the main factors controlling these variables in the Baikal region. We hypothesized that (1) climatic heterogeneity of the study area is one of the leading factors influencing variations in the $\delta^{13}C$ values of SOM; (2) precipitation has the main effect on the SOM $\delta^{13}C$ values; and (3) temperature has no significant effect on the spatial distribution of SOM $\delta^{13}C$ values at the regional scale but plays an important role in extreme climatic conditions.

This study could then provide clues on soil carbon dynamics in sharply continental regions and the potential feedbacks to climatic changes.

MATERIALS AND METHODS

Study area: geographical localization and orography

The Baikal region (Fig. 1) is a vast area in Southeastern Siberia that is strongly heterogeneous regarding physiographic conditions (Mikheev and Ryashin 1977).

The western part of the study area (Fore-Baikal region) is confined to the Siberian platform and its margins. The Eastern Sayan mountains' foothill plains and intermountain depressions, characterized by a predominance of hilly relief and absolute heights of 200-500 m, orographically represent this area (Plateaus… 1971). Deeply incised river valleys dissect this area.

The central part (Sayan-Baikal highlands) of the study area is characterized by extensive high plateaus and alpine-type mountain ranges with absolute heights up to 2300-3000 m or more, separated by deep depressions (Highlands… 1974).

The eastern part (Transbaikal region) is the area of ancient folding characterized by middle mountains, plateaus, and flat (in the southeast) relief. The territory is composed of deep intermountain depressions alternating with mountain ranges. Oriented from northeast to southwest, the mountain ranges are characterized by prevailing absolute heights of 1000-1500 m (Highlands… 1974). The northeast direction dominates the intermountain depressions, which are characterized by predominantly gently sloping terrain divided by hills and low ridges.

Climatic conditions, vegetation cover, and general patterns of pedogenesis

The heterogeneous topography of the study area has a significant impact on climatic conditions, which results in a high spatial diversity of landscapes (Plysnin et al. 2018), an altitudinal distribution of vegetation (Belov et al. 2018) and a diversity of soils (Koposov 1983; Kuz'min 1988). There is a wide range of vegetation types in the region, such as high-mountain (mountain-tundra and subalpine grasslands) and boreal (taiga and steppe) (Belov et al. 2018). Taiga forms the study area's main landscape background. Mountain tundra and steppe are localized as more, or less isolated areas confined to the most highly elevated surfaces (in the case of mountain tundra); high terraces and slopes of river valleys and the bottoms of depressions (in the case of steppes) (Mikheev and Ryashin 1977).

The climate of the study territory is sharply continental, characterized by large daily and annual (up to 30-45°C) temperature fluctuations. The amplitude of temperature fluctuations and degree of climate continentality increase from west (Fore-Baikal) to east (Transbaikal) (Zhukov 1965). The annual air temperature ranges from 0.5 to –4.2°C. The temperature of the coldest month (January) is from –18°C to –35°C, the warmest (July) is 10-15°C at altitudes of 1000-2000 m and 15-20°C in low-elevation sites (Scientific and Applied Handbook… 1989; 1991). The low annual radiation balance causes insufficient heat supply to landscapes (Zhukov 1965; Chimitdorzhieva 2016). The sum of active temperatures (above 10°C) varies from 600°C in the mountain tundra landscapes to 1500-2000°C on the plains and in basins, within the area of distribution of steppe and subtaiga (Zhukov 1965; Scientific and Applied Handbook… 1989; 1991).

The duration of periods with negative temperatures significantly exceeds those with positive ones, affecting the soil temperature regime (Trofimova et al. 2019). There is a strong cooling and deep freezing of soils during the winter and long-term preservation of negative temperatures in the soil profile (Koposov 1983; Abasheeva 1992).

Significant spatial heterogeneity in atmospheric precipitation is one of the most critical climatic features due to the complex topography of the study area (Trofimova et al. 2019). The predominantly northeastern orientation of the ridges and depressions, combined with the prevailing north-western flow of air masses, contribute to condensation and precipitation on the windward slopes. Due to mountain-barrier and basin effects, leeward slopes and intermountain basins are significantly less moistened (Zhukov 1965).

A heterogeneous distribution of precipitation is noted during the year. The first half of the vegetation season (May-June) is characterized by aridity; for July-August, moisture is optimal. At the same time, most precipitation falls in the summer (up to 75% of annually) (Scientific and Applied Handbook… 1989; 1991). The soil moisture fluctuates significantly during the vegetation season, and the soil is intensely dried in the first half (Koposov 1983).

Thus, there is a pronounced contrast between the dry and wet seasons of the warm period, which affects the short periods of optimal combination of positive temperatures and soil moisture for microbiological activity. This circumstance limits the intensity and degree of plant residue decomposition (Volkovintser 1978; Abasheeva 1992; Chimitdorzhieva 2016) and humification (Kuz'min 1988; Chimitdorzhieva 2016).

The most severe variants of the described regional specifics of soil temperature and water regimes are typical of the soils of the mountain tundra. It is also found in steppe soils in a somewhat milder form (Kuzmin 1988; Abasheeva 1992; Chimitdorzhieva 2016). Steppe landscapes are characterized by a dryness index (the ratio of the surface annual radiation balance to the amount of heat required to evaporate the annual amount of precipitation) in the range of 1.5-1.8 which indicates a large moisture deficit (Zhukov 1965). The forming conditions of soils in taiga landscapes are more favorable. Here, the dryness index ranges from 0.45 to 1.0 indicating sufficient and even excessive moisture (Budyko 1971). Such humidity contributes to the optimal water balance of plants that form SOM and high soil microbiological activity. At the same time, the temperature decrease with increasing altitude is often offset by temperature inversions, which are one of the essential features of the regional climate (Trofimova et

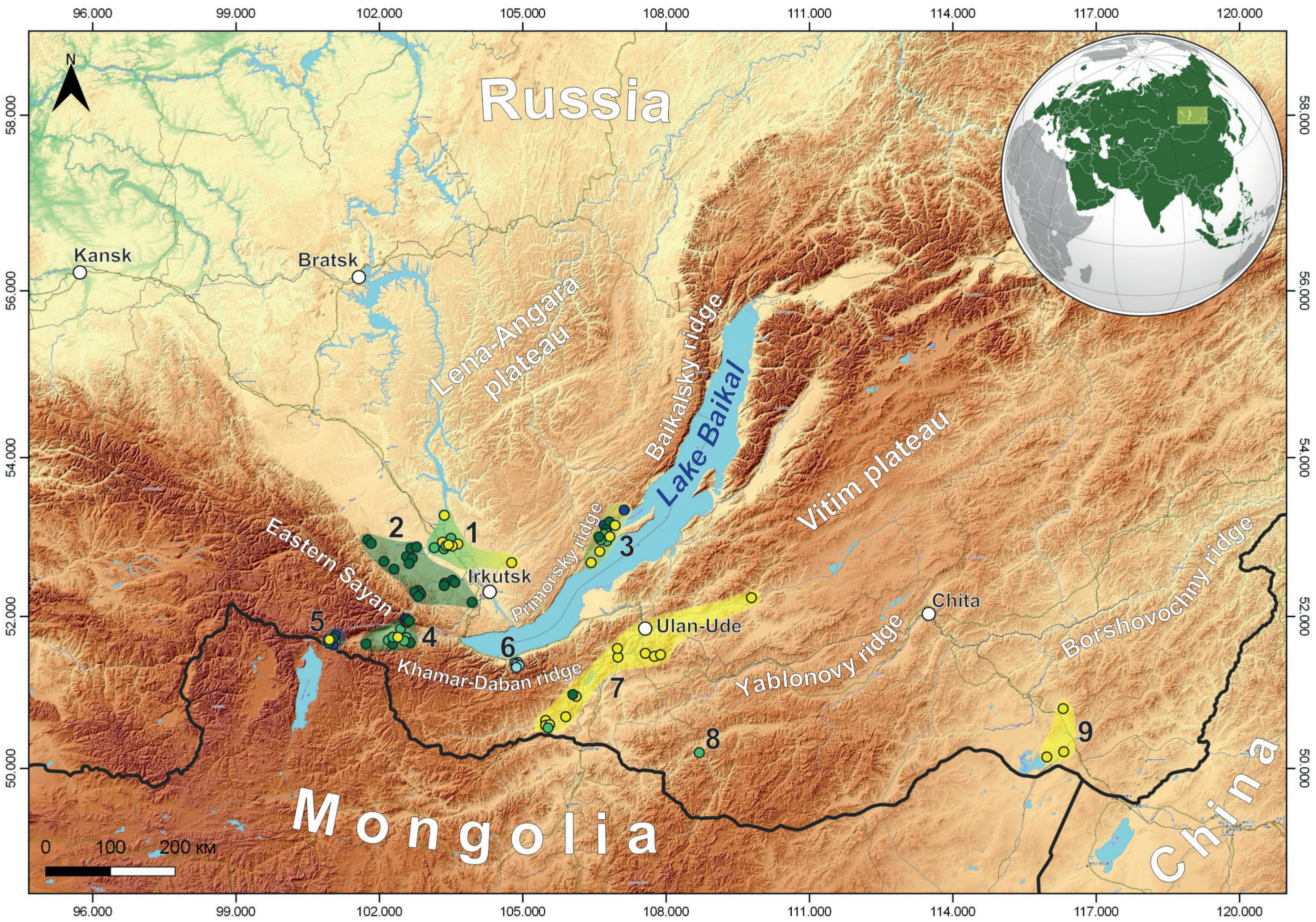

Fig. 1. Location of the study area and main sampling sites. Landscape types: yellow – steppes; dark green – mountain taiga; light green – subtaiga; light blue – subalpine grasslands; dark blue – mountain tundra

al. 2019). As a result, the period with positive temperatures in some taiga areas may exceed that even on the plains and in the bottoms of the depressions.

