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SOIL CHEMISTRY

Assessment of the Mobility and Time of Renewal of the Densimetric Fractions of Organic Matter in Chestnut Soils from the Ratio of Stable Carbon Isotopes

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Abstract—The results of experimental studying of the organic matter status in soils using methods of granulodensimetric fractionation and the geochemistry of the stable carbon and nitrogen isotopes are presented. The organic and organomineral matter in chestnut soils representing a chronoseries—stages of natural changes in the vegetation with different carbon isotope compositions—were investigated. The organomineral fractions and fractions of free organic matter were distinguished, the changes in their proportion due to the changes in the vegetation were analyzed, and the residence time of the carbon in the organic matter of the fractions was assessed.

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INTRODUCTION

One of the most important functions of the soil cover—the regulation of the biogeochemical fluxes in the biosphere—is implemented as the carbon exchange between the soil and atmosphere. In the context of traditional pedological approaches to the investigation of this interesting and significant biosphere function of the soil, one can distinguish between studies of the biological cycle and studies of soil organic compounds. The first line of research is related to the determination of the biogeochemical structure of the carbon cycle in the biosphere (the sources, carbon reserves, and fluxes), while the second one is related to the characterization of the carbon pool (according to the modern terminology) and the carbon pool in the soil. When assessing the soil organic matter, great attention is traditionally paid to both the carbon compounds and the properties of the environment where they are synthesized, transformed, and function. This approach allows assessing the originality (uniqueness) of the soil carbon pool. Despite the fact that the carbon content (1.6×10^{12} t) in soils is much lower than its contents in the earth's sedimentary sphere (81×10^{15} t) and in the World Ocean (40.6×10^{12} t) [1], the soil carbon pool appears to be more active in the present-day biosphere than the carbon reserves in the sedimentary cover. This carbon pool is more stable or conservative than its pool in the World Ocean. The specific features of the soil carbon pool are determined by its openness (the spatial boundary position in the biosphere and the specific internal properties, namely, the substance and structural heterogeneity of the environment (where the carbon compounds are formed and exist)) and by the thermodynamic stability of the soil carbon compounds proper in the present-day environment. At the same time, all this complicates the problem of studying the soil carbon pool, which is especially topical in the recent time, when the technogenic carbon fluxes in the biosphere, as many scientists state, can affect the global climatic processes. The results of the technogenic or natural changes in the cycle components and the imbalance of the carbon content in the atmosphere have already resulted in some climatic trends during the last several decades in many regions of the planet, thus confirming the global character of this phenomenon. These circumstances have caused a severe need to regulate the carbon cycle in the biosphere, and, in our opinion, the soil carbon pool is a key object and relatively available for control (except for the reduction of the technogenic CO_2 emissions). In order to control the carbon cycle, it is necessary to assess its potentialities and the processes and mechanisms responsible for its formation in a volume sufficient to manage the carbon fixation and preservation.

Presently, the factual information on the soil carbon pool behavior is rather limited not only due to the fact that its biosphere importance was recognized in full measure not so long ago but, primarily, due to the objective methodological difficulties. From the standpoint of obtaining adequate and uniquely interpreted data, the principles of the distinguishing and further division of the soil organic pool components are of the greatest importance and need discussion.

The aim of this work was to reveal some specific features of the soil carbon pool—its activity assessed by the residence time of carbon in the soil organic matter and the structure and mechanisms responsible for its fixation in the form of organic compounds. This complicated task was suggested to be accomplished owing to the combination of the experimental possibilities provided by the environmental conditions of the region and the modern instrumental methods, namely, the following:

—The studied object is a unique one, since it allows constructing a chronological series of soils with increasing time of entering of carbon into C3 plants that differ in their carbon isotope composition from the initial C4 plants.

—The mass-spectrometric method of determining the composition of the stable carbon isotopes of organic matter in the soils permits one to detect the contents of carbon entering from the vegetation of the C3 and C4 photosynthesis types.

—The system of isolation of the organic matter fractions reflecting the carbon's native state in the soil; these fractions are supposed to be independent to some extent in their genetic and functional aspects.

