
GENESIS AND GEOGRAPHY OF SOILS

Supraglacial Soils and Soil-Like Bodies: Diversity, Genesis, Functioning (Review)

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Abstract—In the 21st century, glaciers are perceived as a distinct biome that has taken on special significance in today's world of retreating ice. In this paper, we review the results of recent studies of organomineral formations on glaciers, their diversity, genesis, functioning, and the role in the biosphere. The question is raised about the possibility of involving supraglacial organomineral formations in the range of objects of soil science. We review the supraglacial zone as an area of soils and soil-like bodies, the biogeochemical processes in which affect the glacial biome and the surrounding landscapes. Interpretation of supraglacial organomineral formations from a pedological point of view allows us to identify several typical soil processes: accumulation and stabilization of organic matter (OM), its heterotrophic transformation, formation of dark-colored humified OM, accumulation of residual solid-phase products of functioning in situ, fine earth aggregation, and biochemical weathering. Among supraglacial formations, we distinguish pre-soils and soil-like bodies in ice and snow, metastable soil-like bodies on cryoconite, and soils with microprofiles under moss communities on ice, as well as relatively stable soils with macroprofiles on silicate gravelly to fine-earth deposits underlain by moving glacier and dead glacier ice. Labile dissolved OM accumulated and transformed in supraglacial soils and soil-like bodies has a significant impact on the periglacial zone, leading to the reservoir and priming effects. The studies of supraglacial organomineral systems are of fundamental importance for understanding the evolution of ecosystems on Earth, as well as for modeling supraglacial formations of extraterrestrial bodies with a vast cryosphere. Supraglacial soil formation is also a model object for studying common soils under conditions of a continuous external input of organic and mineral components, the contribution of which beyond the glaciers is no less significant, but is masked by the polymineral substrate of soils and parent rocks themselves.

Keywords: organomineral interactions, pre-soils, cryoconites, periglacial zone, greenhouse gas emissions, radiocarbon

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INTRODUCTION

The history of glaciers on Earth extends back to at least 2.9 billion years. This is evidenced by the oldest fossil moraines—tillites of the Kaapvaal Craton in South Africa [26, 249]. Glaciers and accumulations of mineral material on their surface probably appeared almost synchronously, since loose products of rock weathering are easily transported by wind or gravitational processes and accumulate on the ice surface. Microorganisms began to colonize terrigenous rocks at least 3.2–2.6 billion years ago [50, 52, 103, 241]. Therefore, from the very beginning, on the surface of glaciers, in the so-called *supraglacial zone*, along with mineral components, biogenic components could also accumulate. In this regard, it can be assumed that supraglacial organomineral systems can be very

ancient formations. Already at the first stages of the existence of glaciers, such systems could influence energy balance of glaciers by changing the albedo and regulate biogeochemical processes in the supraglacial zone and beyond. The importance of supraglacial organomineral systems increased significantly during large-scale cooling periods both in the Paleoproterozoic (Huronian glaciation, ~2.4–2.2 billion years ago) and Neoproterozoic (Cryogenian period, ~0.72–0.64 billion years ago) eras, when glaciers covered all continents and reached sea level in tropical latitudes [79, 186, 233]. The debate about the extent of Precambrian glaciations continues [102, 113, 124], but recent paleoclimate models show that even under conditions similar to Snowball Earth, there was never a shortage of ice-free areas on the continents [50, 101, 139], and vol-

canic and continental dust accumulated on the ice surface could significantly reduce the albedo [32]. Such ice-free spaces were probably similar to the Dry Valleys of modern Antarctica and served as sources of fine earth. There is evidence of this in the sedimentary record of the Cryogenian in the form of loessites—eolian deposits of periglacial environments that are preserved in a consolidated state [78].

Paleontological data indicate that several groups of photoautotrophic eukaryotes, for example, red and green algae [57, 88, 125, 126], which appeared before the Cryogenian (i.e., before 0.72 billion years ago), survived planetary glaciations. Along with hydrothermal landscapes, lakes, and soils of ice-free “oases” (like Antarctic oases), glacial surfaces covered with organomineral sediments could also act as refugia during large-scale glaciations (see references in the review [101]). The supraglacial zone appears to have been the largest refugium for Precambrian organisms, including protists [62]. Organisms had to adapt to life in the cryosphere, freeze–thaw cycles, high ultraviolet radiation, and extremely limited primary production of organic matter [232]. The authors [253] substantiate the role of cold glacial conditions in the evolution of terrestrial plants and their further expansion on land, as well as the possibility of a return from multicellular forms to unicellular ones in some algae due to adaptation to conditions on glaciers. Glaciers covered with gravelly fine earth clastic materials could act as refugia for vascular plants during the Quaternary glaciations [83]. Some researchers consider supraglacial sediments as model objects for understanding the conditions, under which microorganisms could exist on the ancient Earth, as well as potential analogues of supraglacial formations on extraterrestrial bodies, including Mars and Saturn’s moon Titan [106, 196, 231].

Thus, paleontological and paleoclimatic data obtained in recent years indicate the importance of organomineral sediments on glaciers for the evolution of both individual organisms and ecosystems on Earth. Now we have become contemporaries of the dramatic melting of glaciers, enhanced, among other things, by the dark-colored organomineral material on their surface, which makes more relevant than ever the study of the genesis of such material, its diversity and structure, the development of soil processes in it, as well as its influence on soils and soil covers that evolve after the retreat of glaciers.

Glaciated area in 18 mountain regions of continental Russia reaches 3480 km²; most glaciers are rapidly retreating. Over the 20th and early 21st centuries, their area has reduced by dozens of percent [123]. In particular, from 1960 to 2014, in the Caucasus, 480 km² were freed from ice [230]; from 1997 to 2017, Elbrus glaciers lost 11% of their area (14 km²) [129]; in Altai, since the 1950s to 2008, glaciers shrank by 27% [21]. The glaciers of the Russian Arctic cover a total of 51 592 km², and their annual total mass loss in 2002–2016 was about

11 Gt per year [245]. Such retreat of glaciers has long attracted soil scientists to study the patterns and rates of soil formation on glacial material of various ages [9].

According to our estimates, over the past two decades, international teams have published more than 300 works on supraglacial formations, including cryoconites (from the Greek *kryos*, cold and *konis*, dust)—dark-colored, finely dispersed and well-aggregated organomineral sediments on glaciers, which were described by A. Nordenskiöld more than 150 years ago. This includes several large reviews [37, 64, 97, 98, 187].

Pioneering work of M.A. Glazovskaya [6, 7] on eolian sediments on Tien Shan glaciers was carried out as part of expeditions of the Institute of Geography of the USSR Academy of Sciences and the Kazakhstan Branch of the USSR Academy of Sciences decades before the boom in supraglacial research. Even then, the possibility of biogenic transformation of eolian glacial fine earth in situ, its enrichment in organic matter due to the development of cyanobacteria and algae, as well as the formation of stable granular aggregates, was revealed. Glazovskaya pointed to the redeposition of such fine earth in the near periglacial zone and did not exclude the possibility of the origin of loess-like loams, as well as covering (mantle) loams of the Russian Plain, from eolian sediments on ice sheets.

Given the still significant area of glaciation in Russia, we have a poor idea of the diversity and area of supraglacial organomineral sediments. There are no systematic studies of these formations; until recently, the term cryoconite was rarely used in Russian-language scientific literature.

Currently, data on algal communities and cryoconites have been published on Akkem Glacier in Altai [225] and Glacier 31 in Suntar-Khayata [224, 226]; data on the micromorphology, geochemistry, and biochemical properties of supraglacial fine earth have appeared [2, 29, 31, 127, 128, 131] and antibiotic-resistant bacteria on the glaciers of the Caucasus [147] have been studied. Particle-size distribution data have been obtained for cryoconite from Bolshevik Island of the Severnaya Zemlya Archipelago [28]; radioecological and geochemical properties of cryoconite from the Novaya Zemlya Archipelago [153] and greenhouse gas fluxes from the Caucasus and Altai glaciers covered with cryoconite and other supraglacial materials [14] have been studied.

Supraglacial systems of Antarctic and low-latitude glaciers have also begun to be actively studied. It has been found that, in contrast to the mountainous and arctic regions, the supraglacial organomineral material in Antarctica does not thaw every year, and the age of organic matter, the composition of microbial communities, primary production, respiration, and biomass growth in cryoconite differ significantly [3, 44, 45, 55, 68, 84, 85, 145, 173, 191, 192, 205, 232, 243]. Supraglacial systems in the areas of Russian Antarctic stations have yet to be studied.

In recent years, some researchers [3, 24, 30, 151, 175] began to consider supraglacial organomineral systems as soil-like stratified formations, and also as a source of material enriched with nutrients that comes from the glacier into the periglacial zone and plays the role of a local catalyst of initial soil formation on oligotrophic substrates freed from ice [13, 27, 162].

There are no global estimates of the areas occupied by soil-like systems in snow and ice, but given that glaciers cover about 10% of the land surface [37], and permanent and seasonal snowfields occupy up to 35% [154], these areas are very significant, even now, during the period of shrinking cryosphere. Supraglacial soil-like bodies can be considered from the standpoint of the theory of extreme pedogenesis [11, 12], since they are formed at low temperatures with a long dark period in winter (most of the ice is concentrated in the polar regions) and contrasting, relatively warm conditions with intense solar radiation in summer. The communities inhabiting them go through daily, seasonal, and annual freeze–thaw cycles; experience alternating aerobic and anaerobic conditions, high levels of ultraviolet radiation; exist in conditions of an unstable substrate, a limited pool of nutrients, periodically high water content and can be buried under snow cover even during melting periods.

It is believed that where the glacier ends, the soil begins. New interdisciplinary studies of supraglacial systems have allowed soil scientists to realize that where the glacier begins, soils do not end (or rather, they do not necessarily end). Certainly, there is no well-developed soil cover on the surface of glaciers. However, it is known that this is an area where some processes characteristic of soils are realized and objects are formed that can be classified as at least soil-like bodies. So far, the concept of supraglacial soil formation has not been formulated: the diversity of organomineral systems on glaciers has not been described from the standpoint of soil science, there is no understanding of which objects on glaciers should be classified specifically as soils or soil-like bodies, and which should not, what distinguishes glacial soils from soils outside glaciers, what are their morphotypes, formation processes, absolute age and residence time, and what are their evolutionary trajectories and biosphere functions.

The purpose of this review is to summarize modern ideas about the origin, diversity, and transformation of organomineral bodies on the surface of glaciers; to identify potential objects of soil science among them; and also to describe the connections between supraglacial processes and the initial stages of soil formation in near-glacial environments.

CONDITIONS FOR SOIL FORMATION ON A GLACIER

Such conditions arise with the appearance of solid-phase impurities on the surface of a glacier and the for-

mation of heterogeneous systems of ice, snow, firn, and meltwater with organic, mineral, and organomineral components. The components interact with one another, and, due to the specifics of the environment, a significant part of them becomes water-soluble. According to the figurative expression of N. Takeuchi, a glacier has both a light (ice–snow–firn) and a dark “side”. The complexity of processes on the dark “side” of a glacier, i.e., in supraglacial dark-colored impurities, appears to be comparable to that in other terrestrial systems, such as soils. The surface of a glacier is by no means a clean zone; just like soil, it is a medium for the development of organisms, a filter for numerous introduced compounds, and a critical zone at the junction of the hydrosphere, atmosphere, lithosphere, and pedosphere.

In glaciology, the most important function of glacial solid-phase impurities is considered to be their influence on the reflectivity of the surface—albedo. This effect was noted by Nordenskiöld [163, 164] and Nansen [161]. Impurities enhance direct positive near-surface radiative forcing, which accelerates melting [158, 159, 244]; they are called light-absorbing impurities; this term is also translated into Russian as light-absorbing particles [1]. The first option seems more accurate, since such “particles” can include large fragments and accumulations of organic and mineral material, as well as individual organisms. Compared to the mass and volume of ice and snow, these components are just impurities.

Light-absorbing impurities change the reflective properties of snow or ice in different ways [73, 250]. The spectral signatures of mineral dust [74, 75], algae [122, 214], and supraglacial clastic material [59] differ not only among themselves but also within each group depending on the mineralogical and chemical composition of the components. The most common light-absorbing impurities on glaciers are algae and cyanobacteria, the products of their heterotrophic transformation [223], as well as cryoconites [73]. Black carbon particles [53, 92, 142] and, to a lesser extent, mineral dust [74, 166] are particularly efficient in changing the albedo. Mineral dust is formed either from rock materials or from soils and is introduced from local and distant landscapes, mainly from deserts [132, 204].

In this review, we briefly consider the inorganic matter on the surface of the glacier, but we focus on the organic and organomineral components, which are the basis for the biological cycle, biogeochemical transformation, and soil-forming processes in the supraglacial zone.

Inorganic matter. Studies by Glazovskaya [6, 7] on the flat-topped watershed glaciers of the Tien Shan, where there is no supply of gravitational masses from the adjacent mountain slopes, showed that the main material accumulating on the glacier by the eolian transport and fallout from the atmosphere is coarse and fine dust. The composition of the dust is domi-

nated by quartz and feldspars; clay minerals are present in significant quantities, especially hydromicas (illite, vermiculite) with thin and wide light-weight grains, which contributes to their transfer by wind.

Modern studies carried out at the global level [187] have shown that the deposits of mineral particles on ice generally correspond to the composition of surface deposits of the lithosphere. The ratio of local and long-distance mineral particles varies greatly and depends on the absolute height and position of the glacier (cover, watershed, mountain-valley). Typically, the lower the glacier, the higher the proportion of local material. On low-lying glaciers, the mineral material can have a sandy composition or be polydisperse with a predominance of fine sand and coarse silt fractions [28]. Geological conditions or geographic isolation of the glacier determine the specific geochemistry of the supraglacial material. For example, in cryoconite from Kersten Glacier on Kilimanjaro Stratovolcano, the high content of rare earth elements is associated with the abundance of volcanic glass [69]. Locally, cryoconite can be enriched in calcium or iron, if there are nearby carbonate rocks or exposure to iron ore deposits. The ice cover surrounding ice-free land areas in Antarctica receives eolian material mainly from local sources [72].

Analysis of clay minerals makes it possible to distinguish local material from distant material [131, 170, 228]. This is especially true for indicator minerals. For example, palygorskite indicates formation under arid conditions. Certainly, in comparison with the pioneering research by M.A. Glazovskaya, the set of clay minerals on the glaciers of the world has been expanded significantly: smectites, mixed-layer minerals, and kaolinite have been identified.

Just as in soil, the environmentally significant part of the substance in cryoconite consists of heavy metals and radionuclides [31, 41, 42, 111, 143, 144, 160]. An important additional indicator of the source of mineral dust is its isotopic composition; for example, the strontium isotope ratio indicates the origin of material that was brought on Alpine glaciers [228].

Mineral particles arriving on the glacier surface are often pretreated by the processes of physical and chemical weathering, which increases the specific surface area for interaction with living matter and contributes to further biochemical weathering already in the supraglacial zone [255].