Study sites

The study sites include (Fig. 1; Table A): steppe and subtaiga of the Irkutsk-Cheremkhovo plain and Cis-Baikal depression (1); mountain taiga of the foothills of the Eastern Sayan mountains (2); mountain tundra, mountain taiga, subtaiga, and steppe of the Olkhon region and the southeastern side of the Primorsky ridge (3); Tunka (4) and Mondy (5) basins of the Baikal rift zone; subalpine grasslands of the Khamar-Daban ridge at the southern part of Lake Baikal (6); steppes and subtaiga of intermountain basins of the Selenga middle mountains (7) and the Chikoy river valley (8); steppes of the Southeastern Transbaikal region (9).

The study is based on SOM investigation data from 95 plots. Each plot (10 x 10 m) was characterized by a complex environmental description that includes the characteristics of phytocenoses, geomorphological conditions, and morphogenetic analysis of soils. Soil taxonomic affiliation was determined according to (IUSS Working Group WRB 2015). General information on the sampling sites is shown in Table A. A more detailed description of sites and soil characteristics can be found in previous works (Golubtsov et al. 2021; 2022; 2023 a, b).

Sampling scheme, sample preparation, and laboratory analyses

The fieldwork was carried out in July of 2020-2022. We collected fresh, mature leaves of the dominant species, constituting 80% of the biomass in each plot. Leaf and mineral soil samples were dried to an air-dry state and homogenized. Leaves were oven-dried at 70°C for 60 hours until weight constancy and ground. Samples of mineral soil horizons were sifted through a 1-mm sieve with the subsequent removal of fine roots and washed using 1M HCl to remove carbonates.

The total organic carbon (TOC) content was determined by pyrolysis of samples on an elemental analyzer CHNS Vario Isotope Cube (Elementar, Germany). The stable carbon isotopic composition ($\delta^{13}C$) was measured on a set of equipment CHNS-analyzer Vario ISOTOPE Cube mass spectrometer Isoprime precision IRMS (Elementar, UK) connected in continuous flow mode. The measurements were conducted at the Center for Collective Use "Laboratory of Radiocarbon Dating and Electron Microscopy" Institute of Geography of the Russian Academy of Sciences. The results were expressed in ‰ for the VPDB standard. The standard deviation for measurement of the $\delta^{13}C$ is less than 0.1‰.

Meteorological data

Climatic data for each sampling location were assessed both for the vegetation season (May-September) and for the year as a whole based on the WorldClim 2.0 database (spatial resolution 30") (Fick et al. 2017), monthly data of bias-corrected ERA5-Land reanalysis (reference period from 1981 to 2021) (Muñoz Sabater 2019) with spatial resolution 0.1°×0.1°, which approximately corresponds to 11.0 km in latitude and 7.2 km in longitude for Siberia. For each investigated site, the time series at the nearest mode of the reanalysis grid was used.

The data of microclimatic observations obtained using automatic recorders Elitech RC-51H and RC-4HC (Golubtsov et al. 2022; Golubtsov et al. 2023b) were used for sites located on altitude profiles within mountain-depression areas (Fig. 1, areas 3-5) where the climate changes noticeably at reasonably close distances that are beyond the spatial resolution of the reanalysis. The air temperature, relative humidity (at a height of 2 m above the surface), and soil surface temperature were measured. The temperature measurement accuracy is 0.1°C, and the relative humidity is 3%. The observations have been carried out at 1-hour intervals since 2013. In some cases, the data were corrected based on annual precipitation, taking into account data from the nearest meteorological stations (Scientific and Applied Handbook… 1989; 1991).

Data analysis

Data processing was performed using MO Excel and PAST 4.03 (Hammer et al. 2001). For each data set, a check was carried out for compliance with the normal distribution using the Shapiro-Wilk and Anderson-Darling tests, and according to this, parametric or non-parametric statistical methods were selected. For samples of less than 30 measurements, nonparametric methods and tests were used.

A one-way ANOVA with Tukey's HSD test was conducted for the estimation of differences in $\delta^{13}C$ values in the SOM of the surface horizons of soils in different landscape types. The Kruskal-Wallis H-test (a nonparametric analogue of ANOVA) with Dunn's post hoc test was used to estimate differences in $\delta^{13}C$ values in different life forms of plants and lichens. A paired correlation analysis with the Spearman correlation coefficient (Spearman's rs) was conducted to identify the main climatic factors (air temperature and precipitation for the year and the growing season, altitude) influencing the variation in $\delta^{13}C$ values in plant biomass of different life forms. To confirm the dependence of $\delta^{13}C$ values in the SOM of the surface horizons of soils on the amount of precipitation for the year and growing season, paired linear regressions were calculated. The linear regression residuals were close to a normal distribution.

RESULTS

Differences in climate

Both mean annual temperatures and mean temperatures of the growing season decreased in the direction from the steppes to the mountain tundra (Fig. 2). In many cases, temperatures are quite similar in a large group of steppe, subtaiga, and taiga plots, along with this general trend. In terms of temperatures, the mountain-taiga area was the most heterogeneous. The lowest temperatures here were recorded in high-elevated sites that form in the mountain-depression conditions of the Baikal rift zone (Primorsky ridge (3), the mountains surrounding Tunka (4) and Mondy (5) basins, the Khamar-Daban ridge (6)). A similar differentiation was also typical for the steppes, where low temperatures were observed in the highest-located steppes of the Olkhon region (3), the Mondy basin (5), and the Selenga middle mountains (7).

The studied sites were also heterogeneous in terms of atmospheric precipitation (Table A). The highest mean annual and growing season precipitation was observed in mountain tundra, subalpine grasslands, and high-elevated mountain taiga sites (Fig. 2), especially on the Khamar-Daban ridge (6), windward slopes, and foothills of the Eastern Sayan mountains (2). The leeward slopes and intermountain depressions were significantly drier due to the mountain barrier and basin effects. This was most clearly indicated in the mountain taiga on the southeastern slope of the Primorsky ridge (3) and the Tunka mountains (4). The lowest mean annual precipitation within the study area was observed in the steppe landscapes of the intermountain basins of the southern part of the Selenga middle mountains (7), the Mondy basin (5), and the western coast of Lake Baikal at the foot of the Primorsky ridge (3). Here, the annual precipitation often only slightly exceeds 200-250 mm.