OBJECTS AND METHODS

The field works were conducted at the experimental station of Texas A&M University in the eastern part of the Rio-Grande River basin covering the southern area of the USA (the State of Texas) and northern Mexico. The climate of the region is subtropical with a mean annual temperature of 22°C, the mean annual precipitation is 715 mm, and the evapotranspiration is 1540 mm. For a sufficiently short time period (nearly 120 years), in this territory, the native herbaceous vegetation has been substituted for woody vegetation, which reflects the worldwide trend observed in the last 100–120 years in the steppe and savanna of Europe, North and South America, Africa, Asia, and Australia. For 40 years, since 1950, in the territory studied, the area occupied by woody vegetation has increased by 2 times. In the woody communities, nitrogen-fixing woody leguminous species such as *Prosopis glandulosa* and some species of *Acacia* predominate.

For this research, in the territory studied, the difference in the ratio between the stable ^{13}C and ^{12}C isotopes in the herbaceous and woody plants that substitute for the former is very important. In this connection, we briefly consider the main principles of the formation of the isotope composition in plants; a detailed description of this process was performed earlier [2]. The isotope composition of vegetation depends on the biochemical features of the pathways of the CO_2 assimilated by the plants during photosynthesis. From this standpoint, terrestrial plants can be divided into three main photosynthetic types; each of them has its own unique particularities of carbon isotope fractionation. In the temperate zone, the predominant part of the terrestrial vegetation, including the majority of trees, shrubs, and grasses, uses Calvin's cycle in the atmospheric carbon fixation. They are called C3 plants. There are less widespread plants with another type of carbon fixation in photosynthesis; they use the reaction cycle of Hatch–Slack and are called C4 plants. This type includes many plants of desert and semidesert regions; among cultivated plants are such as maize (*Zea mays*), sugar cane (*Saccharum officinarum*), and sorgho (*Sorghum*). The portion of C4 plants in the floristic composition of the plant communities increases with a reduction in the geographic latitude and the absolute altitude of the sites [7, 12]. C4 plants are more competitive under arid conditions since they assimilate CO_2 more intensely. Lastly, there is one more exotic and poorly investigated type (from the standpoint of the isotope fractionation type), CAM plants, including cactuses and succulents.

Usually, for the expression of the isotope composition, the δ value representing the decline of the proportion of the isotopes studied in the sample compared to the proportion in the substance accepted as the standard (in δ or, more rarely, in %) is used:

$$\delta^{y}X,\ \text{‰} = \frac{(^{y}X/^{z}X)_{\text{sample}} - (^{y}X/^{z}X)_{\text{stand}}}{(^{y}X/^{z}X)_{\text{stand}}}1000,$$

where $^{y}X_{\text{sample}}$ and $^{y}X_{\text{stand}}$ are the values of the intensity of the ion fluxes (volts) of the less spread (heavier) isotope of the chemical element X with mass y in the sample and the standard, respectively, upon the condition that the signals from the more spread (light) ^{z}X isotope are equal. Evidently, the δ value depends on the standard chosen. For measuring the correlation between the isotopes of the chemical elements, a component of the earth's crust containing the main mass of the element investigated or a natural compound that preserves unchanged the correlation between the isotopes of this element from the early stages of the earth's formation and the termination of the nucleus synthesis is usually assumed as a standard. For instance, the ^{15}N to ^{14}N ratio for atmospheric nitrogen is the standard and the reference point for the beginning of the fractionation of all the rest of the nitrogen compounds. As for carbon, its isotope composition is considered in reference to the Pee Dee Belemnite (PDB) standard—calcite of the fossil *Belimnitella Americana* of the Cretaceous age. This substance has been spent by this time, but its derivatives are available. Therefore, there exists the possibility to express the analytical results in reference to the PDB standard. This standard is traditionally used in the international system to express the data on the isotope composition of carbon. The δ value for