Organic matter (OM). Glaciers contain about 6 Pg (petagrams, 10^{15} g) of organic C, most of which is in the dissolved form and about a quarter in solid-phase particles [104]. Up to 96–98% of the global pool of glacier organic C is stored in Antarctic and Greenland ice sheets; 2–4% is stored in mountain glaciers. On a global scale, the reservoir of organic C in glaciers is significant, although it is much inferior to such large reservoirs of the cryosphere as the soils of permafrost areas (~1600 Pg). However, during ablation, the gla-

acier releases large amounts of readily available water-soluble OM, which quickly enters the surrounding soils. This labile reservoir of organic C is still poorly taken into account in the carbon balance of periglacial ecosystems [104].

Supraglacial OM. On the surface of glaciers, OM [36, 91, 104, 141] accumulates due to primary production *in situ*, comes from external sources during long-distance aerial transport, erosion of surrounding soils and sediments, and is also released from the glacier body during ablation [199, 212, 215]. In terms of the variety of forms and complexity of the chemical structure, the OM of the supraglacial zone is not inferior to soil OM.

Various supraglacial organomineral formations may contain from a few to tens of percent of organic C [187], and its distribution has certain spatial patterns within the same glacier [137, 162] and between different glaciers [91]. On the surface of glaciers, a dynamic pool of organic compounds of *in situ* microbial origin and those added with transported plant and soil materials, combustion products of fossil fuels and biomass, marine and microbial aerosols, and many other allochthonous C-containing components, including, for example, silty material of desert areas [130], pesticides [81] and microplastics [34].

Live OM. Glaciers are a separate biome within the cryosphere [37]. Organisms either live in the supraglacial zone or use it as a transit environment, all together causing the effect of a biological albedo reduction [108], which was previously neglected against the background of the great influence of abiotic factors. The living components of the supraglacial system are cyanobacteria and eukaryotic microalgae, mosses, complex heterotrophic microbial communities, invertebrates, birds, and even large mammals.

Prokaryotic and eukaryotic photoautotrophs form the basis of the supraglacial ecosystem. Among them, cyanobacteria, green algae, and diatoms predominate [119, 211]. Green algae in snow are a significant C sink. For example, in Antarctica, they accumulate in quantities from 5 to 5800 g dry phytomass m^{-2} [90], where the upper limit is comparable with the stocks of organic C in soil. In [187], lists of the most common taxa of photoautotrophs are given: among cyanobacteria, these are Pseudanabaenaceae (*Leptolyngbya*, *Pseudanabaena*) and Phormidiaceae (*Phormidium*, *Phormidismis*, *Wilmottia*, *Microcoleus*) families; among green algae, Mesotaeniaceae (*Cylindrocystis*, *Ancylonema*) and Ulotrichaceae (*Klebsormidium*) families; and among diatoms, Naviculaceae (*Pinnularia*, *Navicula*) family. The genus of green algae *Chlamydomonas* of the Chlamydomonadales family is the main, but not the only cause of such a common phenomenon as pink, red or “watermelon” snow. For example, the genus *Sanguina* of the same family also causes the effect of red and orange snow [179].

As in any terrestrial ecosystem, autotrophic organisms exist on glaciers in association with *heterotrophic communities* of bacteria, archaea, and fungi, as well as with protists and viruses [36–38, 119, 146]. The photic zone of glaciers contains 10^{21} – 10^{26} cells of bacteria and archaea [114, 115], and the fungal biomass reaches 10^4 – 10^5 CFU per 100 mL [70, 71, 108]. Bacteria predominate among cryoconite heterotrophs [58, 76, 172]. Mycobiota of cryoconites has received less attention so far [77, 119, 149, 200], although it is known that cryoconites are usually dominated by yeasts rather than filamentous fungi [169, 200]. The mycobiota is dominated by *Cryptococcus gilvoscens*, *Mrakia* spp., *Rhodotorula* spp., *Phialophora alba*, and *Articulospora tetracladia*. Most of their strains are psychrophilic and produce hydrolases that are active at near-zero temperatures [200]. This fact proves that mycobiota can actively participate in the decomposition and transformation of cryoconite OM.

Various pigments, such as the carotenoid astaxanthin [182] and the phenol purpurogallin in algae [183] and melanins in fungi [169] additionally absorb energy, regulating the ambient temperature and releasing meltwater and nutrients dissolved in it.

Recently, evidence has emerged [147, 148] that glaciers are a habitat for modern bacteria and a repository for ancient antibiotic-resistant bacteria. Apparently, the supraglacial zone, like the soil and aquatic environment [18], acts as a reservoir of antibiotic resistance genes that can be transferred to clinically important bacteria through horizontal transfer—the main cause of multidrug resistance.

Viruses (their classification as living OM here is conditional) on glaciers regulate the number of bacteria and indirectly affect the microbial biomass and the amount of OM in general [38, 49, 206].

Plants and plant material. The most famous inhabitants of glaciers are mosses of the genera *Bryum*, *Ceratodon*, *Drepanocladus*, *Racomitrium*, *Schistidium*, and *Hygrohypnella*, which exist in the form of so-called “glacial mice”—mobile spherical (ball-like) colonies and cushions of moss [4, 48, 66, 96, 107, 177]. Such moss cushions have been described on glaciers in South America, Iceland, Svalbard, and especially widely in Alaska, where hundreds of moss cushions can simultaneously be present on a glacier [107]. In Central Africa, on the tongues of retreating high-mountain glaciers, aggregates of protonema (pre-juveniles) and gemmae (brood bodies) are common, i.e., initial phases of the development of bryophytes, for example *Ceratodon purpureus* [234].

Depending on the location of the glacier, its supraglacial zone receives a variety of fresh plant material (Fig. 1) rich in carbon and nitrogen in the form of leaves, needles, herbs, as well as numerous pollen grains [54], which can be transported through the air over long distances.

Owing to the abundance of the aqueous phase, cryoconites contain many *Protozoa*, primarily ciliates [152].

The supraglacial zone is a home to numerous *invertebrates*, from rotifers to ice worms [109, 198, 257], and regular atmospheric fallout of arthropods occurs [95]. The heterotrophic block of cryoconites is represented by such invertebrates as tardigrades (Tardigrada), rotifers (Rotifera), springtails (Collembola), and freshwater crustaceans (Branchiopoda and Maxillopoda) [254]. Ice worms (*Mesenchytraeus solifugus*) influence the algal biomass: they feed on it [156] and introduce nutrients, which, on the contrary, stimulate an increase in algal biomass [110].

The biomass of *arthropods*, especially insects, is rich in carbon (up to 60%) and nitrogen (up to 12%) represented by labile organic compounds, has a low C : N ratio, and at the peak of arthropods fallout in summer its contribution to the pool of available C and N on the glacier is comparable to the contribution of algae [108].

Many *large animals*—from ungulates to felids, as well as birds—temporarily use glaciers and snowfields as a shelter, for searching for food, nesting, and as a transit zone [108]. The main OM that vertebrates bring in is excrement, wool, feathers, as well as food, including the corpses of other animals. Just as in soils, bioproductivity increases in ice and snow near colonies of birds and seals; abundant algal blooms are observed. This phenomenon is especially pronounced in the Antarctic [90, 184]. The ornithogenic factor influences the formation of soil-like bodies in places of cryoconite accumulation [3].

Humans have a local direct impact on the supraglacial zone due to tourist load [86] and scientific research, as well as a large-scale indirect impact through atmospheric fallout of pollutants.

Dead OM from glaciers is very diverse. Along with OM of microbial origin, glaciers contain fragments of compounds that mark the supply of plant and soil material; a significant part of the OM is bioavailable [199]. The composition of OM largely depends on the location and size of the glacier and the periglacial geochemistry. The OM of cryoconites from Antarctic glaciers is predominantly autochthonous, while the composition of OM from cryoconites in the Arctic indicates different combinations of sources [168]. The proximity of forest ecosystems increases the amount of lignin- and tannin-containing fragments, and the location of glaciers along the path of aerial transport from areas with intense anthropogenic load increases the amount of highly condensed aromatic compounds [90].

The supraglacial zone is expected to contain a lot of dissolved organic matter (DOM) [199]. Its concentration in the ice and snow of ice sheets is lower than on the surface of mountain glaciers, since the latter are closer to land-based sources of C and are better warmed up, ensuring high availability of meltwater and increasing the efficiency of primary production [185].

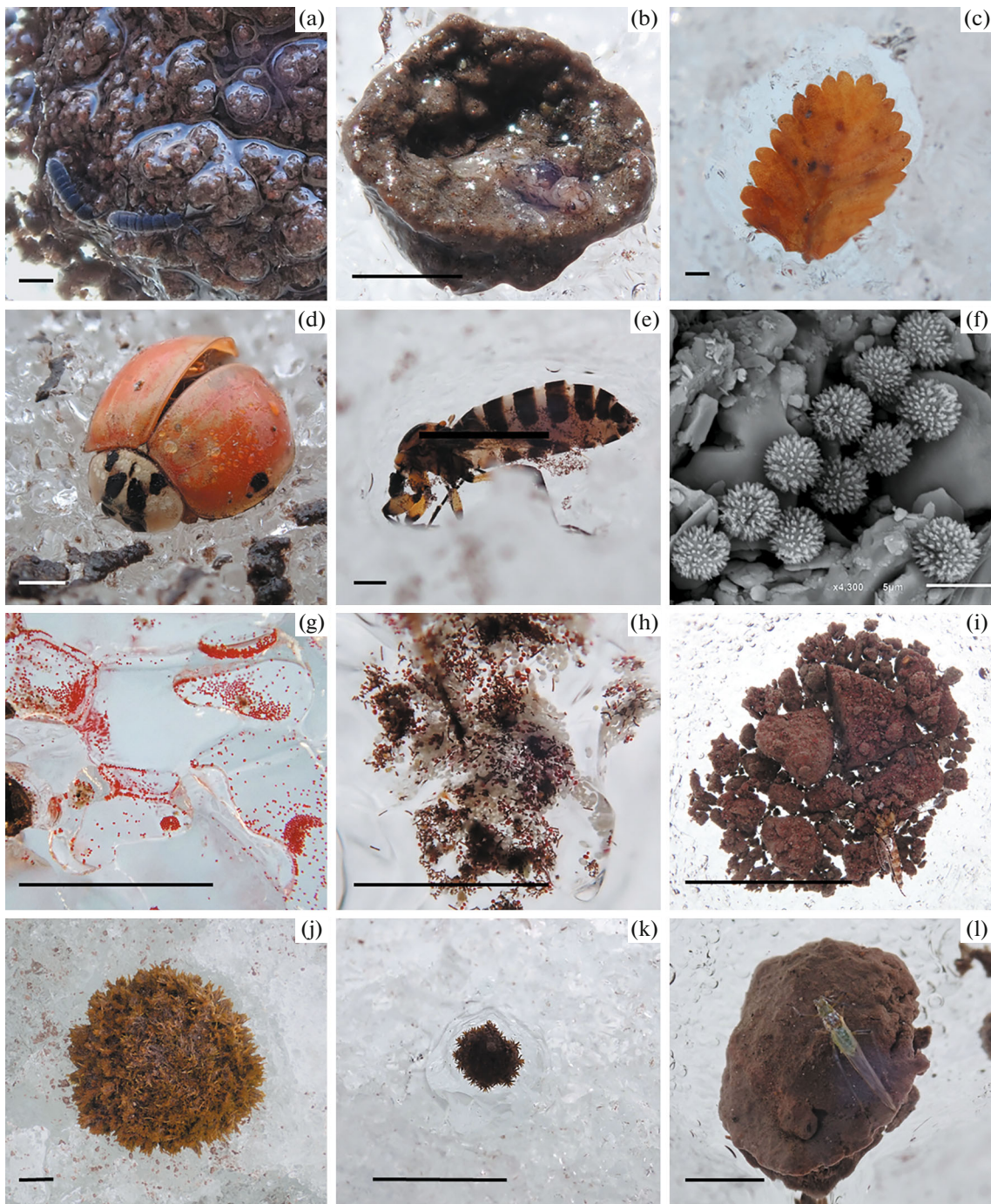


Fig. 1. Fresh OM of the supraglacial zone: (a) “grazing” of springtails on a cryoconite granule (Garabashi Glacier, Caucasus); (b) insect corpse inside a cryoconite granule, photo of a section of the granule (Levyi Aktru Glacier, Altai); (c) *Nothofagus* leaf melted into ice (Perito Moreno Glacier, Argentina); (d) ladybug corpse (Mizhirgi Glacier, Caucasus); (e) insect corpse melted into ice and covered with a biofilm of *Chlamydomonas nivalis* and *Ancylonema nordenskiöldii* algae (Levyi Aktru Glacier, Altai); (f) pollen grains in cryoconite, photo from a scanning electron microscope, secondary electron mode (Levyi Aktru Glacier, Altai); (g) biofilm of *Chlamydomonas nivalis* algae in firn (Mizhirgi Glacier, Caucasus); (h) multicomponent organomineral film composed of *Chlamydomonas nivalis* and *Ancylonema nordenskiöldii* algae and fine mineral particles in the meniscus of the cryoconite hole (Levyi Aktru Glacier, Altai); (i) organomineral sediment at the bottom of cryoconite hole covered with a biofilm of *Chlamydomonas nivalis* algae and insect corpse (Levyi Aktru Glacier, Altai); (j) ball-like moss cushion on an ice pedestal; moss serves as a thermal insulator (Bertil Glacier, Svalbard); (k) moss sprout in a cryoconite hole; moss serves an accelerator of ice melting (Bertil Glacier, Svalbard); (l) insect on a cryoconite granule in a cryoconite hole (Levyi Aktru Glacier, Altai). Scale bar at (a–e, g, h, l), 1 mm; at (i–k), 10 mm; at (f), 5 μ m.

Allochthonous DOM mainly consists of chemically stable fragments of humic-, lignin-, and tannin-like compounds and highly condensed aromatic C [67, 80, 215]. Autochthonous DOM has a predominantly lipid and protein composition due to the development of algae and photosynthetic microorganisms [36, 157]. Heterotrophic bacterial communities are involved in the mineralization and transformation of labile forms of OM with residual accumulation of more stable products, for example, alicyclic compounds with carboxyl groups [36, 98, 157].

According to ^1H NMR spectroscopy, alkaline extracts from cryoconite of Antarctic glaciers are dominated by signals from proteins and peptides of microbial origin, contain phospholipid fatty acids, and lack plant biomarkers except for mosses [168]. In isolated areas of East Antarctica [40], more than 2300 chemical compounds were identified only in the DOM fraction isolated from snow. They were dominated by aliphatic fragments, but condensed aromatic compounds were also present in the structure of dozens of compounds. DOM consisted of lignin-like compounds, proteins, fats, unsaturated hydrocarbons, carbohydrates, and tannin components. Some of the lignin-like compounds were apparently associated with OM of terrestrial origin, and some were associated with alicyclic compounds rich in carboxyl groups. About 90% of the identified compounds (with both aromatic and aliphatic fragments) had high bioavailability for heterotrophs and were transformed by microbial communities in short-term laboratory experiments through parallel processes of degradation and synthesis [40]. Significant amounts of easily available DOM with a high proportion of aliphatic and peptide fragments (for example, released by the glaciers of Svalbard and Greenland), when entering periglacial landscapes, can accelerate the heterotrophic degradation of DOM [120], i.e., lead to a priming effect [133].