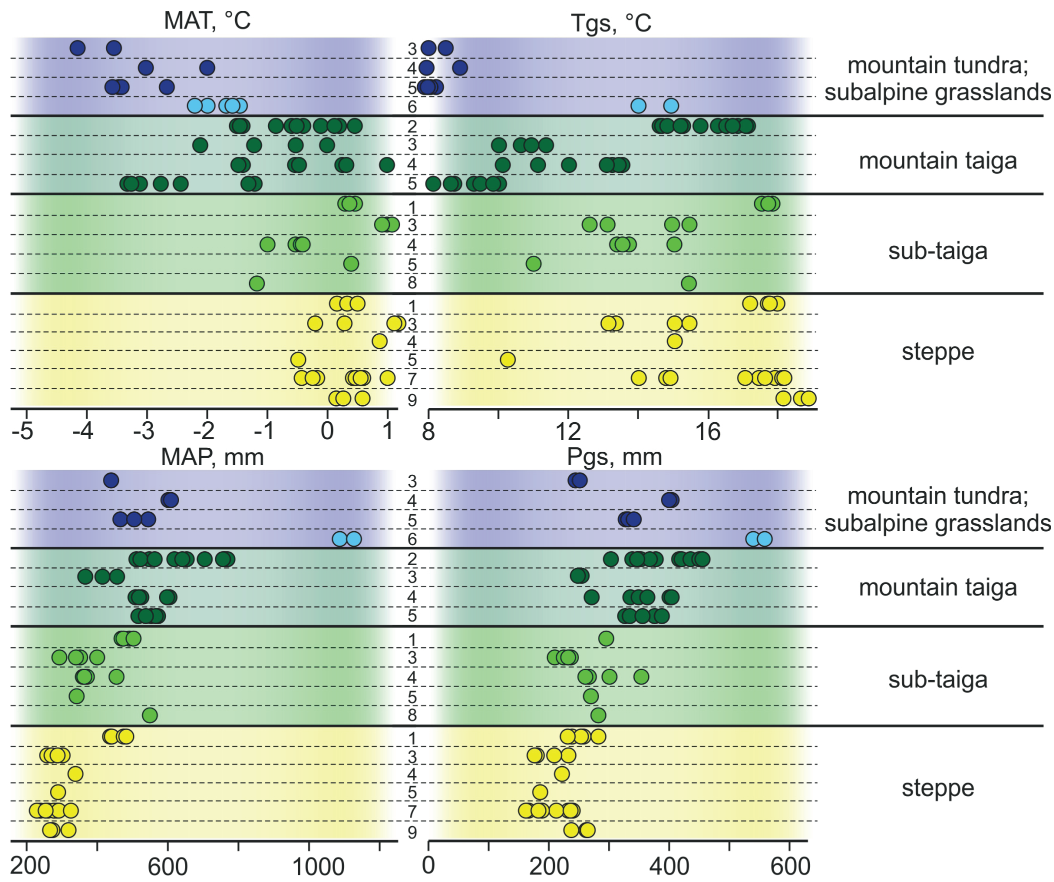

Fig. 2. Mean annual and mean growing season temperature and precipitation of landscapes in the Baikal region. The numbers in the center of the figure correspond to the areas on Fig. 1

Microclimatic observations within individual altitudinal profiles on the Primorsky ridge (3) (Golubtsov et al. 2022) and in the Tunka mountains (4) (Golubtsov et al. 2023b) indicated a linear increase in temperature, the duration of the growing season, and a decrease in precipitation from high to low-elevated sites. At the same time, local variations were associated with the density of the vegetation cover, the landform features, and local atmospheric circulation (Golubtsov et al. 2022; 2023b).

Differences in TOC

High-elevated mountain tundra and taiga soils were characterized by an increased TOC content in the surface mineral horizons compared to low-elevated steppe soils (Fig. 3). High TOC variation (from 2.5 to 45%) in mountain taiga soils was one of the most essential features. This feature is associated with the distribution of taiga landscapes in the Baikal region, both in terms of their vast area and the range of variations in environmental factors (temperature, precipitation, vegetation cover, etc.).

On the contrary, the TOC content in steppe soils is localized in rather narrow ranges (from 0.5 to 12%) (Fig. 3; Table A). The similarity of the steppe soils is probably due to their localization in the region, mainly within the basins, and the pronounced limitations of their areas in terms of precipitation.

The tundra and subalpine soils are very diverse in terms of TOC, which is largely due to the high heterogeneity in temperature and precipitation conditions in the mountains.

Variations in $\delta^{13}C$ values of plant organic matter

$\delta^{13}C$ values were determined for 38 species of higher vascular plants, two species of moss, and two species of lichen growing in the study area. Table 1 presents data only for the main plant species-edificators of landscape types. The $\delta^{13}C$ values in leaves and needles of higher vascular plants were low and varied from −32.6 to −25.5‰ (Fig. 4a).

The range of $\delta^{13}C$ values varied from −32.6 to −32.3‰ for mosses and −24.6 to −23‰ for lichens (Table 1). The mountain tundra and steppe plant species were the most enriched in ^{13}C (mean value −27.9 ± standard deviation 1.3‰ and −28 ± 0.8‰, respectively), the mean value $\delta^{13}C$ in taiga species was lower by 2.2 and 2.1‰.

Analysis of $\delta^{13}C$ values in various life forms of plants showed (Fig. 5) that the greatest range of values was observed in trees (a difference of 6.77‰) and shrubs (5.4‰), and the minimum (0.37‰) in mosses.

When comparing the mean $\delta^{13}C$ values in leaf samples, litter, and the SOM of the surface horizons of soils, it was revealed that the $\delta^{13}C$ value of organic matter became higher during the transition from plants to litter at 0.1-3.9‰, and from litter to the surface humic horizon at 0.3-2.0‰ (Fig. 6). The maximum difference in $\delta^{13}C$ values was observed mainly in the mountain taiga, which may be due to the mixed composition of the litter, including both coniferous and deciduous litter, as well as to varying degrees of their decomposition.

This pattern is not observed at several plots where the plant sample was enriched with ^{13}C compared to the litter, or $\delta^{13}C$ value of the humic horizon SOM was lower than that of the litter (Fig. 6). This difference from the general pattern is probably due to the fact that the soil on these plots was formed under conditions of changes in taiga and steppe landscapes, determined by the dynamics of climatic conditions. The disruption of the ecological balance between the litter and other components of the landscape as a result of the periodic expansion and contraction of the taiga area could be one of the factors that caused the observed differences in the stable carbon isotopic composition of the upper part of the soil organic profile (Golubtsov et al. 2022).

Variations in $\delta^{13}C$ values of SOM in regional soils

SOM of the study soils showed a considerable variation in $^{13}C/^{12}C$ isotopic ratios, covering a wide range of $\delta^{13}C$ values characteristic of C3 plants (O'Leary 1988) (Fig. 4). The topsoil organic matter had $\delta^{13}C$ values from −29.5 to −22.98 ‰ (Fig. 4; Table A).

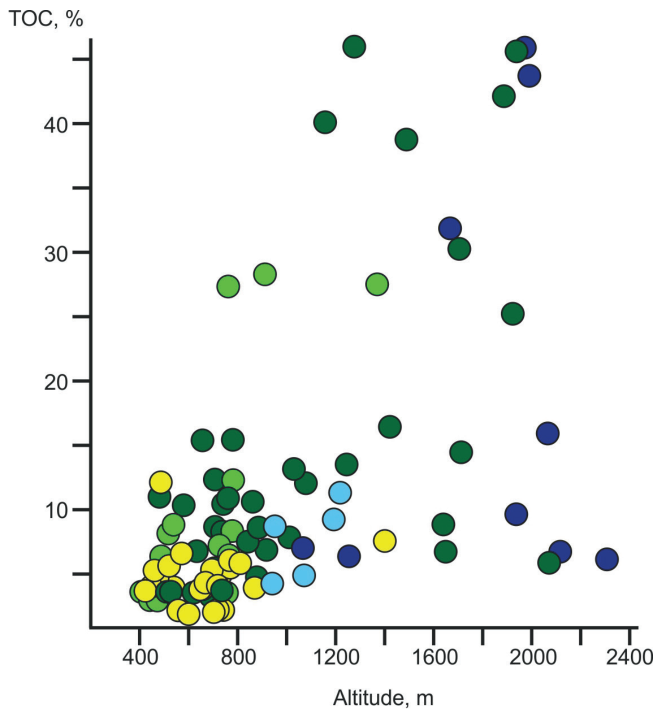

Fig. 3. Altitudinal variations in topsoil TOC content in different landscapes (yellow – steppes; dark green – mountain taiga; light green – subtaiga; light blue – subalpine grasslands; dark blue – mountain tundra)

Table 1. $\delta^{13}C$ values of dominant species of vascular plants and lichens at sampling sites