Table 1. Properties of the soils studied

Depth, cm	pH_{H_2O}	C_{org}, %	CEC, meq/100 g	Adsorbed Na, % of the CEC	$CaCO_3$ equiv., %	Extracted bases, meq/100 g			Particle-size fractions, %		
						Ca^{2+}	Mg^{2+}	Na^+	sand 2–0.05 mm	silt 0.05–0.002 mm	clay <0.002 mm
0–15	7.5	0.66	13.9	1	0	10.1	1.2	0.2	76.0	13.8	10.2
15–30	7.4	0.54	11.4	2	0	9.6	1.3	0.2	76.6	11.2	12.2
30–60	7.4	0.46	14.4	1	0	12.9	2.0	0.1	70.0	12.2	17.8
60–90	7.7	0.32	15.8	1	0	18.2	2.4	0.2	64.4	16.5	19.1
90–120	8.1	0.42	12.9	2	15.2	48.1	2.3	0.2	54.5	22.4	23.1
120–150	8.1	0.33	10.4	2	23.3	46.2	2.5	0.2	50.6	24.1	24.7

the standard sample is zero. For the expression of the data for other compounds different from the standard, the "–" sign indicates that the sample studied is depleted in light isotopes as compared to the standard. In this case, the isotope composition is said to be *lighter*. On the contrary, the "+" sign means that the sample is enriched with a less spread heavier isotope; i.e., this sample becomes *heavier* as compared to the standard.

The isotope composition of the carbon in atmospheric CO_2 is about –8‰. In the course of photosynthesis, the carbon isotopes function so that the plant organic mass is enriched with the light carbon isotope ^{12}C. The degree of this enrichment is related to the type of photosynthesis. The biomass of terrestrial plants of the C3 photosynthesis type is characterized by a range of $\delta^{13}C$ from –22 to –32‰ (the average is ~–27‰), while the biomass of C4 plants varies from –10 to –18‰ (the average is ≈–13‰).

Thus, the isotope composition of the carbon entering the soil with the organic matter depends on the type of the vegetation and the environmental conditions. The portion of C4 (or C3 plants) in the formation of the isotope composition of the soil organic matter can be assessed according to the formula representing the isotope–material balance:

$$C4, \% = \frac{\delta^{13}C_{init} - \delta^{13}C_{C3}}{\delta^{13}C_{C4} - \delta^{13}C_{C3}}100.$$

where $\delta^{13}C4 = -13‰$ (the average for C4 plants), $\delta^{13}C_{init}$ is the humus carbon isotope composition measured, $\delta^{13}C_{C3}$ is the isotope composition of the C3 plants (–27‰), and C4 is the percentage of C4 plants in the composition of the plant community.

Nowadays, in the studied territory, the plant community can be represented as a two-phase system. The indigenous phase is represented by herbaceous plants with the domination of C4 species ($\delta^{13}C$ ~–13‰) continuously distributed over the territory, and the other phase substituting for the indigenous one is represented by groups of C3 plants (trees) ($\delta^{13}C \approx -27‰$) included into the herbaceous biocenosis; it has a discrete distribution. Thus, in this case, the present-day plant cover is characterized by the absence of equilibrium between its two main components that differ in the phytomass structure, as well as in the structure of the plant material, primarily, in its carbon isotope composition.

In the soil cover, solonetzic chestnut soils (according to the USA soil taxonomy, Typic and Pachic Argiustolls) predominate; their analytic characteristics are given in Table 1. Gypsum was not found along the whole soil profiles.

For the analysis of the carbon dynamics in the soil, a chronocatena consisting of plots occupied by woody vegetation during different time periods was selected [10]. The age of the wood biocenoses and, consequently, the duration of the carbon input with C3 plants to the soils were detected according to the annual rings of the oldest tree chosen in the group. Thus, in the territory investigated, a series of biocenoses was formed starting with those where woody species were absent (the 0 moment) to the 80-year-old biocenosis; the intermediate biocenoses were at ages of 14, 23, 29, 32, 35, 40, 52, and 68 years. In each sample plot, at a distance of 0.5 m from the stem of the largest and oldest tree, soil samples were collected (4 replicates) using an auger. A mixed sample was prepared. A sampling plot of the herbaceous biocenosis was chosen at least at a distance of 10 m from the closest groups of trees. Soil samples were taken at the same sites. In addition, from an area of 0.5 × 0.5 m around the sites where the soil samples were taken, the plant falloff and live leaves were collected. The soil and plant samples were dried at 60°C; the soil samples were fractionated by particle size and density using a heavy liquid (sodium polytungstate, $3Na_2WO_4 \cdot 9WO_3 \cdot H_2O$). The fractionation was performed using the method of Shaimukhametov et al. [6]. Some modifications of this methodology were made by the author in order to use sodium polytungstate instead of bromoform. The roots were removed from the soil after its wetting and sifting through a 0.25-mm sieve. In the fractions obtained, the roots and plant material, the C and N