On glaciers surrounded by ecosystems dominated by vascular plants and developed soil cover, compounds of microbial origin in combination with material of lichens and mosses, rather than vascular plants, can still predominate in the OM. Low levels of molecular biomarkers of vascular plants on Alaskan glaciers indicate that the contribution of OM sources, such as forests or peatlands, is insignificant [215]. According to ^1H and ^{13}C NMR spectroscopy data [247], the OM of cryoconite from Athabasca Glacier (Canadian Rockies) contained a large amount of fatty acids, *n*-alkanols, *n*-alkanes, wax esters, and sterols; there was no lignin, and the ratio of alkyl/O-alkyl fragments in solid-phase ^{13}C -NMR spectra differed from the composition of OM in the surrounding soils. Analysis of phospholipid fatty acids confirmed the significant contribution of microorganisms to the OM of cryoconite of this glacier, primarily bacteria, as well as microscopic eukaryotes [209]. However, in modern and ancient ice from a core from the Vetrenyi Dome on Graham Bell Island

of the Franz Josef Land Archipelago [89], despite its remote maritime location, thousands of compounds were identified, including humic (more precisely, humic-like) substances (according to Fourier transform ion cyclotron resonance mass spectrometry (FT/ICR MS)). Humic substances constituted a significant part of the isolated organic material and indicated its supply from terrestrial ecosystems.

Pyrogenic OM. The entire spectrum of products of incomplete combustion of modern biomass and fossil OM is present on glaciers and is similar to the composition of pyrogenic OM in soils and sediments, including soot; charcoal; elemental, black and brown C; lithogenic kerogen, and microcrystalline graphite. Pyrogenic OM in ice cores (black C, levoglucosan, etc.) is an important carrier of information about the chronology of fires and fossil fuel combustion [138, 188, 258]. The classification of the continuum of combustion products is highly imperfect; the categories of pyrogenic OM overlap, but all of them are dominated by polycyclic aromatic hydrocarbons, which can be separately isolated from glaciers [30, 140]. Among pyrogenic OM, particles of black C are considered especially reactive [1, 173, 174], even their small amounts (~ 10 – 100 ppb) significantly change the radiation balance of the glacier surface, absorbing solar radiation in a wide range of wavelengths and reducing the albedo by 1–5% [92]. Brown C (a product of *low-temperature* combustion of woody or herbaceous biomass, household waste, fossil fuels, aerosols from dispersed OM, microorganisms) is capable of absorbing light mainly in the short wavelength range. Brown C contains two large groups: humic-like substances—HULIS, which are similar to OM in soils and aquatic environments and are ubiquitous in the atmosphere [87], and tar materials [60, 246]. The latter are formed during slow combustion of biomass [60] and have the form of spherical nanoparticles (30–500 nm), which are morphologically different from soot. It is believed that brown C is more chemically labile than black C. It is transported over long distances, settles on the surface of snow and ice, where it is involved in the life cycle by microorganisms and then re-enters the atmosphere in the form of CO_2 [246, 248].

We assume complex and multidirectional influence of C-containing components of the supraglacial zone on the radiation balance of glaciers. On the one hand, the accumulation of black C in lighter cryoconite, in which brown C dominates, can enhance the ablative effect; on the other hand, the encapsulation of black C particles inside cryoconite granules can lead to a decrease in light absorption. This complex relationship has not yet been quantified [64].

Other types of anthropogenic agents. More recently, evidence has begun to emerge of the presence of *microplastics* on glaciers in Iceland, the Alps, the Tibetan Plateau, the Andes, and other regions [34, 208, 260, 261] in quantities comparable to microplas-

tics contamination of sea ice, coastal sediments, and some soils. Colored microplastic particles among other organic sediments on a glacier also reduce albedo and can affect the content and rate of transformation of OM and their surfaces, being chemically and mechanically etched during transport and sedimentation, sorb other organic pollutants, and serve as new niches for microorganisms [261]. *Pesticides* are another component of OM that accumulates on ice [81]. Experiments in situ showed that, in addition to photo- and chemical degradation, some of pesticides, such as the organophosphorus insecticide chlorpyrifos, can be biodegraded on mountain glaciers [81].

Radiocarbon age of OM. Just as in soil, OM on the glacier surface consists of several pools with different radiocarbon ages. Despite the presence of a large number of primary producers in the supraglacial zone, radiocarbon ages determined from total organic C are often found to be ancient. For example, the age of cryoconite on the glaciers of Svalbard varied within 8200–3700 radiocarbon years (BP) [13]; in the central part of Norway, it exceeded 5000 BP [94]; in Alaska, it varied within 15000–1000 BP [150]; and in Antarctica (Queen Maud Land), within 11000–4600 BP [145]. This is also true for the DOM fraction, for example, in cryoconite and surface ice of glaciers in Alaska and Wyoming [215] with an age of 7800–2640 BP. The retreating glaciers of the European Alps release biogeochemically diverse DOM, part of which consists of ancient C (8500–600 BP) [199].

Sources of pre-aged C on the surface of glaciers can be aerosols formed during fuel combustion [215], rainfall with ^{14}C age of dissolved C up to 8500 years [181], aerially introduced soil material of Holocene age, as well as C from soils and sediments formed before the last glacial maximum and buried in the body of glaciers [112, 118]. An admixture of 1% of “dead” in terms of radiocarbon activity geological C (kerogen, graphite, coal, fossil fuel combustion products, etc.) results in an error of about 80 radiocarbon years. Even a significant admixture of “dead” C at the level of 50% leads to an aging equal to one half-life of ^{14}C (~5700 years), and cannot fully explain the Early Holocene age of the supraglacial material [13].

A mixture of ancient (^{14}C -active and “dead”) and modern C gives a whole spectrum of different radiocarbon ages, depending on the relative proportion of these components. Therefore, an apparent radiocarbon age of several thousand years based on the total organic C only indicates that the material actually contains some amount of ancient carbon [94]. To approximate the true age, it is necessary to construct a model that takes into account the contribution of OM pools of different ages. For example, if we assume that supraglacial DOM originates exclusively from the pools of “dead” and modern C, then a sample aged 2640 BP would contain 72% recent and 28% fossil C,

and a sample aged 7800 BP would contain 38% recent and 62% fossil C, respectively [215].

In addition, ^{14}C ages are not the same in different fractions of supraglacial material. Separation of cryoconite OM by densitometric fractions [13] made it possible to clarify that fresh OM (<100 yr) from Aldegonda Glacier on Svalbard was contained in the lightest fraction of free particulate OM (FPOM), and ancient OM (11120 ± 40 BP and 8850 ± 30 BP) was contained in heavy fractions (HF1 and HF2) associated with organomineral complexes and adsorbed on the mineral matrix.

Ancient C does not necessarily mean inert. Heterotrophs, both microorganisms and invertebrates, quickly settle on surfaces that have just been freed from the glacier, and actively consume ancient glacial C, ahead of the classical stage of settlement of primary producers, and thereby turning the very concept of primary succession [47, 93]. Moreover, the bioavailability of ancient glacial C may be positively correlated with an increase in its radiocarbon age [105]. Thus, ancient C that melted from the surface or body of the glacier during its retreat has great potential for inclusion in the modern cycle and food chains. This causes a kind of glacial reservoir effect [80], when the radiocarbon age of modern invertebrates living in the glacier forefield and feeding on ancient C, which passed through several links in the food chain, exceeds 1000 years [93].

The most interesting options seem to be those that combine a high degree of biogenicity of glacier material, abundant microbial biomass, and ancient age of OM. For example, cryoconite from Aldegonda Glacier on Svalbard contained 4.6% organic C ($\delta^{13}\text{C} = -25.83\text{‰}$) and 0.2% N, and its age was 7540 ± 25 BP [13]. In this case, the microbial community structure may have shifted toward a heterotrophic block that consumes and recycles ancient C along with modern synthesized microbial photoautotrophs.

On the other hand, with ^{14}C dating of the phospholipid fatty acid (PLFA) fraction of microbial origin, it was shown that cryoconite microorganisms can consume C, which was recently in equilibrium with the atmosphere and was assimilated by primary producers, and at the same time practically do not use ancient C [150]. At least two weakly overlapping C pools have been identified on the glaciers of southern Alaska, which differ in stable isotope composition: (1) a small C pool of a supraglacial microbial community and (2) a large pool of predominantly inactive C of geological origin [150].

Thus, with a variety of sources and composition of OM, the interpretation of the measured ^{14}C activity of supraglacial material is no less complicated than for soil OM. Moreover, in supraglacial systems, as in other nonequilibrium systems, organic C pools have different turnover rates.

DIVERSITY AND GENESIS
OF SUPRAGLACIAL ORGANOMINERAL
FORMATIONS FROM THE POSITION
OF SOIL SCIENCE

Organic components on glaciers rarely exist in isolation and interact closely with mineral matter, which leads to the formation of supraglacial organomineral systems. Here we consider the main types of such systems that exist in various environments on glaciers and also on gravelly to fine earth substrates underlain by ice.

Critical zone of glacier. The glacier surface in the ablation zone is never clean (Figs. 2a and 2b). *The photic layer* on glaciers in most cases has a thickness of up to 2 m, although the short-wavelength part of the solar spectrum can penetrate down to 10 m. Unlike deep ice with limited migration of components, including microbial cells [178], supraglacial photic layer under ablation conditions has a dynamic pore space and is permeable to living and dead OM, as well as mineral particles that migrate with meltwater flows, both vertically and laterally [65, 115]. In this case, vertical heterogeneity appears in the upper few meters on the glacier surface (Figs. 2e and 2f). Horizons arise that differ in the ice crystal structure, shape and pore volume, saturation with meltwater [115], as well as the concentration of organomineral components, which can have a high renewal rate on the surface (fast pool) or remain in the near-surface layer for $n \times 10$ years (slow pool). By analogy with hard rocks, foreign colleagues called this zone a *weathering crust* [115, 155]. During warm periods, especially with cloudless days, the total porosity of near-surface ice progressively increases, and the weathering crust expands vertically to greater depths. In Russian, when applied specifically to glaciers, the term weathering crust probably sounds too radical, which does not negate the essence of the phenomenon of supraglacial organomineral system with transformed and genetically interconnected horizons. In our opinion, such a layer can also be considered as a supraglacial component of *the critical zone*, which is still missing in the concept of critical zone, including critical zone for cold regions [171]. The upper part of such a critical zone is represented by *the melting crust*. This is a generally accepted term in glaciology to designate the upper, highly loosened and highly porous layer of ice with a thickness of a few tens of centimeters, which appears in the region of glacier ablation under the influence of solar radiation [17].

The development of drainage between ice crystals, as well as between cryoconite holes (Fig. 2c) and pore tubes (Fig. 2d), provides a mechanism by which solute-rich water is distributed over the surface and inside the glacier, and its flow turns into a developed hydrological network, including the deep internal drainage system of the glacier [17]. It has been established that microbial cells, organic and inorganic substances penetrate into the melting crust and deeper into the weathering crust/critical zone of the glacier and are

transported along them by meltwater [116]. Depending on the pore size, selectivity arises in the transfer and precipitation of solid particles of different size fractions and associated microorganisms [65]. The migration of microorganisms through the weathering crust of a glacier changes their lighting conditions and access to nutrients, thereby affecting the structure and function of the ecosystem [65]. Diurnal and seasonal changes in the permeability of the weathering crust control the accumulation and movement of cells and nutrients. The structure and function of the microbial community may depend on depth: the horizon near the ice surface is associated with high rates of photosynthesis and the need for photoprotective pigmentation, while the deeper horizon contains organisms with better adaptation to low light and the ability for pure heterotrophy [65]. Studies of ciliates in cryoconite holes revealed a clear stratification of species depending on the depth of immersion in meltwater. Thus, heterogeneity in biotic and abiotic conditions influences microbial communities at the microscale [114].

In 2012, Tashirev et al. [24] were the first to describe the process of soil formation directly on a glacier. Organomineral bodies (Fig. 2i) in the supraglacial zone of Galindez Island (West Antarctica) were vertically stratified into a surface organic horizon with algal–bacterial communities and an underlying horizon with humified OM having a water-resistant structure. Incomplete destruction of biomass was apparently determined by a deficiency of nitrogen, phosphorus, and sulfur, which are necessary for microorganisms to decompose OM from algal–bacterial mats into final products. The formations were distinct from cryoconite (Figs. 2c and 2d), as well as from such widespread but ephemeral phenomena as green or pink snow (Figs. 2g and 2h), associated with the development of algal communities, and were identified as “ice” (glacial) soils. *Ice soil*, according to Tashirev [24], is a dense organic formation on horizontal areas of ice containing a layer of necromass of algal–bacterial cenoses humified by heterotrophic microorganisms. In addition to the presence of biota and ephemeral functioning horizons that differ in temperature, moisture, and OM composition (typical of green/red ice and snow), an important property of ice soil is the in situ accumulation of residual solid-phase functioning products (according to Targulian [23]) in significant quantities, which can be diagnosed macromorphologically as the dark “humus” horizon (Fig. 2i).

Microprofiles under plant communities. On the surface of glaciers, the formation of soils with a microprofile under moss communities is possible. This is due to the phenomenon of so-called “glacier mice” [4], namely metastable spherical (ball-like) colonies of mosses (Figs. 3a–3c), either fixed on the glacier surface or moving along it like tumbleweed plants in arid regions [107].