№	Plant species	$\delta^{13}C$, ‰ (mean ± st. dev.)	Number of samples	Landscape type
Trees				
1	*Betula pendula* Roth	-31.04 ± 0.86	4	Taiga/subtaiga, successional stage
2	*Larix sibirica* Ledeb.	-28.12 ± 1.86	4	Light coniferous taiga
3	*Pinus sibirica* Du Tour	-29.14 ± 1.99	3	Dark coniferous taiga
4	*Pinus sylvestris* L.	-28.74 ± 1.49	9	Light coniferous taiga and subtaiga
5	*Populus tremula* L.	-29.69 ± 1.53	4	Taiga/subtaiga, successional stage
Shrubs				
6	*Betula fruticosa* Pall.	-29.57 ± 1.39	3	Mountain tundra
7	*Duschekia fruticosa* (Rupr.) Pouzar	-29.66 ± 1.26	3	Taiga
8	*Rhododendron parvifolium* Adams	-26.72	1	Mountain tundra
9	*Rhododendron dauricum* L.	-30.87 ± 0.94	3	Light coniferous taiga, subtaiga
Subshurbs				
10	*Dryas oxyodonta* Juz.	-29.63	1	Mountain tundra
11	*Empetrum nigrum* L.	-27.33	1	Mountain tundra
12	*Vaccinium vitis-idaea* L.	-30.50 ± 0.83	9	Taiga
13	*Vaccinium myrtillus* L.	-32.14 ± 0.55	4	Taiga
Herbs				
14	*Agropyron cristatum* (L.) Gaertn.	-28.57 ± 0.69	3	Steppe
15	*Artemisia frigid* Willd.	-28.43 ± 1.04	3	Steppe
16	*Koeleria cristata* (L.) Pers.	-28.16 ± 1.76	2	Steppe
17	*Maianthemum bifolium* (L.) F.W. Schmidt	-30.84 ± 0.42	2	Taiga
18	*Poa botryoides* (Trin. ex Griseb.) Kom.	-29.56 ± 0.52	2	Steppe
Mosses				
19	*Hylocomium splendens* (Hedw.) Bruch et al.	-32.46	1	Taiga
20	*Pleurozium schreberi* (Willd. ex Brid.) Mitt.	-32.45 ± 0.17	4	Taiga
Lichens				
21	*Flavocetraria nivalis* (L.) Kärnefelt et A. Thell	-23.03	1	Mountain tundra
22	*Cladonia rangiferina* (L.) F. H. Wigg.	-24.58	1	Mountain tundra

A non-linear distribution of $\delta^{13}C$ values in the altitudinal profile was noted, which was strongly correlated with the landscape type (Fig. 4b). One-way ANOVA with Tukey's HSD test showed differences ($p \leq 0.0001$) in $^{13}C/^{12}C$ ratios in the surface horizons of steppe soils (Table 2) from subalpine grasslands, mountain taiga, and subtaiga forests. The differences between $\delta^{13}C$ values in mountain tundra, subtaiga, and steppe soils have not been identified ($p > 0.05$, Table 2).

Even though subalpine grasslands like mountain-tundra soils were formed under rather severe climatic conditions, they were much lighter in stable C isotopic composition ($p = 0.001$) and were distinguished as an independent group.

The highest variations in $\delta^{13}C$ values were observed in mountain taiga soils (from –27.06 to –24.72‰) associated with a reasonably wide taiga distribution under different climatic conditions (from wet dark-coniferous forests to drier pine forests). At the same time, differences were revealed between the soils of the mountain taiga and subtaiga ($p < 0.05$), which occupied a transitional position between the taiga and the forest-steppe zone (Fig. 4, Table 2).

DISCUSSION

Control of C3 and C4 plants on $^{13}C/^{12}C$ ratios of SOM

The $^{13}C/^{12}C$ isotopic ratio in the plant tissues is one of the main predictors of SOM's stable carbon isotopic composition (Ehleringer 2005). It is essential to pay

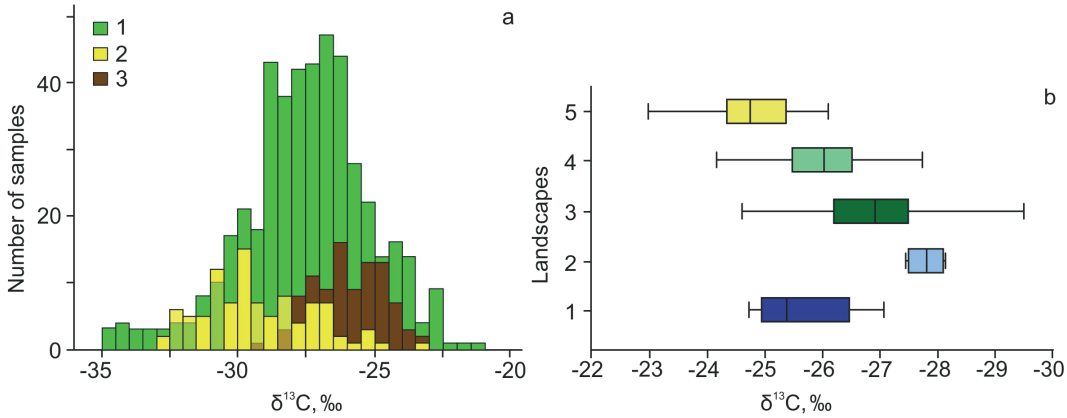

Fig. 4. Histogram of the δ13C values of surface soils and plants in the Baikal region (a). Table of symbols: 1 – data of C3 plants come from O'Leary (1988); 2 – δ13C values of study plant species; 3 – δ13C values of surface humic soil horizons. The δ13C values of surface soil samples in different landscape types in the Baikal region (b): 1 – mountain tundra; 2 – subalpine grasslands; 3 – mountain taiga; 4 – subtaiga; 5 – steppe

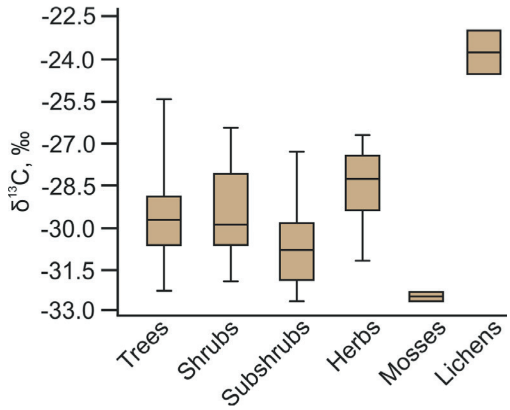

Fig. 5. δ13C values in different life forms of higher vascular plants, mosses and lichens

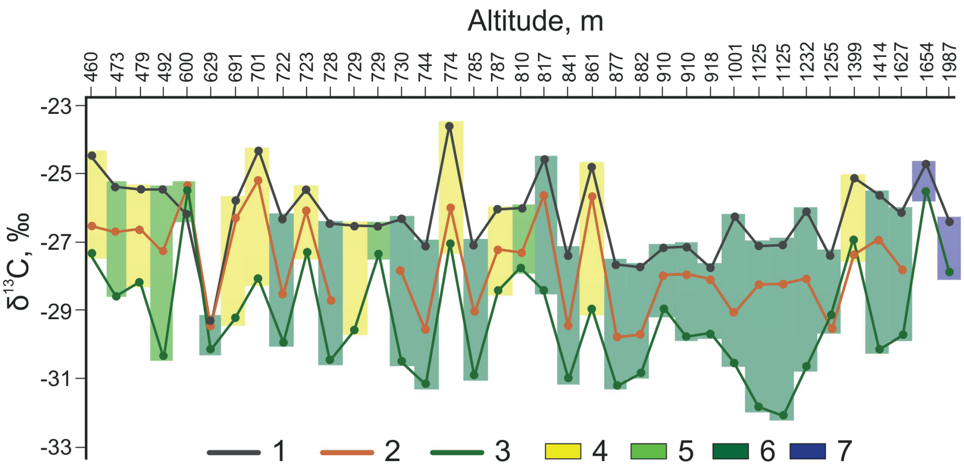

Fig. 6. The mean δ13C values: 1 - in the SOM of the surface humic horizons of soils, 2 – in litter samples, 3 – in leaf samples. Landscape types: 4 – steppe, 5 – subtaiga, 6 – mountain taiga, 7 - mountain tundra

Table 2. The results of the estimation of differences in the δ13C values of surface soil samples in different landscape types in the Baikal region by ANOVA

Landscape type	δ13C, ‰ (mean ± st. dev.)	Tukey's Q below the diagonal, p-value above the diagonal				
		mountain tundra	subalpine grassland	taiga	subtaiga	steppe
mountain tundra	–25.39 ± 0.82		0.001	0.005	0.912	0.051
subalpine grassland	–27.8 ± 0.32	5.7		0.231	0.002	< 0.0001
taiga	–26.8 ± 0.98	5.0	3.0		0.007	< 0.0001
subtaiga	–25.96 ± 0.84	1.3	5.4	4.9		0.0001
steppe	–24.7 ± 0.82	3.9	9.2	12.9	6.6	

attention to some issues, given the rather heterogeneous plant cover of the study area.

The first is the relative abundance of plants with different photosynthetic pathways (Ehleringer 2005; Rao et al. 2017). Higher terrestrial plants are divided into three main groups according to their photosynthetic pathways: C3 (Calvin-Benson), C4 (Hatch-Slack) and CAM (Crassulacean acid metabolism). CAM plants will not be considered in this paper because they are only present in highly specialized ecosystems (for example, deserts). C3 plants include almost all trees, shrubs, and cool-season grasses, while C4 plants comprise mainly warm-season grasses (Ehleringer 2005; Sage 2005; Rao et al. 2017). C3 and C4 plants have $\delta^{13}C$ values ranging from -20‰ to -32‰ and -10‰ to -17‰, respectively.