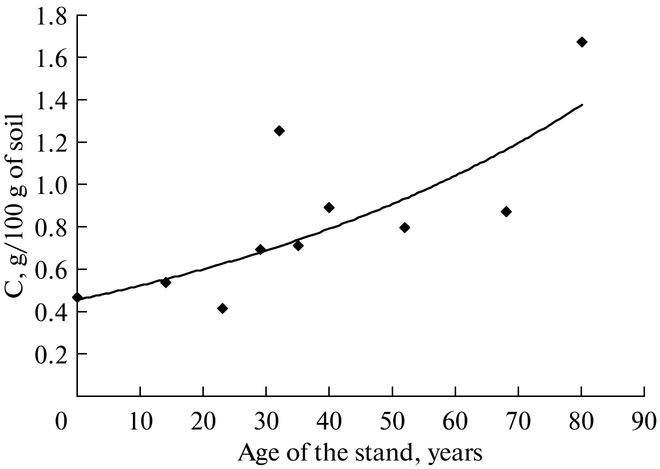

Fig. 1. Changes in the carbon content of the organic matter in the 0- to 15-cm soil layer as related to the age of the stand.

contents, and their isotope composition were determined. The measurements were carried out using an element analyzer (EA1108, Carlo Erba, Lakewood, NJ, USA) connected to an isotope mass spectrometer (Delta Plus, ThermoFinnigan, Germany) working under a continuous flow of helium (the carrier gas). The accuracy of the measurements for δ^{13}C is < 0.1‰; the values are given in reference to the PDB standard.

RESULTS AND DISCUSSION

The correlation between the fractions, carbon mass, and the concentration in the soil fractions. The analysis for the carbon content in the soils under the tree stands of different ages demonstrates a trend of increasing carbon concentration during 80 years (Fig. 1). Below is the consideration of the structure of the soil organic carbon pool and its transformation as the trees reclaim the territory by comparing the two most contrasting variants—the soil under the herbaceous biocenosis and the soil under the oldest forest. About 86% of the soil mass (Table 2) is represented by the heavy organomineral fraction with a bulk density of >2 g/cm^3. Evidently, its content is mainly controlled by the lithological factors—for instance, by the particle-size composition of the parent rocks. Judging from the permanently low content of the fine component (<2 μm) in the heavy fraction (>2 g/cm^3), the loamy–sandy composition does not favor the relatively stable microaggregation of the soil. In connection with the changes of the biocenoses, the contents of the coarse (2–0.2 μm) and fine (<0.2 μm) clay in the weakly stable microaggregates are variable. The fine fraction (<2.0 g/cm^3) containing free organic matter unbounded with clay minerals changes to the greatest extent. In the soil of the herbaceous biocenosis, the content of this fraction is minimal; it increases by several times as an area is forested. In the composition of this fraction, the amount of readily decomposable organic matter (the fraction >45 μm) representing the organic residues of plant, animal, and microbial origin increases to a greater extent [4]. In the forest biocenosis, not only the weakly decomposed material accumulates but also its humification is activated, as evidenced by the increase in the content of thermodynamically stable organic matter (the fraction <45 μm).

The carbon content in the soil of the woody biocenosis increases in all the fractions as compared to that in the soil under grass (Table 2). Its concentration is maximal in the light fraction. Judging by the similar trends in the changes of the C to N ratio (Table 2), the

Table 2. Proportion between the fractions in the soils of the forest and herbaceous biocenoses