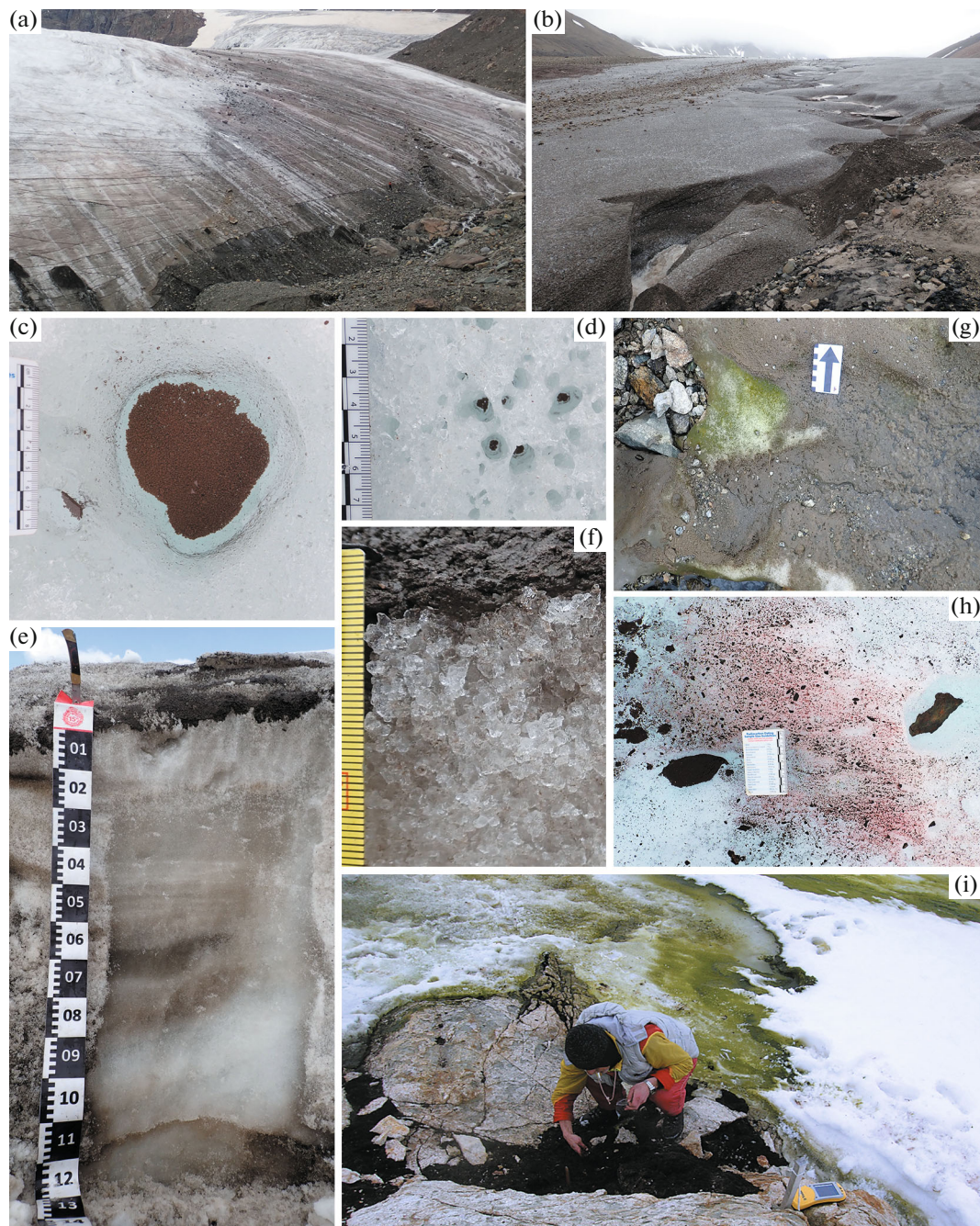


Fig. 2. Critical zone of the glacier and ice (glacial) soils: (a) dirty supraglacial zone on the Levyi Aktru Glacier tongue, Altai; (b) analogous zone on the Bertil Glacier tongue, Svalbard; (c) large glacier hole with cryoconite; (d) tubular pores and small holes in the melting crust with the initial stages of the formation of cryoconite granules on Bertil Glacier, Svalbard; (e) profile of the critical zone (weathering crust) on Garabashi Glacier, Caucasus; (f) penetration of cryoconite between ice crystals in the upper part of the critical zone (Garabashi Glacier, Caucasus); (g) green snow with cyanobacteria and eukaryotic microalgae under cryoconite on Mizhirgi Glacier, Caucasus; (h) red ice with *Chlamydomonas nivalis* algae on Levyi Aktru Glacier, Altai; and (i) ice (glacial) soil on a horizontal part of glacier surface on Galindez Island, Antarctica [24].

In Figs. 3d–3i, a microprofile of poorly developed soil under a ball-like colony of moss attached to the organomineral material of cryoconite in a small hole directly on the surface of Bertil Glacier (Svalbard) is shown. The microprofile reaches 4–6 cm and consists of four microhorizons, including analogues of such soil

horizons as moss litter with some peat material (O), moderately decomposed OM with inclusion of raw humic material (O/A), weakly developed humic or proto-humic layer (A), and transitional layer to assumed parent material (BC). Granular organomineral material of cryoconite is ubiquitously present

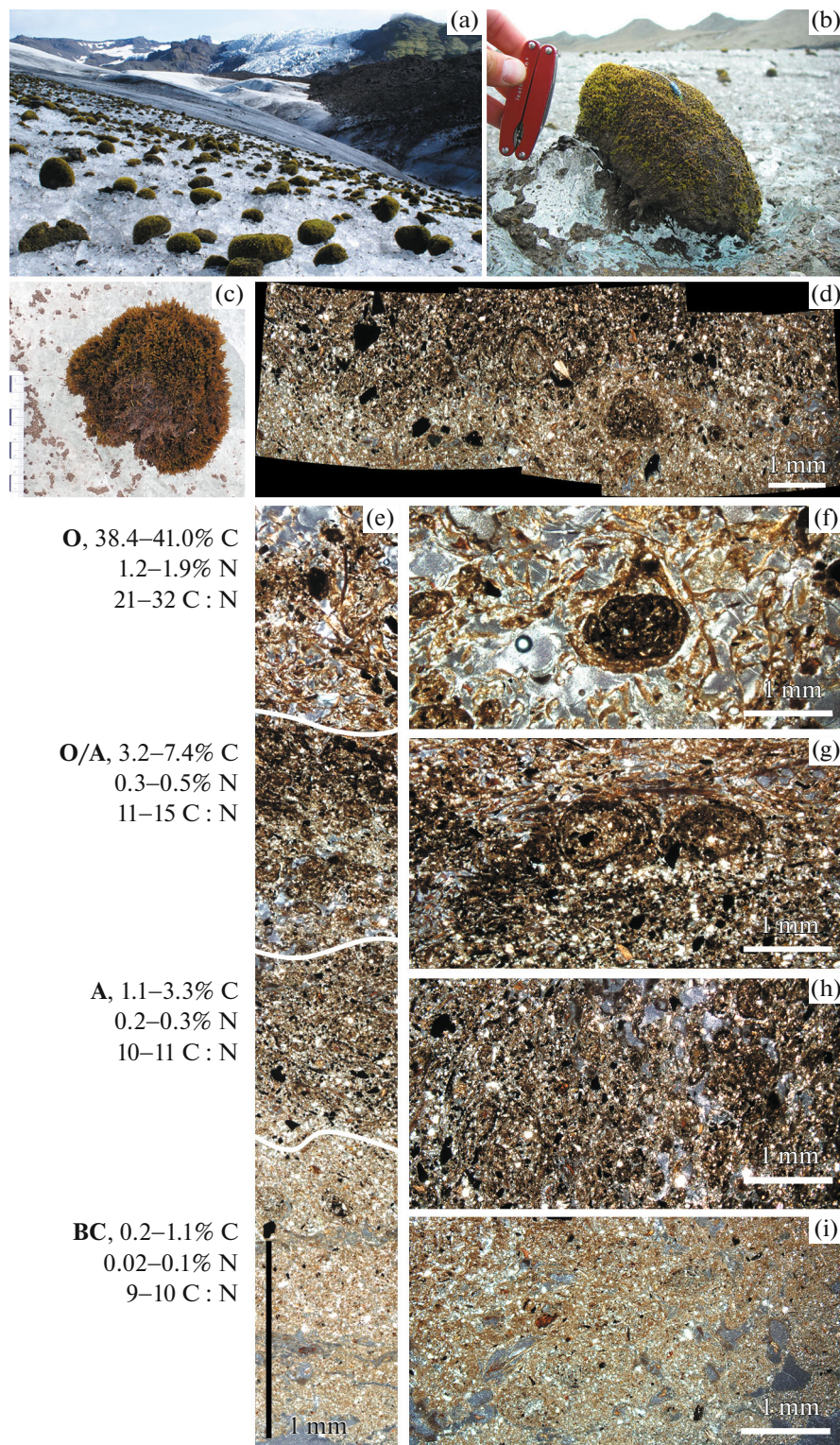


Fig. 3. Moss communities on glaciers: (a) Virkisjökull Glacier, Iceland (photo by Darrel Swift); (b) Ruth Glacier, Alaska, USA (photo by Tim Bartholomaus); (c–i) soil microprofiles under the moss community on Bertil Glacier, Svalbard (photos by N. Mergelov); (c) ball-like moss cushion on ice—“glacier mouse” (view from the top), (d) vertical panoramic section through the O/A and A horizons, (e) full microprofile, (f, g, h, i) vertical sections for each horizon separately. All micromorphological images are in transmitted light, PPL.

between the moss rhizoids. Apparently, moss communities use biogenic elements (C, N, P), which are abundant in cryoconite. The O horizon contains 38.4–41.0% C_{org} (C), 1.2–1.9% N, and the C : N ratio is in the range of 21–32; the O/A horizon, 3.2–7.4% C, 0.3–0.5% N, C : N = 11–15; the A horizon, 1.1–3.3% C, 0.2–0.3% N, C : N = 10–11; the BC horizon, 0.2–1.1% C, 0.02–0.1% N, C : N = 9–10. The BC horizon, which is used by moss communities as the main substrate for attachment and acquisition of nutrients, consists of quartz, feldspar, calcite, aragonite, and fossil coal particles (unpublished author's data on Raman spectroscopy). Its mineral composition reflects the composition of the rocks surrounding the glacier and indicates the dominance of local material. Morphology, stratigraphy, and organo-accumulative type of OM distribution indicate the formation or preservation of the soil microprofile under moss on the glacier surface.

Although the phenomenon of “glacier mice” has been described many times by botanists and glaciologists [4, 48, 96, 107], we note signs of initial pedogenesis and soil microprofiles in these objects for the first time. It is believed that these are ephemeral formations, and rotational mechanisms (rolling and turning over of moss cushions), as well as sliding, play a significant role in their movement along the glacier. However, the discovered soil microprofiles clearly indicate long periods of stability in the existence of these objects. Such profiles could not be formed within one season; the duration of soil differentiation processes had to reach at least several years. Here we highlight two hypotheses of the soil microprofiles formation: (1) periglacial origin—a cushion of moss with an attached substrate differentiated to the horizons was brought from the landscapes surrounding the glacier and (2) supraglacial origin—the microprofile was formed directly on the glacier, when moss was fixed on a mineral substrate, for example, on clastic ablative moraine material; the spores or protonema of the moss were carried by wind or birds. Inclusion of cryoconite granules in all horizons of the soil profile (in “glacier mouse”) and close integration of granules with moss rhizoids and mineral mass additionally indicate joint long-term formation in a supraglacial setting.

Botanical studies [48] have shown that moss cushions appear on the glaciers of Svalbard not only as a result of introduction from the surrounding landscapes, but also during the development of plants from protonema in situ, in particular, the branching of mosses on cryoconite accumulations rich in nutrients. Each moss cushion represents a small ecosystem on the ice, goes through several successional stages, and provides habitat for other organisms [48]. On Alaskan glaciers, after reaching a mature size, moss cushions persisted for at least six years [107]. When mosses are associated with the mineral fine earth this time is likely sufficient to initiate soil processes and form a microprofile. The upper time limit for the existence of one cushion of moss on a glacier is still unknown.

Cryoconite. The most important product of organomineral interactions on a glacier is cryoconite. It is a finely dispersed, often structured and dark-colored organo-mineral material on the surface of glaciers and perennial snowfields. Cryoconite is a complex system of mineral fine earth and OM from near and distant sources, microorganisms and mesofauna, products of their transformation and, at present, numerous anthropogenic pollutants.

Since, apparently, the first description of cryoconite in the 19th century in Greenland [163, 164], its biogenic granularity, as well as its influence on glacier melt, have been documented. In Russian literature, the term cryoconite was used, for example, by Fersman:

{... in the subpolar regions, accumulations of cosmic dust (cryoconite) are detected, which in other parts of the earth are mixed with earthy dust and are not captured. Although cryoconite still remains a mysterious and little-studied formation, its significance in the general geochemical history of the Earth cannot be denied.} [25].

Glazovskaya, based on microscopic studies of cryoconite, although without naming it as cryoconite, wrote: “... each structural grain is a plexus of multicellular filaments of blue-green algae that hold particles of mineral dust inside the ball” [7]. It is now generally accepted that the granular structure of cryoconites is associated with filamentous cyanobacteria and their extracellular polymers, and such a product of in situ heterotrophic transformation of OM, as humified OM newly formed in cryoconite granules, absorbs more sunlight than the surrounding ice and cryoconite components separately [216, 217].

The formation of granular cryoconite is an extremely common process found on glaciers in both hemispheres. Several detailed descriptions of its morphology and structure were previously available [116, 137, 216, 217, 227, 255]. In 2022, a major synthesis [187] revealed differences in the color, morphology, and geochemistry of cryoconite from 33 glaciers around the world. Importantly, cryoconite was represented by both loose and well-aggregated material, and the occurrence of granules increased with the increasing content of organic C (from 1 to 38% C for 33 glaciers).

Classification of cryoconite. We propose to separate *in situ and redeposited cryoconite* (Fig. 4). *In situ* cryoconite accumulates and acquires structure in the melting crust of the glacier in small pores—tubes and larger rounded forms—cryoconite holes, as well as in their clusters and associations [43, 61, 84, 85, 136, 216, 217]. Such niches provide relatively stable conditions for the formation of:

I. Primary (elementary) aggregates with the participation of filamentous cyanobacteria (according to [187], with modifications), which are divided into:

- aggregates of regular rounded shape—*microgranules* (Fig. 4a);