Plant phylogenetic lines that are common mainly in the tropics and subtropics tend to use C4 photosynthesis (Sage 2005). However, the flora of the study region has distinct boreal features (Belov et al. 2018). Nevertheless, according to various estimations, the relative abundance of C4 plants in the plant communities of arid and semi-arid areas of Central Asia can reach 30% (Ehleringer 2005).

The steppes presented in the study area had floristic similarities with the Central Asian and East Asian steppes and belonged to the Mongolian-Chinese florogenetic type (Belov et al. 2018). Several authors noted the presence of C4 plants in the steppe communities of the study area (Ivanova et al. 2019; Golubtsov et al. 2022). These findings agreed with the nature of C4 plants adapted to high-intensity photosynthesis at elevated temperatures and light, mainly due to the possibility of fixing CO_2 at high stomatal resistance under water stress conditions (Sage 2005).

However, an abundance of plants with C4 photosynthetic pathways in the steppes of the Baikal region was insignificant (Ivanova et al. 2019). For example, on several plots in the steppes of the Olkhon region (3), plants of this group were represented singly, or their projective cover did not exceed 1% (Golubtsov et al. 2022). Hence, C4 plants did not contribute to variations in the study area's $^{13}C/^{12}C$ isotopic ratios of SOM. The obtained $\delta^{13}C$ values of SOM supported this assumption since they indicated the C3 plant biomass as the primary source of SOM (Fig. 4).

Variations in the $\delta^{13}C$ values of SOM can also correlate with stable C isotope fractionation features in C3 plant tissues depending on their growing conditions (Dawson et al. 2002; Diefendorf et al. 2010; Rao et al. 2017). It was noted above that the SOM $\delta^{13}C$ values were distributed nonlinearly depending on the elevation. The SOM of low-elevated (steppe) and high-elevated (mountain tundra) sites was most enriched in ^{13}C. The lowest $\delta^{13}C$ values were observed in the SOM of mid-elevated mountain taiga landscapes (Fig. 4). A similar nonlinear distribution of $\delta^{13}C$ of plant biomass depending on altitude was observed (Fig. 6).

Significant height amplitudes caused pronounced changes in the air temperature and precipitation (Fig. 2), determining plant communities' species composition and growth conditions. It is also known that $\delta^{13}C$ can vary in C3 plants of different layers and life forms depending on moisture availability, temperature, and other factors (Dawson et al. 2002; Tiunov 2007). The results of the Kruskal-Wallis H-test (H=35.71, $p < 0.0001$) revealed statistically significant differences in the $^{13}C/^{12}C$ isotopic ratios between the different life forms of the studied plants and lichens (Table 3). Differences were observed between the upper and lower layers ($p < 0.05$): subshrubs and mosses were lighter in composition of stable C isotopes than trees and shrubs (Fig. 5). The observed isotopic differences could primarily be associated with the effect of the forest canopy when, in dense forest stands, the concentration of ^{13}C in plant leaves had a pronounced vertical gradient and was minimal in plants of the lower layers (Tiunov 2007). The effect was associated with the peculiarities of photosynthesis under shading conditions, as well as the fixation of ^{13}C-depleted carbon dioxide released by soil and litter by plants.

Higher $\delta^{13}C$-values were found for herbs (Table 1, Fig. 5) than for other groups of plants except lichens. This was because the herb sample was represented mainly by steppe species that grow in drier conditions and are unaffected by the forest canopy effect. Of the taiga species, the sample included *Maianthemum bifolium* (Table 1), which had the lightest isotopic composition among herbs.

A separate group with a high level of significance ($p < 0.05$) included lichens (Fig. 5, Table 3) that dominated the plant cover of the mountain tundra on the Primorsky ridge (3). The increase in ^{13}C content in lichens compared to vascular plants in the study area was associated with the peculiarities of C fractionation in this group of organisms (Byazrov et al. 2010). The $\delta^{13}C$ values we obtained indicated insufficiently favorable conditions for lichens in the mountain tundra and were consistent with the data obtained for *Flavocetraria nivalis* in the tundra in Western Yamal, where the $\delta^{13}C$ varied from –24.3 to –24.4 ‰ (Kuznetsova et al. 2020).

A paired correlation analysis was conducted to identify the main climatic factors influencing the variation in $\delta^{13}C$ values in plant biomass. It was revealed that $\delta^{13}C$ of coniferous tree species negatively correlates with annual precipitation (Spearman's rs is –0.77, $p = 0.0001$). For deciduous tree species, such a correlation has not been identified.

Table 3. The results of the estimation of differences in the $\delta^{13}C$ values between the different life forms of plants and lichens by Kruskal-Wallis H-test

Plant life form	$\delta^{13}C$, ‰ (mean ± st. dev.)	Dunn's post hoc test, raw p-value					
		Trees	Shrubs	Subshurbs	Herbs	Mosses	Lichens
Trees	–29.50 ± 1.63		0.68	0.02	0.02	0.001	0.02
Shrubs	–29.23 ± 1.68	0.68		0.02	0.11	0.0009	0.04
Subshurbs	–30.74 ± 1.38	0.02	0.02		< 0.0001	0.09	0.001
Herbs	–28.50 ± 1.19	0.02	0.11	< 0.0001		< 0.0001	0.15
Mosses	–32.45 ± 0.15	0.001	0.0009	0.09	< 0.0001		0.0001
Lichens	–23.81 ± 1.10	0.02	0.04	0.001	0.15	0.0001	

A strong positive correlation (Spearman's rs = 0.71, p = 0.002) with the absolute height of the area and an average negative correlation (Spearman's rs = –0.59, p = 0.02) with air temperature for the year and the growing season were found for shrubs. The shrub sample includes species from each studied landscape type except steppe. Therefore, this plant group demonstrates that under low-temperature conditions, processes of inhibition of photosynthesis occur, which, like dry conditions, led to the enrichment of plant tissues with ^{13}C. Subshrubs also showed similar correlations, although they were significantly less pronounced ($p > 0.05$). The lack of a reliable correlation of $\delta^{13}C$ with climatic parameters in subshrubs was because the sample was represented mainly by taiga species that grow in the low layer and were most susceptible to the forest canopy effect. Also, no significant correlation existed between $\delta^{13}C$ of herb biomass and climate parameters.

As a result of the analysis, it was established that for plants in taiga landscapes growing in subordinate layers, the effect of the forest canopy often overlapped with the influence of climatic factors on the $^{13}C/^{12}C$ isotopic ratios. However, the climatic conditions caused by altitudinal and latitudinal zonality largely determined the structure of plant communities. As a result, we considered the climatic heterogeneity of the study area to be one of the leading factors influencing variations in the composition of stable carbon isotopes of SOM.

Relationship between natural ^{13}C abundance in SOM and climate

With changing altitudes, the change of such climatic factors as temperature and precipitation were the most prominent. These climatic parameters were the main driving factors of C isotopic discrimination in C3 plant tissues during photosynthesis (Dawson et al. 2002; Diefendorf et al. 2010; Rao et al. 2017). These factors affected plant C isotope discrimination because they could control the intensity of CO_2 exchange between the plant and the atmosphere (Farquhar et al. 1989).

In conditions where the moisture in the air or soil is reduced or precipitation is decreased, the plants will close the stomata to decrease water evaporation, which can lower the stomatal conduction coefficient, followed by a decrease in the partial pressure of CO_2 in the leaf intercellular spaces (P_i) (Farquhar et al. 1989) which leads to a reduction in C isotope fractionation and an increase in $\delta^{13}C$ values of plants and SOM (Xu et al. 2015; Rao et al. 2017). Such a relationship has been found globally (Diefendorf 2010; Kohn 2010; Rao et al. 2017) and regionally (Chen et al. 2015; Lee et al. 2005; Zhang et al. 2020; Golubtsov et al. 2021; Golubtsov et al. 2022). There is a negative relationship between precipitation and the plant/SOM $\delta^{13}C$ values in which increased precipitation can result in a lighter carbon isotopic composition, especially in areas with water stress (Kohn et al. 2010; Diefendorf et al. 2010; Rao et al. 2017).

Previous studies have also reported that plant C isotope discrimination may be affected by temperature (Wang et al. 2013; Xu et al. 2015; Rao et al. 2017; Zhao et al. 2017; Zhang et al. 2020). The activity of enzymes in plants that assimilate CO_2 during photosynthesis is influenced by temperature. The activities increase as the temperature rises, then the photosynthetic rate increases, and the assimilation rate of CO_2 accelerates, and P_i is consequently lowered, which results in a decrease in the carbon isotope fractionation and an increase in the $\delta^{13}C$ values of plants (Farquhar et al. 1989).