Biocenosis	Fractions (% of the soil mass)						
	2–0.2 μm	<0.2 μm	>2.0 g/cm^3		<1.8 g/cm^3		1.8–2.0 g/cm^3
			<2 μm	>2 μm	>45 μm	<45 μm	
Herbaceous	5.7	5.6	1.1	84.6	0.1	0.4	0.2
80-year-old forest	5.1	3.4	1.1	84.3	2.6	1.1	0.7
Carbon content in the fractions (% of the fraction mass)							
Herbaceous	3.1	1.5	3.3	0.02	27.1	28.8	7.7
80-year-old forest	5.3	4.4	3.7	0.04	31.6	31.2	8.1
C/N ratio in the soil fractions							
Herbaceous	9	8	9	4	18	17	10
80-year-old forest	8	8	9	4	16	14	12
Soil carbon bound in the soil fractions (carbon mass, % of its total content in the soil)							
Herbaceous	36.6	17.3	7.7	3.8	5.6	25.3	3.7
80-year-old forest	15.8	8.7	2.5	2.0	48.3	19.2	3.5

character of the plant material humification is identical in both cenoses. As was assumed, in the organic matter bounded with clay minerals, humic acids relatively enriched with nitrogen predominate, although to a lesser degree than in the free organic matter [3]. According to the results of Titova et al. [3], the C/N ratio in the fractions of the organic matter sorbed on the mineral material is low and well agrees with our materials. The free organic matter of the light fraction is relatively enriched in nitrogen, and its fraction <45 μm, as was shown experimentally [3, 5], has a higher humus content than the sorbed carbon compounds.

The assessment of the contributions of each fraction to the carbon pool of the soil in the herbaceous biocenosis demonstrates the leading role of the organo–clay complexes and free organic matter, although the total mass of these fractions is small (Table 2). The correlation between the components of the carbon pool drastically changes as a result of the development of the forest biocenosis during 80 years. The light fraction of the organic compounds (<1.8 g/cm^3) fixing almost 70% of the soil carbon becomes dominant, and the readily decomposable organic matter turns out to be significant in it.

The heavy fraction (>2 g/cm^3) exerts the lowest effect on the soil carbon pool in the herbaceous and forest biocenoses. At the same time, this fraction comprises the main part (85%) of the soil mineral substance, which confirms its dependence on the lithological factors.

The isotope composition of the organic matter carbon in the soil fractions. The isotope composition of the terrestrial above-ground and root biomasses—a carbon source of the soil organic matter—varies significantly due to the fact that, in the territory studied, plants of two photosynthesis types grow. For the above-ground biomasses, the isotope composition ranges from –27.4 to –19.2‰; that of the root mass varies from –26.6 to –21.2‰. In the both cases, the lower value characterizes the woody C3 plants, while the higher one is characteristic of the herbaceous plants of the C4 type [8, 9]. For the organic matter in the upper part of the humus horizon, the same authors give ranges from –21.0 to –17.4‰. Thus, under the general trend of the changes toward the lighter isotope composition of the organic carbon due to the plant succession in the studied savanna area, the soil is a more conservative component of the biogeocenosis, which preserves the information on the last predominance of C4 plants to the greatest extent. As for the isotope organic matter composition of the fractions, it can be controlled by two factors. The first evident factor actual for the whole soil mass is the variable isotope composition of the carbon source—the biomass of the plants, the photosynthesis type of which changes from C4 to C3. This phenomenon is accompanied by the *lighter* character of the organic matter carbon entering the soil. The second factor, which is not clear and needs further investigation, is the possible fractionation of the carbon isotopes due to the specific soil processes of the humification. The experimental data on the carbon isotope composition of the soil fractions demonstrate its regular changes related to the age of the stand, i.e., to the time during which the organic matter enters from the new C3 source (woody vegetation) different in the isotope composition from the former vegetation with the predominance of C4 plants (Fig. 2). As stated above, in this case, the carbon isotope composition of the fractions changes due to the replacement of the vegetation and the fractionation of the carbon isotopes as a result of the soil processes.