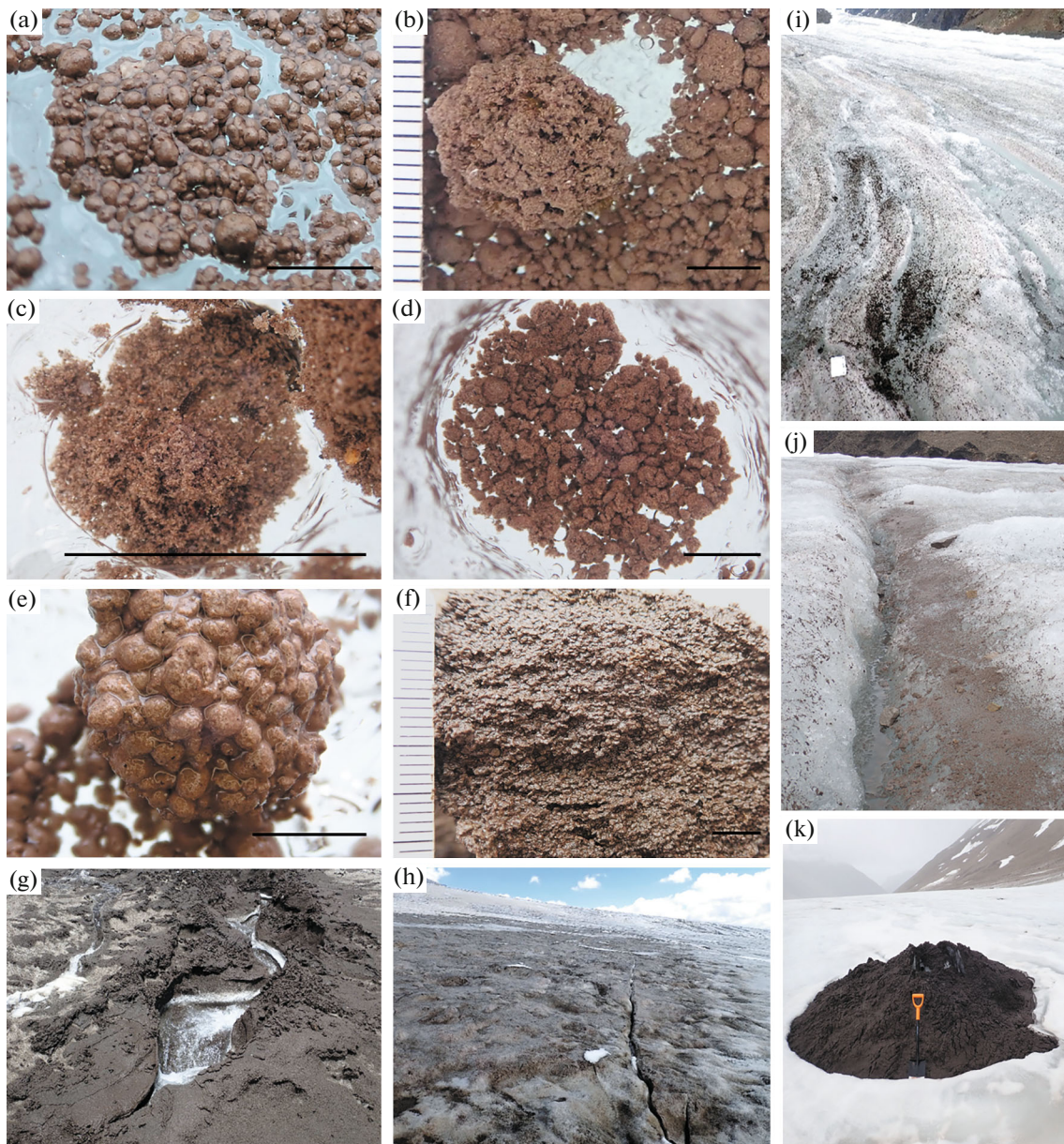


Fig. 4. Forms of cryoconite on glaciers: (a) aggregates of regular rounded shape, (b) rounded aggregates with uneven edges, (c) loose aggregates covered with filamentous cyanobacteria, (d) accumulations of aggregates of various shapes and sizes, (e) secondary associations of primary aggregates >1 cm in size (megagranules), (f) preserved granular structure in a fresh accumulation of redeposited cryoconite, (g) accumulation of redeposited cryoconite, (h) fields of dispersed cryoconite, (i, j) striated forms of redeposited cryoconite, and (k) redeposited cryoconite of conical shape with an ice core. Images were taken at (a–f, j, k) Bertil Glacier, Svalbard; (g, h) Garabashi Glacier, Caucasus; (i) Levyi Aktru Glacier, Altai. Scale bar (a–f) is 5 mm.

- rounded aggregates with uneven edges (Fig. 4b);
- weakly formed loose aggregates covered with filamentous cyanobacteria (Fig. 4c);
- accumulations of aggregates of various shapes and sizes (Fig. 4d);

II. Complex rounded aggregates with concentric layers and stratification between the zones of primary production and transformation of organic matter, and with redox microprofiles in the center–periphery direction—*mesogranules* (e.g., Figs. 5b, 5d, 5h);

III. Secondary associations of primary aggregates with a total size of 1 cm or more—*megagranules* (Fig. 4e).

Based on the characteristics of their internal structure, granules can be divided ([221] with modifications) into (Fig. 5):

- granules with concentric layers (*type 1*);
- granules with several subgranules inside (*type 2*);
- granules without a specific internal structure (*type 3*);

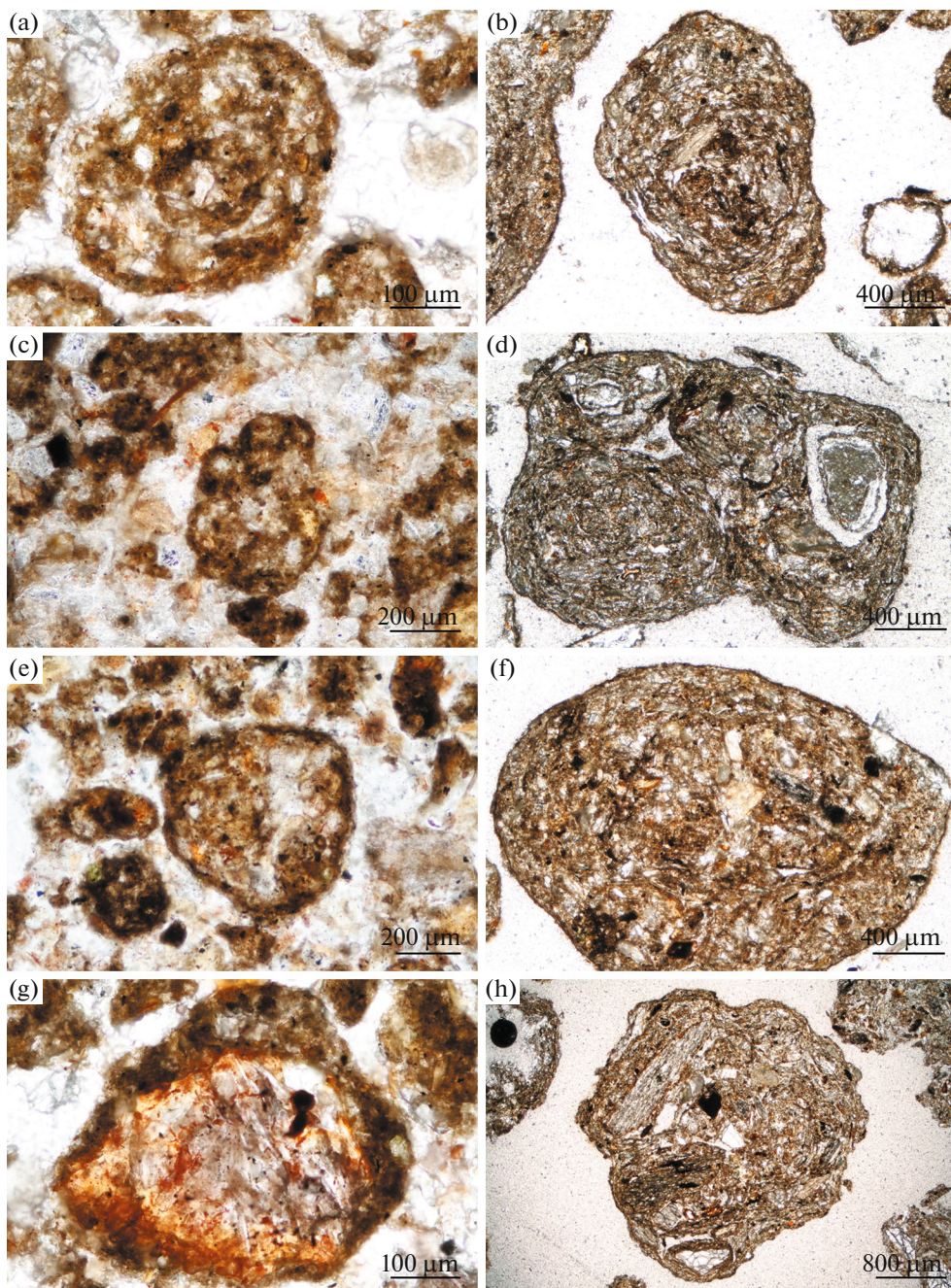


Fig. 5. Four types of internal structure of cryoconite granules: (a, b) granules with concentric layers, (c, d) granules with several subgranules inside, (e, f) granules without a specific internal structure, and (g, h) granules with one or more embryonic grains of minerals. All granules are impregnated with brown organic matter, the content of which is increased in the walls of the granules. Images were obtained in transmitted light, PPL. Left column—granules from Garabashi Glacier (Caucasus); right column—granules from Levyi Aktru Glacier (Altai).

– granules with one or more large mineral particles inside—embryonic grains (*type 4*).

Such morphotypes of granules could occur in various combinations.

In the unstable conditions of the supraglacial zone, cryoconite undergoes multiple translocations within daily, seasonal, and annual cycles and produces new

accumulations of *redeposited cryoconite* (Figs. 4f–4k) in the form of:

– fresh accumulations with preserved granular structure, often as a result of melting with minimal movement (Fig. 4f);

– fields of dispersed cryoconite (Fig. 4h):

- with a well-preserved granular structure (melting with minimum movement);
- with a disturbed granular structure;
- without signs of granular structure and other biogenic aggregation (according to [187], with modifications):
 - finely dispersed material with a predominance of particles <0.5 mm;
 - material with particles >0.5 mm.
- striated accumulations along meltwater flows (Figs. 4i and 4j);
- large accumulative conical forms with an ice core (Fig. 4k) protected by the cryoconite layer of several centimeters in thickness [162]. In glaciology, similar forms are called “ant heaps,” however this term is broader, since the layer that screens solar radiation is often formed not by cryoconite, but by moraine of various types, as well as by sediments of problematic genesis [16, 19]. Some researchers classify such forms as specific supraglacial kames [210, 211].

The morphotype of cryoconite granules depends on the nature of the glacier (ice cap, glacier of cirque or valley, ice sheet, etc.), season, specific year, position on the glacier. For example, in the upper part of the ablation zone, granules may be absent, and closer to the glacier tongue they can be large and darker; within the season they can be disintegrated due to rain, active melting, gravitational movement, as well as katabatic winds [187]. There is a relationship between the size of granules and the content of C and N in them: the larger the granule, the more OM it contains [235]. It is assumed that the size of cryoconite granules, which correlates with OM content and color, has a decisive influence on the ablation rate [135, 252].

The color of cryoconite depends on its mineralogical composition and texture, presence of black C, and, to a greater extent, on the content and structure of in situ OM, the degree of its transformation and interaction with mineral components. These principles of color formation are similar to those realized in soil. The higher the OM content in cryoconite, the more likely the formation of granules and the darker their color. The effect of OM on the color of cryoconite was demonstrated in a simple combustion experiment at $T = 550^{\circ}\text{C}$, after which even the darkest granules became significantly lighter throughout the entire volume of the aggregate [187]. Consequently, it was the impregnation with dispersed OM that made a significant contribution to the darkening of the granules. In the case of Antarctic cryoconite in stable cryoconite holes, approximately 35% of the surface of the granules is covered with microbial biofilms, of which diffuse impregnation with extracellular polymers accounted for ~19% [203]. The dark color of OM can be explained by the synthesis of microbial pigments, for example, melanin [190]; the residual accumulation of highly polymerized compounds formed during the bacterial decomposition of OM [137, 216]; as well as

the sorption of allochthonous particles of black carbon. Particles of black C produce a strong but spotty color change and cannot cause uniform darkening. However, the latter is easily explained by the primary production of OM and its subsequent transformation in situ, which involves DOM and mineral-associated OM.

Thus, in addition to the mineralogical composition of granules, for example, the initial presence of melanocratic minerals (pyroxenes, amphiboles, biotite, etc.) and the aeolian supply of black C, the synthesis of OM by microorganisms and its subsequent in situ transformation serve as important additional drivers of the biological albedo reduction on glaciers [216, 217, 220, 222].

Mechanisms of granule formation in cryoconite have not been fully elucidated. On the one hand, this is the entanglement of mineral particles by filamentous cyanobacteria and the formation of glomeruli—essentially microbial mats of a quasi-spherical shape [7, 216, 217]. The role of adsorption and adhesion with the participation of clay minerals in the composition of eolian dust is noted, as well as the binding role of extracellular polymeric substances (polysaccharide matrix) secreted by cyanobacteria and other microorganisms [136, 137, 203]. Extracellular polymers glue organic and inorganic components into aggregates of regular or irregular round shape [136, 137, 221, 235]. The polysaccharide matrix of microbial origin has a large specific surface area on which particles are adsorbed due to electrostatic forces. In addition, the matrix contains hydrophilic and hydrophobic biopolymers. The latter play an important role in the stability of the granules. According to Raman spectroscopy, methyl functional groups ($-\text{CH}_3$) are confined specifically to the walls of the granules, which indicates the hydrophobic properties of their surface necessary for stabilizing the microecosystem of the granule in meltwater (unpublished data by the authors). Microaggregates with the high polysaccharide content may trap suspended solids from meltwater streams in the supraglacial zone of the glacier.

Microbially induced formation of round aggregates (granules) occurs in a wide range of conditions with diverse composition of microorganisms and their metabolism. It was recently shown [167] that cryoconite granules and the so-called oxygenic photogranules formed in wastewater treatment systems [33] are similar to each other and represent OM-rich quasi-spherical microbial aggregates. How can granules with significant similarity form in such different environments—sewage sludge and glacier meltwater? The fact is that in both cases a universal mechanism operates: filamentous cyanobacteria, which form mats in a wide range of conditions, envelop mineral particles and form round aggregates. This network of filamentous organisms and extracellular polymers stabilizes the environment by connecting with other microorganisms and mineral particles. Among the conditions for

the formation of microbial-induced quasi-spherical aggregates, the occurrence of representatives of the Oscillatoriaceae order in the microbial community (primarily, *Oscillatoria* and *Phormidium* genera of cyanobacteria) is important, as they secrete abundant extracellular polymers and are capable of sliding movement along a solid surface [207] towards light and away from it (phototaxis). The movement of cyanobacteria together with extracellular polymers enhances the binding of mineral particles and also attracts other microorganisms that use cyanobacterial metabolites (polymers and oxygen), promoting further aggregation of cryoconite [136, 216, 217].

For the formation of granules, both hydrostatic conditions in the cryoconite holes, which promote the growth of spherical microbial mats, and the presence of a small hydrodynamic shift (with a slow flow of water through the holes and their coalescence), which leads to further formation of the granules, may be important [167]. In our opinion, there is a bias in the interpretation of the genesis of cryoconite granules towards their exceptional biogenicity, while the role of the formative mechanical impact has been poorly studied. Indeed, benthic cyanobacteria grow in stable environments, for example, on gently sloping mountain glaciers and ice sheets with limited surface water movement, which is favorable for the formation of granules [187]. However, these conditions are not strictly necessary for cryoconite granulation. Under conditions of intense ablation, granules are also formed on the glaciers of Svalbard [256 and authors' observations]. Moreover, cryoconites from mountain glaciers at high altitudes are often better granulated than cryoconites from flatter, lower-lying glaciers in the Arctic and Antarctic. For example, the microaggregate composition of fine earth from the Tien Shan glaciers [8] revealed the predominance of mesogranules 1–3 mm in size. It is likely that cryoconites on glaciers with a steeper surface, experiencing hydrodynamic influence, can in some cases explain better granulation [167].

An analysis of the microscopic structure of cryoconite granules from the Bertil and Aldegonda polar glaciers (Svalbard) and the Garabashi mountain glacier (Caucasus) revealed that their internal part consists predominantly of larger mineral particles, tens of microns in size, while the periphery is composed of densely packed smaller particles <10 μm (authors' unpublished data). In the wall of a granule, flat mineral particles are often oriented subparallel to the outer surface of the granule, whereas within the granule the particles are randomly oriented. Inside the granular wall, mineral particles form concentric layers and are usually held together by a polymer matrix. The presence of biopolymers and the specific orientation of mineral particles in the wall of the granule indicate the combined action of both biological and mechanical factors in the structuring of the cryoconite material.

It is assumed that the concentric layers reflect stages in the formation of granules, a kind of "growth rings" that are formed at a rate of ~ 0.2 mm/year [221]. This is a promising, but so far undeveloped direction for determining the age of granules. It should also be taken into account that the proliferation of cyanobacterial biomass and extracellular polymeric matrix, on which mineral particles "settle," can occur several times during the season.