However, when speaking about living organisms, it is necessary to consider the limitations beyond which C3 plants begin to experience physiological stress. An increase in annual temperatures is closely related to an increase in evaporation and can lead to the plant growing under water-stress conditions, the effect of which, in terms of $\delta^{13}C$ values, is described above. At the same time, plant growth at low temperatures will favor various physiological adaptations (for example, leaf morphology). Low temperatures may increase water viscosity, inhibit sapwood water movement, and thereby decrease plant water potential, resulting in partial stomatal closure and an increase in $\delta^{13}C$ values in plants and producing SOM as a consequence (Cernusak et al. 2013; Xu et al. 2015).

Thus, previous studies indicated a somewhat contradictory effect of temperature on the $^{13}C/^{12}C$ ratios in plant tissues and the SOM, which various factors can explain. The increase in SOM $\delta^{13}C$ values coinciding with the rise in annual temperatures was correlated with the relative abundance of C4 plants (Jia et al. 2016; Rao et al. 2017). It is well-known that C4 plants have a competitive growth advantage over C3 plants in high temperatures and aridity (Farquhar et al. 1989; Ehleringer 2005; Sage 2005). At the same time, no relationships between $\delta^{13}C$ values of SOM and temperature in C3-dominated plant communities or under pure C3 vegetation were found (Feng et al. 2008; Lee et al. 2005; Jia et al. 2016). This observation was supported by findings indicating an insignificant (in terms of $\delta^{13}C$ values) response of C3 plants to changes in mean annual temperatures compared to mixed C3/C4 phytocenoses (Rao et al. 2017).

Indeed, a linear regression analysis of the $\delta^{13}C$ SOM values driving factors showed a non-significant correlation for the average annual temperatures in the study area. Therefore, it would be most reasonable to search for the effect of temperature on the isotopic composition of soils and plants under ecologically unfavorable conditions. In our case, these included steppes with their relatively high temperatures and low moisture, and mountain tundra and subalpine grasslands that formed under conditions of low air temperatures (Fig. 2). As a result, SOM was most enriched in ^{13}C (Fig. 4).

At the same time, the optimal temperatures in combination with wet conditions in the taiga landscapes of the study area (Fig. 2) provided pronounced discrimination of ^{13}C in plant tissues during CO_2 fixation and caused the accumulation of ^{12}C in the SOM of taiga soils (Fig. 4). Low $\delta^{13}C$ values, both according to our data and according to (Andreeva et al. 2013) were found in the SOM on the northern slope of the Khamar-Daban ridge (Fig. 1, region 6) which is associated with favorable climatic conditions (Fig. 2), namely the combination of the highest precipitation and relatively high (for a given landscape belt and absolute heights) air temperatures.

Thus, the observed $\delta^{13}C$ variations reflected the impact of climatic factors on ^{13}C discrimination during the photosynthesis of C3 plants. Thus, the surface-soil $\delta^{13}C$ signatures in the study area were primarily dependent on the MAP ($r = 0.67$, $p < 0.0001$), and the dependency became a little stronger when only the vegetation season precipitation was considered ($r = 0.69$, $p < 0.0001$) (Fig. 7).

The pronounced response of regional soils to moisture conditions from the standpoint of the composition of stable carbon isotopes confirmed the point of view of previous authors (Koposov 1983; Kuzmin 1988) that one of the main predictors of soil and vegetation cover differentiation in the study area was the heterogeneous distribution of atmospheric precipitation. In addition, this response

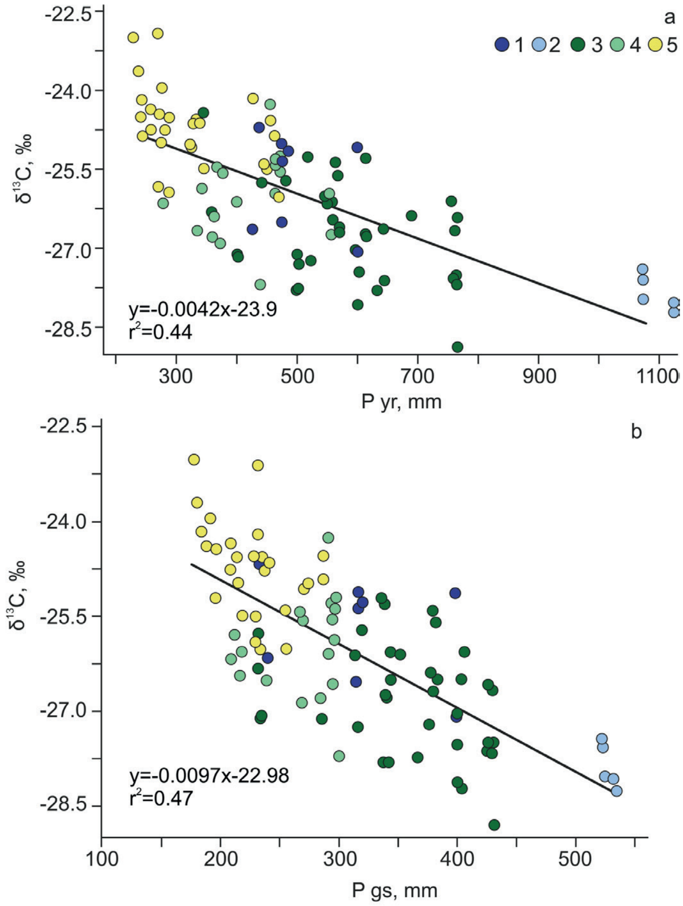

Fig. 7. Relationship between δ^{13}C values of surface soils and mean annual precipitation (a) and mean precipitation of growing season (b). Landscapes: yellow – steppes; dark green – mountain taiga; light green – subtaiga; light blue – subalpine grasslands; dark blue – mountain tundra

agreed with global models (Kohn 2010), indicating that regions characterized by annual precipitation up to 800 mm (most of the soils we studied belong to this range) showed the greatest response to precipitation changes.

Linear regression analysis showed that MAP correlated with the soil δ^{13}C values in the Baikal region (Fig. 7, $p<0.05$) and the sensitivity of the δ^{13}C values response to MAP was –0.42 ‰/100 mm and to vegetation season precipitation was –0.97‰/100 mm. Such values indicated that the soils studied had a high sensitivity to changes in moisture conditions.

This relationship was comparable well with previous quantitative studies, which showed that the coefficient between soil δ^{13}C values and precipitation could be –1.16‰/100 mm in Mongolia (Feng et al. 2008) and varied from –0.3 to –0.8‰/100 mm in different regions of China (Zhang et al. 2020). In semiarid conditions in the northeastern part of China bordering the Transbaikal area, it was more pronounced in some places and amounts to –1.9‰/100 mm (Chen et al. 2015).

CONCLUSIONS

Variability of δ^{13}C in plants and surface soils was investigated in the Baikal region, which is strongly heterogeneous regarding physiographic conditions. The soil δ^{13}C varied by vegetation type and showed a high spatial diversity. At regional scale, the δ^{13}C values of modern plants and surface soils mainly responded to precipitation, with more negative values corresponding to greater precipitation.

Mean annual and mean growing season temperature were not significant factors to explain the spatial distribution of SOM δ^{13}C values at a regional scale. However, temperature was an essential plant growth-limiting factor that negatively affected soil microbial activity and C isotope discrimination in the unfavorable climatic conditions of mountain tundra (coldest) and steppe (warmest) landscapes. There was a powerful synergistic effect of temperature and precipitation in steppe landscapes where high temperature and water stress co-occurrences were found.

In fact, in some areas of the Baikal region, including mountain tundra and steppes, where extreme environmental conditions alternate throughout the year, biological activity was not especially favorable to SOM transformation. Under such conditions, the relative importance of abiotic constraints such as temperature and precipitation was critical for SOM formation.