The morphology of cryoconite is also similar to ooids in periglacial soils. Soil scientists [202, 237] pointed to the role of cryoturbation and solifluction (gelifluction) in the formation of round granular structures in Cryosols. In general, in paleogeographic reconstructions, ooids are interpreted as indicators of periglacial and/or permafrost environments [236]. On the one hand, by analogy with soils, gelifluction processes (movement along a frozen surface) can play a role in the formation of cryoconite granules; on the other hand, some of the ooids in periglacial soils can be granules preserved during the redeposition of cryoconite.

Stratification in cryoconite holes and granules. According to [176], an anaerobic layer of several millimeters in thickness is formed already one hour after the deposition of organomineral material in a cryoconite hole. The vertical heterogeneity of oxygen concentration is influenced by the structure of the sediment, for example, the size of cryoconite granules, the thickness of the meltwater layer, as well as by oxygen consumption during heterotrophic respiration. The resulting oxygen profile, in turn, changes the structure of the microbial community, and the activity of anaerobes increases in oxygen-free microniches [176].

Similar zoning of redox conditions, or redox stratification, occurs within large cryoconite granules [197]. Measurements of O_2 , NH_4^+ , NO_2^- , and NO_3^- concentrations with microsensors revealed that on the surface of granules colonized by cyanobacteria, aerobic conditions prevail, and photosynthesis and denitrification occur. In the central part of the granules, anaerobic conditions can form, and microbial decomposition of OM, nitrification, and denitrification occur. Thus, cryoconite granules determine the heterogeneity of microprocesses and the formation of solid-phase functioning products. If cryoconite material is widespread, this affects the carbon and nitrogen cycle in the supraglacial zone as a whole. When there is an abundance of quartz grains in cryoconite and light penetrates through them, primary production can occur not only on the surface of cryoconite granules but also in their internal part [99]; the existence of peculiar endocryoconitic primary producers is assumed. Elementary microgranules may not have zoning of redox conditions.

Thus, the main structural unit of cryoconite, a granule, is an independent, complexly organized multi-order microecosystem, both isolated in an aquatic supraglacial environment, but also permeable

to flows of matter owing to a developed pore system. The physical and chemical mechanisms of stabilization of organomineral aggregates of cryoconite are similar to the principles of aggregation in complex colloidal systems, such as soils: (1) occlusion/encapsulation; (2) formation of granules/ooids; (3) adsorption on active mineral surfaces, such as clay minerals; and (4) hydrophilic–hydrophobic interactions.

The influence of cryoconite on the periglacial zone. Cryoconites are an important source of microorganisms for primary successions on glacier deposits immediately after its retreat [119], and therefore for the initiation of soil formation. However, the role of cryoconite in these processes has been assessed qualitatively, but not yet quantitatively. In essence, cryoconite is a “bioreactor” of the supraglacial system, in which interaction occurs at different levels: from molecular to landscape. The most active flow of nutrients passes through the cryoconite; quasi-equilibrium temperature and photic conditions are maintained in the cryoconite holes, which is important for the development of microbial communities.

The nearest to glacier parts of periglacial zone are located in the area of redeposition of organic, mineral, and organomineral facies from the supraglacial zone. In the area of terminal moraine and lateral moraine, accumulations of redeposited cryoconite material are locally formed (usually, in the form of lenses and sediments in puddles), on which initial soils can form under stable conditions. Owing to the high content of nutrients and fine dispersion, cryoconite material has more favorable conditions for soil formation compared to moraine material [13]. Therefore, the processes of initial soil formation on redeposited cryoconite proceed an order of magnitude faster than on surrounding clastic substrates, and soils inherit the properties of cryoconite: granularity, layering, and increased contents of C, N, P.

Soils on silicate gravelly fine earth material above glaciers and dead ice. On average, ~4.4% of the area of glaciers (excluding Antarctica and Greenland) is covered with supraglacial debris, the proportion of which increases significantly for high-mountain glaciers [195]. In this case, we do not consider rock glaciers [10], although they may also have soils and soil covers, and their varieties, for example, “ice-cored rock glaciers” [5], are probably genetically close to moraine-covered blocks of dead ice, buried and armored glaciers (debris-covered glaciers).

With a predominance of specific organomineral formations on the surface of glaciers, in some cases there are objects that are classified as soils, and some fully meet all the criteria for soils, and not only underdeveloped soils.

Soils on silicate deposits with shallow embedding by glacier ice. A common case is the phenomenon of the appearance of vascular plants and the formation of soils on ablative moraine and colluvial sediments

accumulating on the surface of a glacier [82]. At the same time, the glacier remains capable of movement.

The soils in these cases remain relatively underdeveloped, but are already classified as soils. Figure 6 shows examples of soil formation in the Ong Valley in the Transantarctic glacier mountains in the dry continental regions of Antarctica, where vascular plants are absent because of bioclimatic reasons [46]. According to the *US Soil Taxonomy*, these soils are classified as Glacic Haplorthels, when the glacier ice is found at a shallow depth, and as Glacic Anhyorthels, when it lies deeper than 50 cm. Analogous thin soils are known on Miage Glacier in the Italian Alps [63] and on Garabashi Glacier in the Caucasus (authors' data). They can be classified as Leptosols according to the WRB system [117] and as gravelly pelozems according to the Russian soil classification system [15]. In all soils, stratification into horizons is noticeable, and the underlying massive ice of the glaciers of Antarctica is at a depth of 10–60 cm. In the case of continental Antarctica, the glacier material, due to very low temperatures, is not as much subject to melting and erosion as in the glaciers of the highlands and polar regions of the northern hemisphere as to the eolian removal of fine earth and the formation of a stone pavement and sublimation of the glacial body, which weakens as the thickness of the quartz–silicate cover above it increases [194]. The organic C content ranges from hundredths to a few percent. The pH values range from highly alkaline with close underlying glaciers to slightly alkaline with increasing depth of the underlying by ice and increase in age of the surface; the salt content increases with time, which is due to the eolian input [194].

Soils on fine earth above blocks of dead ice. It is in this case that full-profile soils can form, although not always. The occurrence of this option is not yet very clear. It is evident that the phenomenon of accumulation of substrate over blocks of dead, i.e., stationary ice, can occur only in cases, where the difference in heights from the accumulation zone to the ablation zone is very significant and amounts to more than one thousand meters. In these cases, the tongues of glaciers “travel” far down the valleys, e.g., into the forest zone. In these already relatively warm conditions, glaciers begin to melt and lose mobility. In addition, the conditions for the possibility of fine earth accumulation on the glacier surface (aside from ablation moraine), for example, volcanic tephra, must be met. Such cases were noted for Andean glaciers in Chile on the border with Argentina at 4° S [180], glaciers of Alaska [209] and Klutlan Glacier in northwestern Canada [193] (Fig. 7).

In the Andes, a transitional case is observed—part of Casa Pang Glacier on the slopes of Mount Tornado is in a moving state, and part is under moraine ridges in an almost motionless state, which made it possible to form here forest communities with a pre-

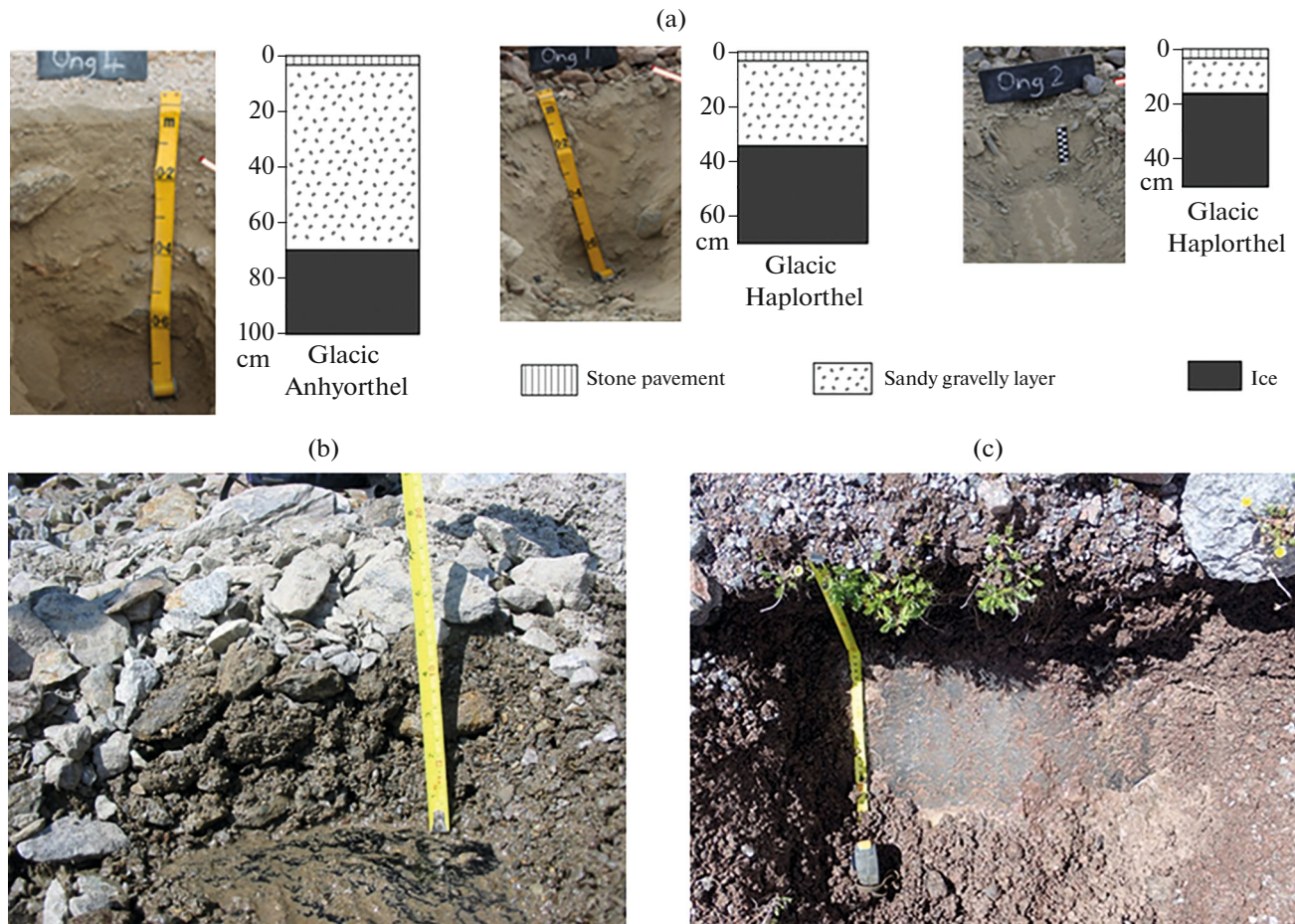


Fig. 6. Soils with close underlying by moving glaciers: (a) continental areas of Antarctica [46]; (b) primitive Leptosols on Miage Glacier, the Italian Alps [63]; and (c) poorly developed supraglacial soils on Garabashi Glacier, Caucasus (authors' data).

dominance of *Nothofagus dombeyi* with dwarf shrubs and ferns in the undergrowth and with mosses and lichens in the ground cover. The thickness of the supraglacial material on the ridges is 3–4 m. The soils are acidic (pH 4.6 to 4.7), N content is from 0.01 to 0.02%, and C_{org} content is from 0.1 to 0.2%, which indicates poor soil development. The Ca content is low, but, in general, the content of the biophilic elements Ca, K, and P is higher than in the glacier material. The authors associate this with soil formation.

Near the glaciers of Alaska, probably, in some locations where true closed forests of Sitka spruce (*Picea sitchensis*) with an alder understory and moss cover grow on thin, loose rocky substrates close to the glacier. The soil under such a forest was described on Kushtaka Glacier (60°25' N, 144°05' W) [209]. It included the following horizons: 13 cm—forest litter, 3 cm—light gray horizon (according to author, A2), 15 cm—dark brown horizon (B), up to 90 cm—very gravelly sandy loamy horizon of gray color (horizon C), underlain by dense clean ice. The author defines this soil as an initial (underdeveloped) podzol and points out that similar soils also occur near Herbert and

Mendenhall glaciers. According to the current WRB classification, this soil can probably (since there are no chemical characteristics) be classified as Protospodic Glacic Cryosol.

Finally, the most developed soils have been described on tephra that accumulated on a moraine with an ice core of Klutlan Glacier (Yukon Province, Canada) lying at a depth of about 2 m (61°36' N, 140°39' W) [193]. The studies were carried out at an altitude of 1050 m. Ice karst is clearly visible here—uneven melting of the ice core leads to the formation of heterogeneous soil and vegetation covers (Fig. 7d, 7e). In the most stable areas, forests of white spruce (*Picea glauca*) with an admixture of birch are formed; in disturbed areas, the vegetation cover is dominated by alder shrubs, and in very severely disturbed areas there is a biological crust of lichens. Accordingly, the authors classified the most developed soils as Brunisols according to the Canadian soil classification. Comprehensive chemical data allowed them to be classified as Dystric Cambisols (Arenic, Gelic, Humic) in the WRB system (Fig. 7g). Less developed soils (Canadian Regosols) are classified as Dystric

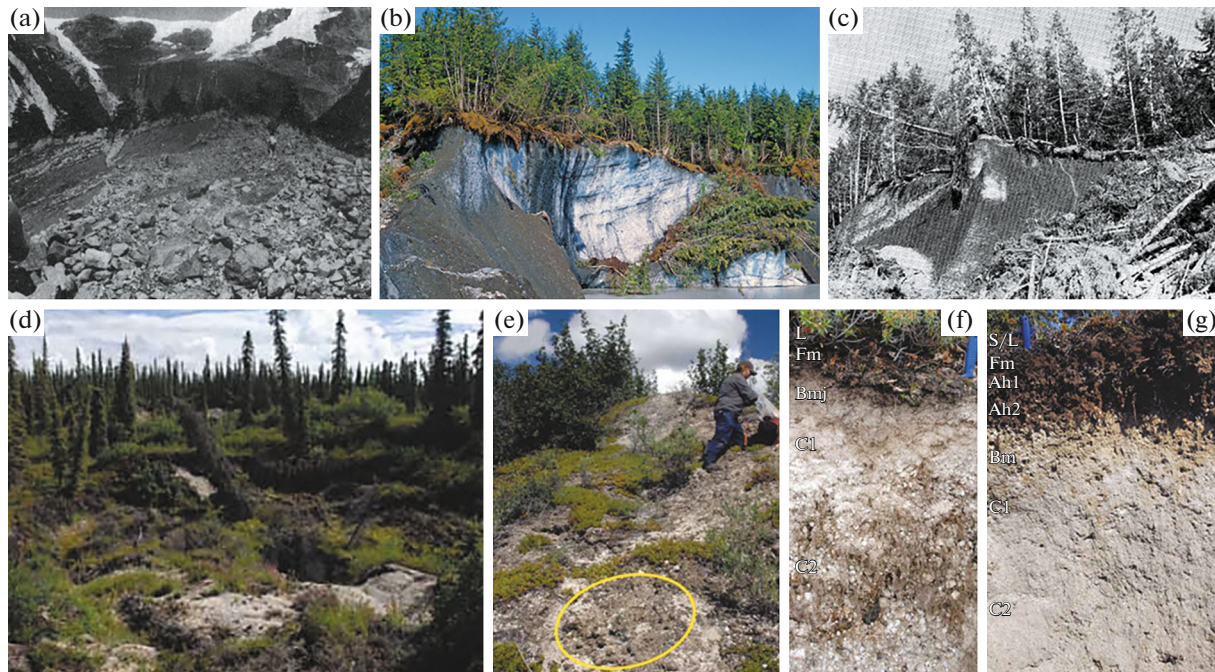


Fig. 7. Soils underlain by weakly moving glaciers and by dead (not moving) ice blocks: (a) Casa Pang Glacier, Chile; *Nothofagus* trees grow on poorly developed rocky soils [180]; (b, c) blocks of dead ice near Alaska glaciers under a well-developed Sitka spruce forest (b) Fairweather Glacier (photo by Tom Bean), no data on soils; (c) Kushtaka Glacier, initial podzol soil underlain by pure ice at a depth of 90 cm [209]); and (d–g) soils on tephra accumulated on a moraine with an ice core of Klutlan Glacier lying at a depth of about 2 m, Yukon province, Canada [193] ((d) general view: forms of ice karst with a “drunk forest” and tephra outcrops on steep slopes, (e) sparse ground cover and biocrust on the surface of steep slopes, (f) Dystric Brunic Arenosol (Gelic), (g) Dystric Cambisol (Arenic, Gelic, Humic)).