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APPENDICES

Table A. General information of the sampling sites and stable carbon isotopic composition (δ¹³C) of surface soil horizons

№	Section	Area	Coordinates	Lanscape	Soil type	Elevation, m	MAT, °C	Tgs, °C	MAP, mm	Pgs, mm	TOC, %	$\delta^{13}C$, ‰
1	Nizhniy Bulay-2-20	1	N 52°51'34.73" E 103°06'47.31"	subtaiga	Luvic Phaeozem	492	0.2	17.6	481	298	6.74	-25.52
2	Nizhniy Bulay-2-18	1	N 52°51'34.0" E 103°06'46.7"	subtaiga	Luvic Phaeozem	482	0.2	17.6	481	298	2.19	-25.25
3	Buret'	1	N 52°58'25.82" E 103°28'15.32"	subtaiga	Luvic Phaeozem	403	0.2	17.9	465	298	3.39	-24.15
4	Berezovyi	1	N 52°51'48.44" E 103°21'28.14"	subtaiga	Luvic Phaeozem	417	0.0	17.9	470	298	2.77	-25.85
5	Mikhailovka	1	N 52°59'25.9" E 103°18'01.8"	steppe	Luvic Chernozem	532	0.1	17.8	445	253	4.17	-25.45
6	Fedyaevskyi	1	N 53°15'41.7" E 103°21'54.1"	steppe	Calcic Chernozem	479	0.3	17.8	446	235	4.90	-25.50
7	Osinovyi-20	1	N 52°53'07.55" E 103°19'12.15"	subtaiga	Luvic Phaeozem	473	-0.4	17.8	474	298	2.89	-25.43
8	Osinovyi_18	1	N 52°53'05.6" E 103°19'12.2"	subtaiga	Cambic Calcisol (Loamic)	462	-0.4	17.8	474	298	2.36	-25.31
9	Taiturka-1	1	N 52°52'29.97" E 103°28'12.55"	steppe	Luvic Chernozem	429	0.4	18.0	467	290	4.11	-24.51
10	Taiturka-2	1	N 52°52'57.01" E 103°25'01.88"	steppe	Luvic Chernozem	436	0.4	18.0	470	290	5.30	-24.82
11	Belaya	1	N 52°50'06.8" E 103°20'31.8"	steppe	Pantofluvic Fluvisol (Arenic)	404	0.0	17.7	476	254	3.38	-25.98
12	Lastochkino Gnezdo-2	1	N 52°48'18.9" E 104°47'12.2"	steppe	Calcic Chernozem	514	0.1	17.2	418	236	5.56	-24.19
13	Onot-1	2	N 52°37'45.75" E 102°03'01.75"	mountain taiga	Folic Spodic Cambisol	1232	-2.6	14.5	769	403	13.78	-26.06
14	Onot-2	2	N 52°42'24.35" E 101°55'55.92"	mountain taiga	Sceletic Cambisol	1001	-1.8	14.6	771	402	7.67	-26.31
15	Yulinsk-1	2	N 52°41'44.0" E 102°21'22.7"	mountain taiga	Mollic Phaeozem	511	-1.0	16.3	608	340	3.66	-26.79
16	Yulinsk-3	2	N 52°42'34.5" E 102°23'01.3"	mountain taiga	Mollic Leptosol	591	-1.2	16.3	608	340	10.50	-26.81
17	Yulinsk-4	2	N 52°42'17.2" E 102°22'53.5"	mountain taiga	Leptic Luvisol	687	-1.3	16.3	608	340	2.72	-25.23
18	K1/20	2	N52°19'58.30" E102°51'17.98"	mountain taiga	Folic Cambisol	740	-0.7	15.4	692	375	10.86	-26.34
19	K3/20	2	N 52°28'14.34" E 103°06'50.92"	mountain taiga	Leptic Luvisol	722	-0.6	16.7	556	344	5.37	-26.43
20	Kitoy-2019	2	N 52°28'14.8" E 103°06'43.3"	mountain taiga	Leptic Luvisol	712	-0.6	16.7	556	344	5.81	-26.01
21	Mezhdurech'e	2	N 52°51'28.0" E 102°28'52.1"	mountain taiga	Leptic Luvisol	536	-1.2	17.1	509	339	3.63	-25.37
22	Novostroyka-1	2	N 52°57'33.0" E 101°47'38.3"	mountain taiga	Phaeozem	614	-1.8	15.8	632	425	6.59	-27.60
23	Novostroyka-2	2	N 52°57'30.55" E 101°47'36.90"	mountain taiga	Phaeozem	609	-1.8	15.8	632	425	3.25	-26.49
24	Bol'shaya Belaya	2	N 52°54'51.7" E 102°31'46.3"	mountain taiga	Cambic Phaeozem over Fluvisol	464	-1.0	17.1	509	364	11.13	-27.17

25	Irkut-2	2	N 52°08′48.3" E 103°53′47.1"	mountain taiga	Fluvisol	461	0.5	16.4	616	343	4.75	-27.76
26	2K/21		N 52°11′02.50" E 102°41′29.62"	mountain taiga	Folic Leptic Cambisol	841	-1.5	14.7	771	429	10.66	-27.47
27	4K/21		N 52°10′00.01" E 102°42′29.50"	mountain taiga	Folic Fluvisol (Arenic)	728	-1.1	14.7	771	429	3.23	-26.52
28	5K/21		N 52°09′53.90" E 102°41′58.70"	mountain taiga	Folic Cambisol	738	-1.1	14.7	771	429	10.34	-27.47
29	6K/21		N 52°10′34.21" E 102°41′28.30"	mountain taiga	Folic Cambisol	877	-1.4	14.7	771	429	8.65	-27.68
30	7K/21		N 52°19′34.32" E 102°50′59.60"	mountain taiga	Folic Fluvisol (Arenic)	629	-0.7	15.2	770	430	15.75	-29.50
31	12K/21		N 52°25′36.91" E 103°12′39.50"	mountain taiga	Folic Phaeozem over Fluvisol	520	-0.1	16.6	560	297	8.26	-26.51
32	C1	3	N 53°07′03.83" E 106°41′22.33"	mountain tundra	Folic Cryosol	1654	-4.6	7.9	429	235	32.08	-24.72
33	C2		N 53°07′09.20" E 106°42′49.00"	mountain taiga	Entic Folic Podzol (Differentic)	1414	-4.1	10.0	429	235	16.53	-25.61
34	C3		N 53°07′09.89" E 106°45′02.47"	mountain taiga	Entic Folic Podzol	1125	-3.2	10.5	400	235	12.29	-27.14
35	C5		N 53°06′17.60" E 106°46′21.08"	mountain taiga	Entic Folic Leptosol	910	-2.8	11.3	400	235	13.52	-27.18
36	C6		N 53°05′51.49" E 106°47′04.19"	subtaiga	Folic Leptosol	810	-2.0	13.1	400	216	12.71	-26.05
37	C8		N 53°05′32.02" E 106°48′02.94"	subtaiga	Calcaric Skeletic Phaeozem	600	-1.4	15.5	280	206	8.12	-26.22
38	C9		N 53°05′19.02" E 106°48′51.51"	steppe	Calcaric Mollic Leptosol	460	-1.0	13.4	290	215	12.69	-24.55
39	Ch1		N 52°58′07.43" E 106°48′37.26"	subtaiga	Folic Phaeozem over Skeletic Phaeozem	915	-1.6	12.5	340	216	28.51	-25.85
40	Ch2		N 53°02′23.82" E 106°40′10.74"	mountain taiga	Folic Phaeozem	1160	-1.8	10.9	364	235	40.49	-26.38
41	Ch3		N 53°01′07.74" E 106°40′44.10"	subtaiga	Hyperskeletic Leptosol	701	-2.8	14.8	360	216	8.50	-26.29
42	Sarma-1		N 53°06′32.0" E 106°48′52.90"	steppe	Sceletic Cambisol Protocalcic	626	-1.9	15.5	260	206	3.56	-24.75
43	Horga		N 53°04′32.80" E 106°47′29.70"	steppe	Cambic Skeletic Leptosol	564	-1.3	15.0	260	206	1.93	-24.34
44	Anga		N 52°47′31.70" E 106°34′10.80"	steppe	Calcaric Cambisol	570	-0.6	15.5	241	190	6.52	-24.42
45	Krestovyi		N 52°40′49.40" E 106°23′55.20"	steppe	Calcic Chernozem (Tonguic)	627	-0.5	13.3	280	190	3.33	-23.97
46	Trekhgolovyi		N 53°18′06.60" E 107°06′49.80"	mountain tundra	Lithic Skeletic Leptosol	1225	-3.7	8.3	407	235	5.50	-26.52
47	293	4	N 51°44′24.8" E 102°35′17.3"	subtaiga	Vitric Anthroumbric Leptic Entic Podzol (Loamic, Aric, Endoeutric)	763	-1.0	14.8	385	265	3.30	-25.47
48	357		N 51°40′54.95" E 102°13′42.11"	mountain taiga	Someriumbric Entic Podzols(Arenic, Albic)	744	-0.6	13.5	500	280	8.30	-27.12