Brunic Arenosols (Gelic) (Fig. 7f). On steep slopes of ice-karst forms, tephra and other mineral particles are held together by organic detritus and lichens to form a surface soil crust. The authors do not classify these formations in any way, but according to the WRB, they correspond to Protic Arenosols (Gelic?). Interestingly, despite the significant role of tephra, the soils do not meet the andic and vitric criteria.

TYPIFICATION OF SUPRAGLACIAL ORGANOMINERAL FORMATIONS FROM THE POSITION OF SOIL SCIENCE

Exposed surfaces of ice and snow. Pre-soils. How can one consider thin, not always organized into micro-horizons or structural aggregates organic, organomineral, and mineral heterogeneous dispersed systems on the surface of glaciers from the position of theoretical soil science? Such objects were not directly considered either in the general theory of pedogenesis [22] or in the concept of extreme soil formation [11, 12], because the concept of soil-like bodies (soloids) requires at least some organization of the material. These objects are closest to such an element of soil systems as “pre-soils” [22], i.e., in situ bodies undifferentiated at the level of the solid phase, in which certain zoning according to temperature and moisture conditions and to the functioning of microbial communities can be

observed. However, the concept of soil systems assumes that, over time, such pre-soils will become soils. With systems, such as green/red snow and ice, biofilms, and organomineral fillings between ice crystals of the melting glacier surface, this evolution does not always happen. In the glacier accumulation zone, they may be covered with snow and pass into the glacier body in the form of separate layers; in the ablation zone, they may either (1) be washed away by meltwater, transported down the slope, and end up in the solid runoff along the drainage network, or (2) be organized in cryoconites, which are significantly different supraglacial formations. It is proposed to introduce the concept of *supraglacial pre-soils* for such organomineral systems, among which both ephemeral varieties and those that will develop into soil-like bodies and soils can be present (Fig. 8a).

Soils and soil-like bodies. Previously, glaciologists were interested in the properties of impurities to determine albedo and provenance—the source of the influx of particles. In situ changes, for example, from green and red colors of surface algae to brown and dark gray colors after photochemical and heterotrophic transformation of OM remain virtually unstudied. Before Tashirev [24], the word soil in relation to objects on the open surface of glaciers was mentioned only when analyzing the source of light-absorbing impurities,

and the term soil-like bodies has begun to be applied to cryoconite accumulations only recently [1, 28–30, 151, 175].

The most obvious formations that can be classified as soils without overlapping ice with a cover of clastic (gravelly fine earth) sediments are organomineral systems with long-term residual accumulation of humified OM on horizontal and inclined surfaces of glaciers or snow covers (Fig. 8b). These formations were described on Galindez Island in West Antarctica as *ice (glacial) soils* [24] and are probably more widespread in the subantarctic zone on low islands, where glaciers have a slightly inclined surface, and the climate is relatively warm and humid (compared to continental Antarctica) to ensure relatively thermostatic conditions favoring the development of thick algal–bacterial mats on the surface of snow and ice. Ice soils form in gravitational traps associated with local melting of ice and snow, or on a mechanical barrier such as coarse material or large rock fragments that prevent the removal of OM and its transformation products. The striking feature of ice soils is stratification into macroscopic horizons of primary OM production (organogenic horizons) and into horizons with heterotrophic transformation of OM (humic horizons).

Supraglacial soils should also include *organomineral bodies under moss communities* on a glacier with a minor contribution from mineral material (Fig. 8c). In this case, *soil microprofiles* are formed; they consist of the moss litter (O), transitional horizon with moderately to well decomposed raw-humic OM (O/A), weakly developed humic or proto-humic layer (A) and mineral horizons with the participation of cryoconite and small amounts of unstructured fine earth or gravelly material (BC). These are metastable formations that are either fixed or moved along the surface of the glacier; they can exist and develop on it for at least several years. Such soils with microprofiles are most widespread on the glaciers of Alaska, Svalbard, Iceland, Greenland, South America and Subantarctic islands.

Metastable soil-like bodies on cryoconite (Fig. 8d) can form both inside granules (Fig. 8d₁) with concentric microhorizons of primary production on the surface and heterotrophic transformation of OM inside granules, and on accumulations of redeposited cryoconite (Fig. 8d₂), on which the degradation of OM and temporary accumulation of humification products occur. The structure of large granules is the most complex: the outer horizon is maintained due to the close interaction of clay minerals, filamentous cyanobacteria, and other microscopic photoautotrophs with an abundant polymer matrix, and in the inner concentric microhorizon (core) anaerobic conditions can predominate and residual solid-phase products of OM transformation accumulate. The presence of numerous mineral surfaces (clay minerals, weathered grains of primary minerals) creates additional conditions for stabilizing OM. Metastable soil-like bodies on cryo-

conite appear to be present in a very wide range of conditions on all types of glaciers, except for the most remote interior portions of ice sheets. The lifetime of such soil-like bodies depends on local conditions and varies from weeks to tens of years, for example, in cryoconite holes sealed with an ice lid on the glaciers of Antarctica.

Surfaces with ice covered by silicate gravelly fine earth material. Under conditions of a stronger and more stable polymineral matrix, soil formation passes through the stages of pre-soil and soil-like bodies; after years and decades *soils* with a poorly developed (Fig. 8e) and then with a fully developed profile are formed (Fig. 8f). A wide range of options is possible: from petrozems, gravelly pelozems and psammozems, initial Podzols and Cambisols, to more developed soils, depending on the depth of the underlying ice, the stability of the substrate, and the duration of soil formation. Supraglacial soils on gravelly fine earth material on glaciers and dead ice have been described in Antarctica, the Andes and the Alps, the Caucasus, in Alaska and Canada, and are likely to be very widespread, and their classification diversity is far from being fully elaborated.

Having typified supraglacial organomineral formations, we identify the following processes characteristic of soils and soil-like bodies (Table 1): accumulation of OM, its photochemical and heterotrophic transformation, stabilization of OM on mineral surfaces, formation of dark-colored “humified” OM, accumulation of residual solid-phase products of functioning in situ, aggregation of fine earth and its biogenic structuring, biochemical weathering of mineral particles, and ultimately the formation of micro- and macroprofiles of soils and soil-like bodies on ice or on silicate deposits above the ice.

EXCHANGE OF GREENHOUSE GASES ON THE SURFACE OF GLACIERS AND IN THE PERIGLACIAL LANDSCAPE

Four potentially significant components influencing the cycle of biogenic greenhouse gases (GHGs) (CO₂, CH₄, and N₂O) in the structure of the glacial biome can be identified: (1) the surface of glaciers with an active supraglacial system, (2) periglacial sediments, (3) meltwaters, and (4) subglacial deposits. This reflects a unique feature of the glacial biome compared to others: in the modern “melting” world, it serves as an autotrophic–heterotrophic ecosystem and a donor of nutrients accumulated both during past glacial epochs and in the modern supraglacial zone. Various components of the supraglacial zone and periglacial landscape can be both net sinks and sources of GHGs (Fig. 9).

Surface of glaciers. As calculations based on a few published data show [100], glaciers during the ablation season (usually this estimate is taken as annual respi-

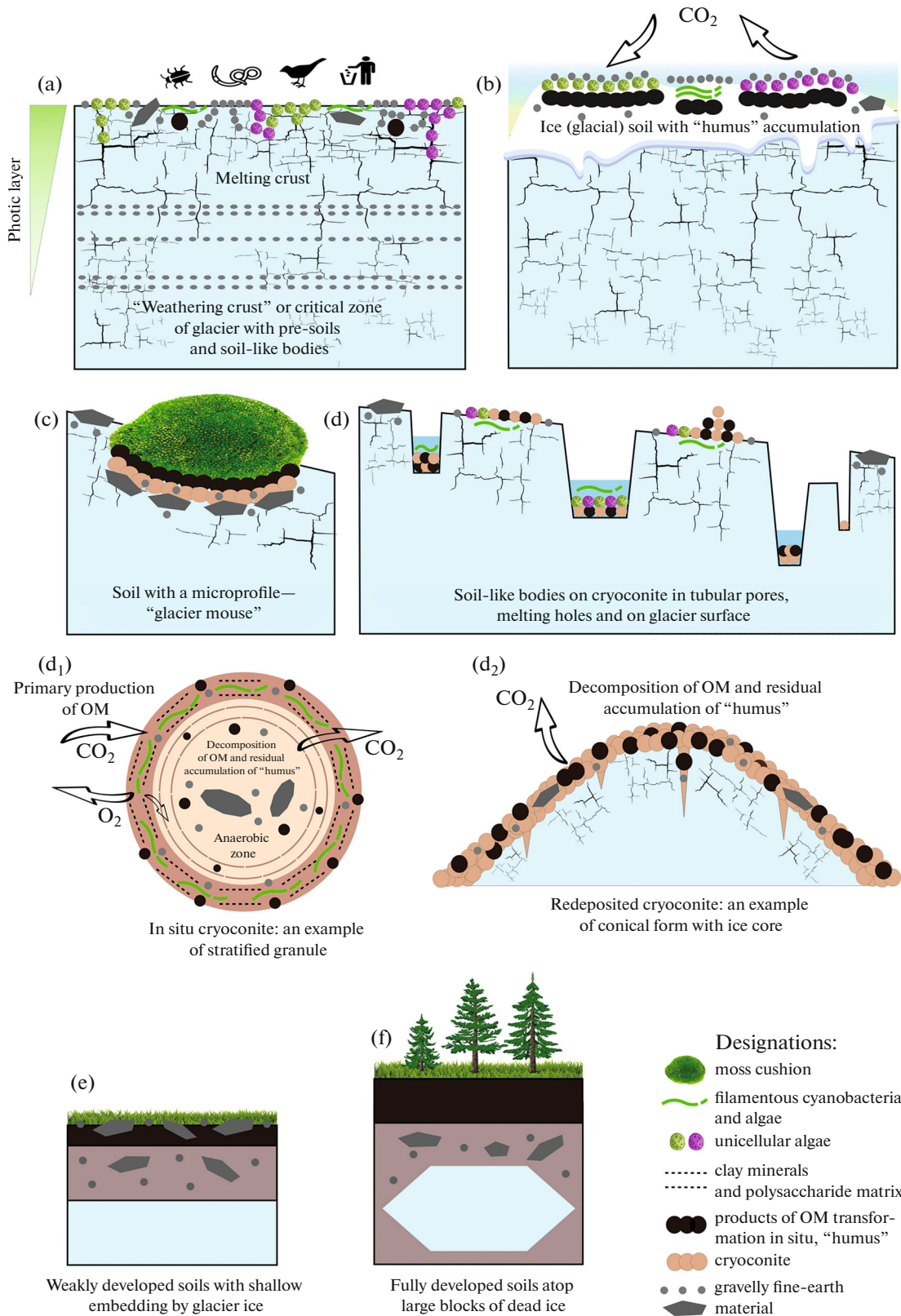


Fig. 8. Typification of supraglacial organomineral formations from the standpoint of soil science: (a) pre-soils and soil-like bodies in ice and snow, (b) ice (glacial) soils with residual in situ accumulation of solid-phase products and macroscopic horizons, (c) metastable soils with microprofiles O(O/A)—A—C under mosses ("glacier mice"), (d) metastable soil-like bodies on cryoconite with microprofiles inside large granules (d₁) and on redeposited cryoconite (d₂), (e) soils on the fine earth—gravelly material on glaciers, and (f) soils on buried/dead ice.