49	Haribyaty	4	N 51°38′56.55" E 102°16′06.10"	mountain taiga	Someriumbric Entic Podzol (Arenic, Albic)	785	-1.4	13.3	500	320	15.60	-27.22
50	278		N 51°42′14.20" E 102°24′44.60"	steppe	Hypocalcic Chernozem (Arenic, Aric)	723	0.9	14.6	330	216	5.00	-25.50
51	Zaktui		N 51°41′19.60" E 102°40′52.39"	mountain taiga	Umbrisols (Siltic, Hyperdystric)	918	0.2	13.5	500	340	6.20	-27.79
52	501/1		N 51°44′22.25" E 102°19′44.40"	subtaiga	Entic Podzol (Arenic, Endoeutric)	764	-0.5	13.4	420	300	27.10	-27.74
53	185/1		N 51°49′24.63" E 102°29′18.85"	subtaiga	Vitric Skeletic Folic Leptic Entic Podzol (Loamic, Endoeutric)	773	0.4	13.8	380	265	8.10	-25.54
54	179		N 51°43'10.1" E 102°35′18.5″	subtaiga	Entic Podzol (Arenic, Endoeutric)	739	-0.4	13.6	385	365	7.00	-26.93
55	110		N 51°56′43.8″ E 102°26′25.29″	mountain tundra	Hypereutric Somerimollic Orthoskeletic Leptosol	2063	-4.7	7.6	600	400	15.80	-24.91
56	113		N 51°56′31.77" E 102°26′25.04"	mountain tundra	Folic Sombric Leptosol (Protospodic)	1932	-4.0	8.6	600	400	9.30	-27.06
57	114		N 51°56′16.69" E 102°26′24.58"	mountain taiga	Skeletic Folic Leptic Entic Podzol	1712	-4.0	10.1	600	400	14.50	-27.02
58	114/1		N 51°56′06.27" E 102°26′15.86"	mountain taiga	Hypereutric Somerimollic Folic Skeletic Leptosol	1493	-4.0	11.1	600	400	39.10	-28.19
59	115/1		N 51°55′42.18" E 102°26′11.41"	mountain taiga	Somerirendzic Folic Skeletic Leptosol	1255	-3.0	12.0	600	400	48.00	-27.49
60	Turan		N 51°38′24.37" E 101°38′44.55"	mountain taiga	Histic Stagnic Eutric Gleysol (Arenic)	882	-1.5	13.1	500	360	29.36	-27.75
61	Mondy-1	5	N 51°41′10.42" E 100°55′45.10"	steppe	Cambic Leptic Calcisol	1399	-1.0	10.2	280	195	7.35	-25.14
62	Mondy-2		N 51°39′59.81" E 100°57′01.70"	subtaiga	Cambic Leptic Calcisol	1373	0.0	10.8	360	280	27.20	-26.84
63	Mondy-3		N 51°37′21.47" E 100°55′20.89"	mountain tundra	Folic Lithic Leptosol	1987	-3.5	8.1	490	311	47.16	-26.40
64	Mondy-4		N 51°39'27.05" E 100°54′38.97"	mountain taiga	Folic Albic Leptic Podzol	1672	-1.2	10.0	549	310	7.71	-26.17
65	98 p.		N 51°43′59.67" E 101°00′15.27"	mountain tundra	Umbric Leptosol	2315	-3.5	7.1	489	311	6.07	-24.98
66	99 p.		N 51°43′29.57″ E 101°00′12.82″	mountain tundra	Umbric Leptosol	2156	-2.7	7.7	489	311	6.65	-25.39
67	100 p.		N 51°43′24.6" E 101°0′4.32"	mountain taiga	Cambic Leptosol	2078	-3.1	7.5	585	380	5.44	-26.46
68	101 p.		N 51°43′7.25" E 100°59′51.72"	mountain taiga	Umbric Leptosol	1932	-2.4	8.3	585	380	25.13	-25.59
69	102 p.		N 51°42′23.08" E 101°0′5.26"	mountain taiga	Histic Cryosol	1687	-3.4	9.4	585	380	30.20	-25.42
70	103 p.		N 51°37′21.68" E 100°55′20.03"	mountain tundra	Histic Eutric Turbic Cryosol	1990	-3.5	8.1	490	311	44.24	-25.35
71	104 p.		N 51°37′48.39" E 100°53′19.90"	mountain taiga	Histic Cryosol	1874	-2.8	8.5	490	311	42.15	-25.68

72	41 p.	5	N 51°42'21.60" E 101°00'5.06"	mountain taiga	Histic Cryosol	1996	-3.4	9.4	585	380	29.81	-26.65
73	43 p.		N 51°39'25.78" E 100°54'39.28"	mountain taiga	Eutric Leptosol	1627	-1.2	10.0	550	353	6.29	-26.10
74	506 (M1.1)	6	N 51°23'36.96" E 104°50'35.34"	subalpine grassland	Folic Umbrisol (Hyperdystric)	950	-1.6	14.7	1082	520	3.14	-27.65
75	507 (M1.2)		N 51°24'38.10" E 104°50'43.98"	mountain taiga	Folic Umbrisol (Hyperdystric)	1050	-1.8	14.7	1082	520	4.99	-27.92
76	508 (M1.3)		N 51°23'28.38" E 104°50'45.18"	subalpine grassland	Haplic Umbrisol (Hyperdystric)	950	-1.9	14.7	1082	520	7.59	-27.45
77	509 (M2.4)		N 51°22'52.44" E 104°51'31.56"	subalpine grassland	Cambic Umbrisol (Hyperdystric)	1200	-2.2	14.0	1120	532	11.53	-28.13
78	510 (M 2.5)		N 51°22'45.00" E 104°51'27.48"	subalpine grassland	Haplic Umbrisol (Hyperdystric)	1200	-2.1	14.0	1120	532	9.02	-28.01
79	2-Z-18	7	N 51°31'55.3" E 107°06'14.8"	steppe	Skeletic Cambic Leptic Calcisol Turbic	633	-0.5	17.9	224	177	0.85	-23.01
80	5-Z-18		N 50°38'02.8" E 105°23'09.1"	steppe	Skeletic Cambic Leptic Calcisol Hypercalcic Yermic	787	-0.9	17.3	293	239	5.65	-26.09
81	9-Z-18		N 50°36'21.1" E 105°25'52.3"	steppe	Skeletic Cambic Leptic Calcisol	691	-1.1	17.3	278	234	5.63	-25.90
82	11-Z-18		N 50°43'35.4" E 105°54'15.7"	steppe	Skeletic Cambic Leptic Calcisol Hypercalcic Yermic	861	-0.5	18.0	244	212	4.10	-24.82
83	1-Z-21		N 50°35'25.4" E 105°26'56.4"	subtaiga	Cambic Leptic Calcisol	729	-1.3	17.3	319	239	6.31	-26.58
84	1-Z-21 ovrag		N 50°35'31.70" E 105°26'51.70"	steppe	Cambic Leptic Calcisol	701	-1.2	17.3	293	239	1.56	-24.75
85	2-Z-21		N 50°58'30.6" E 106°05'03.8"	steppe	Luvic Chernozem	774	-0.6	18.1	224	177	5.57	-23.61
86	3-Z-21		N 50°58'45.1" E 106°04'11.1"	mountain taiga	Folic Albic Leptic Podzol	817	-0.6	18.1	325	176	7.62	-24.59
87	4-Z-21		N51°35'39.70" E107°03'17.94"	steppe	Cambic Leptic Calcisol	701	-1.2	17.6	278	198	1.40	-24.32
88	B.Kunaley-1		N 51°25'30.89" E 107°34'30.00"	steppe	Luvic Chernozem	735	-1.7	14.6	315	230	5.15	-24.49
89	B.Kunaley-2		N 51°25'6.61" E 107°36'11.97"	steppe	Luvic Chernozem	724	-1.7	16.9	315	230	1.51	-24.50
90	Khorinsk		N 52°13'41.98" E 109°49'51.16"	steppe	Cambic Leptic Calcisol	715	-1.9	14.0	236	187	0.83	-24.17
91	Pesterevo		N 51°30'41.5" E 107°29'16.7"	steppe	Calcic Chernozem	606	-1.7	14.6	315	230	0.64	-24.60
92	Ust'-Menza-1	8	N 50°13'28.90" E 108°37'33.10"	subtaiga	Fluvisol	734	-1.7	15.7	551	291	2.91	-26.03
93	16-Z-18	9	N 50°50'01.0" E 116°19'36.2"	steppe	Luvic Chernozem	679	0.2	18.2	312	261	4.27	-24.92
94	17-Z-18		N 50°10'36.5" E 116°17'26.7"	steppe	Leptic Chernozem	723	0.1	18.7	312	263	2.90	-24.84
95	19-Z-18		N 50°07'21.8" E 115°58'43.4"	steppe	Leptic Chernozem	650	0.5	19.0	283	238	2.61	-22.98

Note: MAT – mean annual air temperature, Tgs – mean air temperature of vegetation season, MAP – mean annual precipitation, Pgs - mean precipitation of vegetation season, TOC – total organic carbon