Table 1. Processes in supraglacial organomineral formations characteristic of soils and soil-like bodies

Organomineral formations	Processes										References	
	Accumulation of OM	Photochemical transformation of OM	Heterotrophic transformation of OM	Formation of dark-colored humified OM	Stabilization of OM on mineral surfaces	Accumulation of residual solid-phase products of functioning in situ	Aggregation of fine earth, biogenic structuring	Biochemical weathering	Formation of microprofiles	Formation of macroprofiles		
In ice and snow on glacier												
Green, red, “watermelon” ice/snow	++	++	+	?	+	++	?	?	–	–	[39, 40, 90, 108, 192, 201]	
Ice (glacial) soil	++	++	++	+++	+	+++	++	+	+	++	[24, 39, 40, 192]	
Layered sediments in critical zone of the glacier / annual layers / fine earth between ice crystals	+	+	+	?	?	?	?	+	?	?	Authors’ data	
Under plants on glacier												
Soils with microprofiles under moss communities—“glacier mice”	++	?	+	+	++	++	++	++	+++	–	[4, 48, 107, 110, 177, authors’ data]	
In situ cryoconite on glacier												
Elementary aggregates (microgranules)	++	++	+	++	++	++	++	+	+	–	[137, 187, 192, 216, 217, 221, 255]	
Complex aggregates (mesogranules)	+++	++	++	+++	++	+++	+++	++	++	–		
Redeposited cryoconite on glacier												
Fields of dispersed cryoconite and striated accumulations along streams / conical forms with ice core	+	+++	+++	+++	++	++	+	+	+	–	[Authors’ data, 128, 162]	
On silicate gravelly fine earth material on glaciers and dead ice												
Soils on silicate sediments with shallow embedding by glacier ice	+	n/a	++	+	++	++	+	++	n/a	+	[46, 63, 82, 83, 194]	
Soils on gravelly fine earth material on glaciers and dead ice	++	n/a	++	+	++	++	++	++	n/a	+++	[180, 193, 209]	

Degree of manifestation of the process: weak (+), moderate (++), strong (+++), unknown (?), not applicable (n/a). Microprofiles mean the sequence of soil horizons under moss communities on a glacier, or a redox stratification inside biogenic aggregates (granules) and their accumulations on a glacier.

ration period), taking into account cryoconite in cryoconite holes and dispersed cryoconite, release 12–14 kg C/km² in the course of microbial respiration. This gives a total of only 0.22 million tons of C per year, which cannot be considered a significant contribution to biosphere respiration. Only Russian tundras at the end of the 20th century released 474 million tons of C per year during respiration, which is three orders of magnitude higher [252]. If we take into account primary production on glaciers, it turns out that these are predominantly net-autotrophic systems, where primary production prevails over gross respiration [34, 64, 100], although net-heterotrophic cryoconite ecosystems also occur [99]. However, natural estimates of the actual NEP (net ecosystem production, gross primary production minus ecosystem respiration) on glaciers, to this day, is very small. According to [64], inside cryoconite holes, gross respiration varies from

1.86 to 42.1 μg C–CO₂/(g day), and net balance changes from –2.03 (C-source) to +14.6 (C-sink) μg C–CO₂/(g day). According to the same review, of the 27 publications found in the literature on primary productivity, ecosystem respiration, or the net balance of supraglacial material, only 10 studies contained estimates of the latter indicator covering the period from 2007 to 2015. The authors of [212, 213] conducted a unique large-scale survey along a 70-km-long transect across the edge of the Greenland Ice Sheet and reliably showed that cryoconite ecosystems absorb carbon during melting, i.e., they are predominantly autotrophic. The main factors of the net carbon balance are the influence of meltwater and the steepness of the slope—further from the edge to the center of the ice sheet, the slope becomes less steep, which contributes to an increase in the amount of cryoconite mass in cryoconite holes and its thin-layer distribution in the

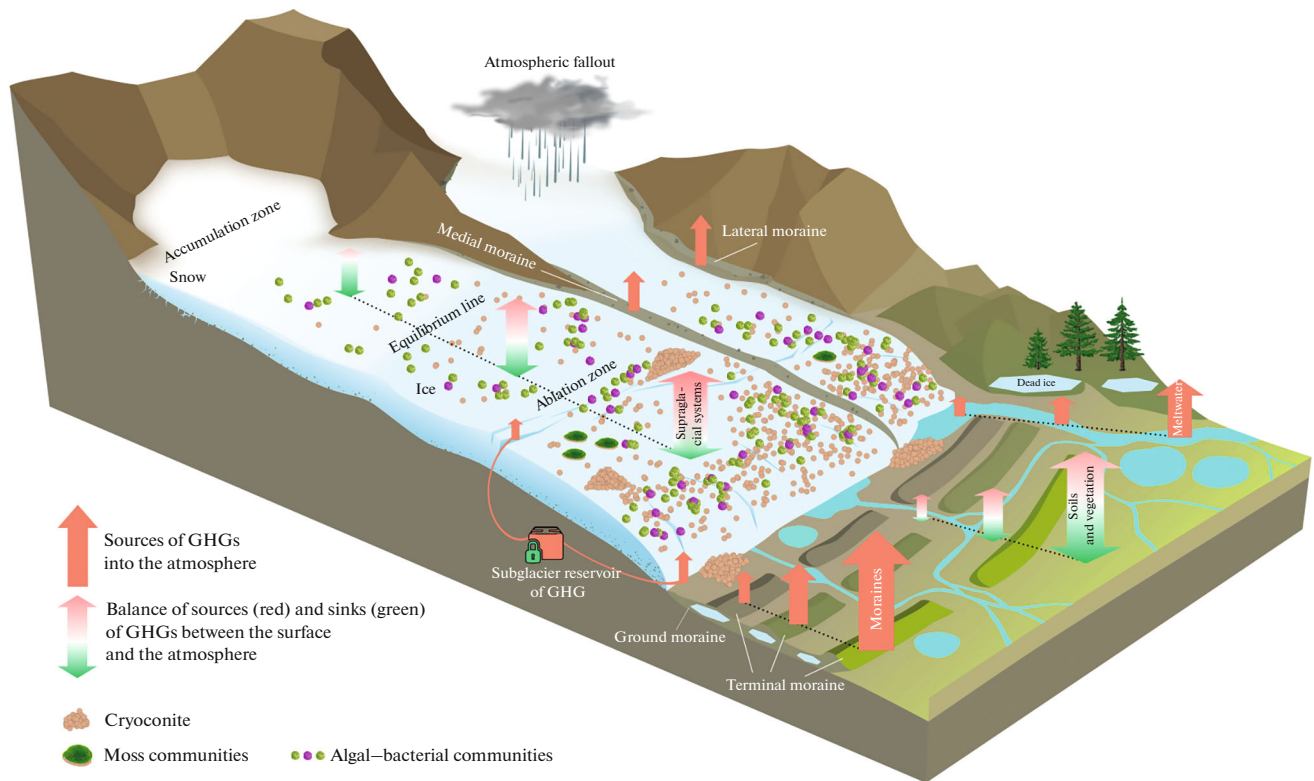


Fig. 9. Glacial biome and greenhouse gas (GHG) fluxes in the supraglacial and near periglacial (proglacial) zones.

holes. Such cryoconite holes retain nutrients longer, and their primary production is higher. Closer to the ablation zone, where there is a powerful washout with meltwater, the mass of cryoconite and microbiota decreases, and the net carbon balance also decreases. The zone of active melting gradually moves upward during the warm season. Thus, one cannot simply extrapolate data from small (mountain-valley) glaciers to giant ice sheets.

The effervescence of cryoconite as a result of bubble transport is noted as an important route for the entry of CO_2 and methane into the aboveground air from melting ice, which can abruptly and synchronously increase the background concentration of CH_4 by 20 ppb and of CO_2 by 5 ppm over time intervals of about 10–15 min [259]. As a result, up to 79% of methane emission from glacial meltwater is due to bubble transport [262].

The first (and so far the only) attempt to estimate the global contribution of the glacier surface to the biosphere net CO_2 balance was undertaken by the authors [36]. Extrapolating the available data and based on the predominance of autotrophy in cryoconite holes, the authors estimated the global cryoconite sink size at about -0.064 million t C per year for the surface of glaciers outside Antarctica. Comparing again with the above estimate for the Russian tundra zone (sink of about 11.6 million tons C per year), one

can see how insignificant the first value is. However, average specific area estimates of cryoconite production and respiration are 3–4 orders of magnitude higher than those for the total surface of glaciers and are comparable to emissions from Arctic lakes. Using these data, the total contribution of cryoconite to the net exchange of CO_2 between the surface of glaciers and the atmosphere can be minimally estimated at 77%.

Such extrapolations relate exclusively to ice melting seasons, while the longest period, winter, is actually ignored, which can significantly underestimate ecosystem respiration and overestimate the net annual CO_2 sink. In addition, almost all the data relate to the ice sheets of Greenland and Antarctica, as well as to large glaciers of Svalbard. Currently, active research on GHG exchange has begun on the Tibetan Plateau, the largest glacial massif on land after Antarctica and Greenland [259]. The Russian territory of Eurasia remains virtually unstudied in this aspect, except for a recent work in the periglacial zone of the largest ice cap on the Novaya Zemlya Archipelago [20] and the first estimates of the balance of greenhouse gases on the glaciers of the Caucasus and Altai [14].

However, the role of glaciers as biospheric agents is determined not by the lifetime value of CO_2 balance on their surface (which largely depends on the presence of cryoconite), but by the conservation properties of great ice masses, as well as by their high albedo. In

this context, it is more important to study the balance of CO₂ and other biogenic GHGs in the areas of melting glaciers and in places of recent glacier collapses and complete destruction.

Apart from studying periglacial GHG fluxes on a local scale, studies covering the *glacial landscape as a whole*, including the contribution of abiotic processes, are insufficient. One of the pioneering works here is the work [239] on the Western Tien Shan, which showed that the entire glacial basin during the melting season turned out to be a small but significant CO₂ sink of $-0.05 \mu\text{mol}/(\text{m}^2 \text{ s})$ over pure ice and $-0.07 \mu\text{mol}/(\text{m}^2 \text{ s})$ over young moraines with an ice base. It is important that the main mechanism in this case is the abiotic absorption of CO₂ by meltwater with the formation of carbonic acid. Thus, the “chemical pump” mechanism of CO₂ sequestration from the atmosphere replaces the biological pump operating through photosynthesis. At the same time, strong fluctuations in the CO₂ sink were noted at an hourly measurement step, which is associated with the abrupt release of gases from bubbles contained in melting ice [189]. In principle, such a mechanism should lead to an increase in the emission of GHGs such as CO₂ into the atmosphere, which contradicts the results demonstrating its preferential absorption. However, the air in the bubbles formed during melting has a concentration of CO₂ lower than in the modern atmosphere, i.e., ~295 ppm, which is typical for the pre-technological era. Apparently, the release of these additional molecules is compensated by their chemical binding on a landscape and seasonal scale. At the same time, 75% of the C from melted glaciers is carried out in dissolved form into the ocean, in the form of HCO₃⁻, and only 25% in gaseous form immediately enters the atmosphere [189].

This fact indicates an active research direction that has emerged over the past decade regarding *the emission of CO₂ and methane from glacial meltwater, periglacial formations, and subglacial sediments*, which includes the glacial landscape as a whole, together with terrestrial and aquatic biomes bordering glaciers, into the research orbit. The exchange of the Greenland Ice Sheet has been studied in most detail in this regard [35, 56, 134, 189], and a detailed study of the Tibetan Plateau has begun [259]. All rivers and lakes emit respiratory CO₂ into the atmosphere, but in glacial rivers and lakes the contribution of chemical weathering with the participation of carbonates and silicates becomes important, during which CO₂ transforms into the form of the bicarbonate anion HCO₃⁻, which can be considered a carbon sink if we use the characteristic lifetimes of the bicarbonate anion in land waters (10²–10³ years) or in the ocean (10⁵ years) [189]. Multiple analyses of ocean water samples show that with distance from the Greenland Ice Sheet, the concentration of dissolved CO₂ gradually decreases and the con-

centration of dissolved HCO₃⁻ increases; then, as equilibrium with the atmosphere is established (approximately 20 km from the glacier), the C content in these forms ceases to change. According to preliminary estimates, 50% of this is CO₂ contained in surface water flowing from the glacier, and the rest is released from under the ice sheet itself. The water coming from under the Greenland Ice Sheet is very saturated with CO₂, but at the same time this gas is released into the atmosphere in very small quantities—about 0.11 million tons C per year. However, models show that under the most intense climate scenarios, upon reaching the lower layers of ice saturated with CO₂ during melting, emissions could reach 100–170 million tons of C(CO₂) per year by 2100 [189], which is comparable, for example, to modern emissions from forest fires in Russia.

Concentrations of CO₂ and methane in glacial meltwater can differ by orders of magnitude between sections of the Greenland Ice Sheet, and the waters themselves can be either oversaturated or undersaturated with these gases [165]. In particular, significant differences in the magnitude of subglacial sources of CO₂, with relative equality of rates of atmospheric runoff lead to the fact that local meltwaters are either net sources or sinks of atmospheric CO₂. These results highlight a previously unrecognized degree of heterogeneity in greenhouse gas dynamics beneath the ice sheet. At the same time, glacial waters always remain sources of methane [259]. In general, methane contained in the melted glacial waters of the Earth is a potential source of about 1 million tons per year into the atmosphere, with its reserves under glaciers in the form of gas hydrates up to 480 billion tons C—CH₄, and the total supply of carbon in all forms with ablation is at least 6 million tons of C per year [238].

The latest large-scale studies have now emerged assessing the balance of C-containing GHGs in all major components of the glacial landscape of the Tibetan Plateau, including supraglacial surfaces, cryoconite holes, periglacial sediments, and periglacial rivers and lakes [259], which show a significant increase in the emission of methane and CO₂ from moraine sediments and glacial meltwater as a result of climate warming. The spread of stagnant moisture, lakes and swamps on the Tibetan plateau as a result of the progressive retreat of glaciers increased the emission of CH₄ to 0.96 million tons per year in the 2000s, which, according to independent estimates, exceeded the rate of its absorption in the same area (0.74 million tons per year) [242]. Despite the fact that cryoconite holes in this landscape can be both net sinks and sources of CO₂, there is no doubt that they act as net sources of methane [259].

Among the three most important biogenic GHGs, the least number of studies in the glacial and periglacial zones are devoted to N₂O, which is explained by

the insufficient size of its fluxes for instrumental assessment. Significant N₂O emissions are recorded only on moraine deposits, when vegetation appears on them [226], or under the influence of guano in bird colonies [240]. In this case, important factors in the balance between gas production and absorption are pH and the degree of soil aeration; a significant contribution of abiogenic nitrous oxide production is also possible.

Continued climate warming dictates the need to combine all components of the glacial biome (supraglacial surfaces, subglacial sediments, moraines, and meltwaters) in predictive models of biosphere dynamics and take into account significant positive feedbacks. At the same time, it is important to include the biotic component of the GHG balance of glaciers in biosphere models. There is a lack of data on the GHG exchange of ice sheets and mountain glaciers in the territory of Russia, which still represents a blank spot in this regard. The need for such studies is due to the fact that the indicators of GHG production, absorption, and transport can significantly depend on local features of the structure of glaciers, climate, anthropogenic influence, and previous geological history.

CONCLUSIONS

Supraglacial organomineral formations have the most diverse composition and the highest rate of biological turnover among all components of the glacial biome. During the period of ablation on the glacier there is no shortage of liquid water, organic and mineral impurities, the interaction of which leads to the emergence of organomineral bodies with new properties that change the structure of the biological community, the chemical composition of the substrate and the spectral characteristics of the glacier surface, as no one else would have done from components separately (emergent properties of the system). Analysis of the diversity, structure, and formation processes of supraglacial organomineral bodies showed that some of them have features related to soils and soil-like bodies. The most important of these features are the accumulation of OM, its stabilization on mineral surfaces and in aggregates, the transformation of OM with the residual accumulation of dark-colored solid-phase products in situ. Among supraglacial formations, it is proposed to distinguish pre-soils, including ephemeral pre-soils, and soil-like bodies in ice and snow; metastable soil-like bodies on cryoconite; soils with microprofiles under moss communities on ice; as well as relatively stable soils with macroprofiles on gravelly fine earth sediments underlain by glaciers and dead glacier ice. Easily available OM produced in the supraglacial “reactor” and its “cells”—soils and soil-like bodies—has a significant impact on the periglacial zone, leading to reservoir and priming effects and affecting all links of the trophic chain, as well as the formation of the initial soil cover after glaciers retreat.

The supraglacial zone is an autotrophic–heterotrophic ecosystem. Its organomineral components, including soils and soil-like bodies, under different conditions are both net sinks and sources of greenhouse gases. However, in the modern “melting” world, the biospheric significance of glaciers is determined not only by the lifetime value of the balance of greenhouse gases on their surface, but also by the conservation properties of the ice masses, as well as their albedo. Therefore, the balance of greenhouse gases must be studied in all components of the glacial landscape both during the melting of glaciers and after their retreat.

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CONFLICT OF INTEREST

The authors of this work declare that they have no conflicts of interest.

CONSENT TO PARTICIPATE

Informed consent was obtained from all individual participants included in the study.

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