The plant succession is clearly manifested in the changes of the carbon isotope composition of the different fractions (Fig. 2). The increase in the content of “light” carbon in the fractions of the soil organic matter by 4–6‰ mainly takes place due to the input of carbon with the C3 plants (woody plants). In addition, one can note the significant differences in the carbon isotope composition between some fractions. In the herbaceous and forest biocenoses, they amount to 7 and 4‰, respectively. Probably, these differences in the isotope composition of the carbon fractions are explained by several reasons. The first reason is the different rates of the carbon cycle for the separate fractions. In the case considered, it is displayed in the isotope records due to the changes in the vegetation. Another reason for the differences is the humification processes, which may fractionate carbon isotopes. Unlike the previous process, this one is a soil process governed by the biochemical regularities of the humification. Lastly, the rest of the process affecting the proportion of the carbon isotopes in the different fractions is the transfer of some fractions between the biocenoses. This is especially notable in the isotope composition of the light fraction (<1.8 g/cm^3 > 45 μm). Precisely this fact is responsible for the significant dispersion of the data on the carbon isotope composition between the fractions in the herbaceous biocenosis. Here, the isotope composition of the fraction <1.8 g/cm^3 > 45 μm is much lighter than that of the roots and the other less mobile fractions due to the fact that this fraction appears to enter partly from the woody C3 biocenoses with the lighter carbon composition of the plant material. We can add that the content of the fraction <1.8 g/cm^3 > 45 μm in the forest biocenosis exceeds its content in the herbaceous biocenosis by more than 20 times. In other words, the forest biocenosis is a rather powerful source of this fraction. It is not inconceivable that the carbon of this fraction also has some qualitative and quantitative effect on the other fractions in the grass and adjacent uneven-aged forest biocenoses. This is manifested in both the carbon content and its isotope composition.

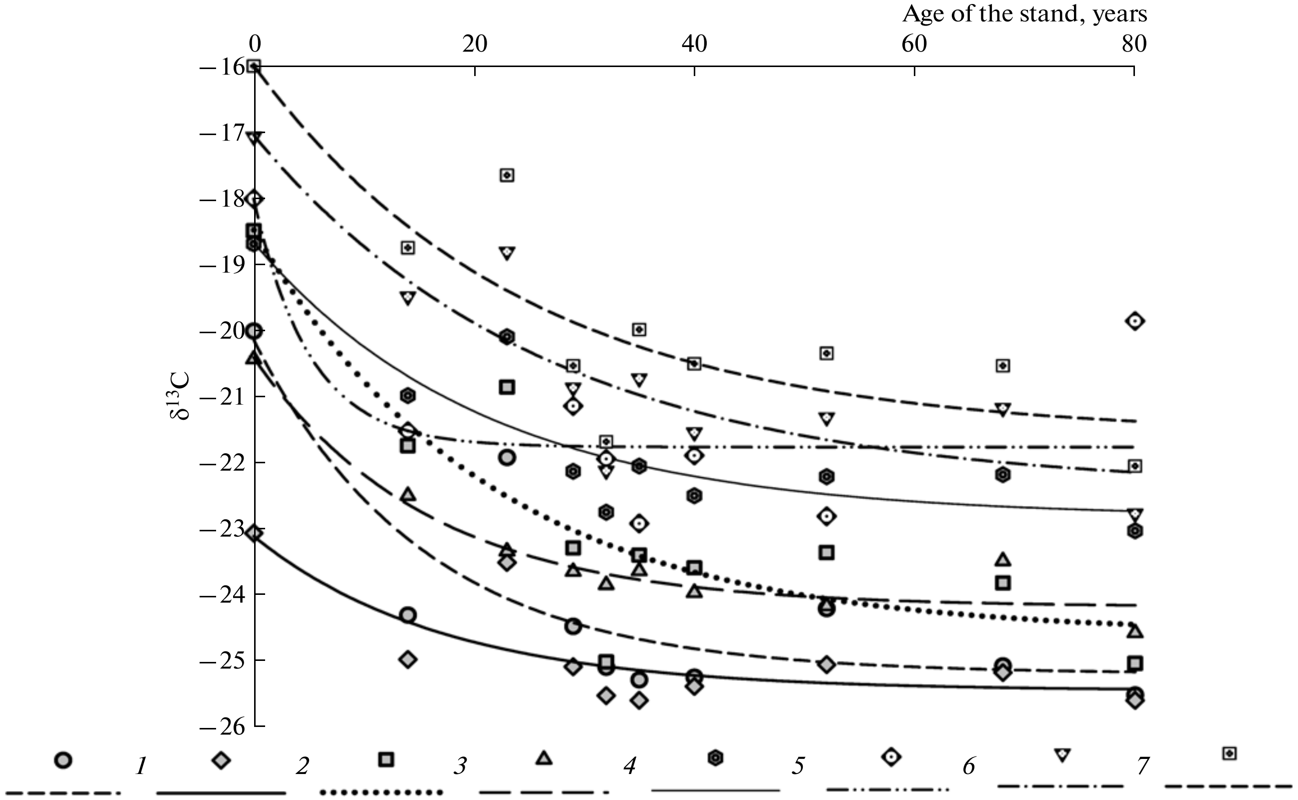

Fig. 2. Changes in the isotope composition of the carbon in the fractions as related to the age of the stands. Hereinafter:

Designation	Fraction
1	roots
2	<1.8 g/cm^3 >45 μm
3	<1.8 g/cm^3 <45 μm
4	1.8–2.0 g/cm^3
5	>2 g/cm^3 <2 μm
6	>2 g/cm^3 >2 μm
7	<0.2 μm
8	0.2–2 μm

However, the effect of this fraction is assessed as low because of the asynchronous changes in the isotope composition of all fractions with the age of the stands. The single exception pointing to the same fact is the change in the isotope composition of the organic carbon in the <1.8 g/cm^3 < 45 μm fraction. In this case, in the herbaceous biocenosis (the C4 plants are relatively *heavy* by their carbon), the carbon isotope composition of the free organic matter is enriched with a heavier carbon isotope to a greater extent than that of the uneven-aged forest C3 plants (relatively poor in the light isotope).

One of the tasks of our studies was to assess the time of carbon residence in the different structural components of the carbon pool. In this connection, for each member of the chronoseries, the correlation between the carbon contents entering from the different sources in the course of the plant cover replacement was determined. The calculation was performed using the equations of the material–isotope budget given above. The portion of carbon remaining from the herbaceous C4 biocenosis in each soil fraction was compared with the duration of the tree growth (the C3 biocenosis). Figure 3 shows that this portion, as expected, decreased gradually. However, an equation should be obtained to describe this process quantitatively.

For this purpose, the known equation of the first order describing the decomposition of the plant residues was used [11]:

$$A_t = A_0 e^{-kt},$$

where A_t is the remaining carbon content in the herbaceous C4 biocenosis at the moment t (year), A_0 is the carbon content in the grass biocenosis at the initial moment ($t = 0$), and k is the index of the decreasing of the initial carbon content in the grass C4 biocenosis per year (percent/yr).

Using the equation given, the experimental data were approximated to produce the curves that describe the decrease in the carbon portion in each fraction of C4 plants per year and present the values of the relative rate of its changes (k). Finding the $1/k$ value, we obtain the average residence time of the carbon in the fractions (expressed in years). These results are given in Table 3. They show that the mean residence time for the carbon in the free organic matter was twice less than in the sorbed organomineral compounds. This means that the free organic matter is the most active part of the biological cycle. The indication of the importance of these fractions consists in their relative high capability to reflect the bioclimatic conditions of the soil formation and the changes in the biocenoses. In addition, the fractions of free organic matter and, primarily, fraction <1.8 g/cm^3 < 45 μm containing specific thermodynamically stable soil organic matter can adequately indicate the paleoclimatic conditions when studying buried soils. Judging by the carbon residence time, the sorbed organic matter is more conservative in the biological cycle and, probably, contains less information. However, the investigations of the properties of the sorbed organic matter remain topical, since evidence indicates that the compounds forming in the course of humification have their own specific features [4]. This fact is important, since it enables one to study the processes and mechanisms of the humus formation using isotope methods and physical fractionation. The comparison of the authors' data on the carbon residence time in the fractions with the mean values determined for the total organic matter in the soils by the radiocarbon method shows that these results are quite comparable—the data of the radiocarbon analysis for the upper horizons of some soils gave estimates from 34 to 81 years [9].

Table 3. Carbon residence time (years) in the soil fractions

Fraction	Residence time
<1.8 g/cm^3 >45 μm	30
<1.8 g/cm^3 <45 μm	35
1.8–2.0 g/cm^3	40
>2 g/cm^3 <2 μm	65
>2 g/cm^3 >2 μm	65
<0.2 μm	70
2–0.2 μm	75

The dynamics of the nitrogen isotope composition demonstrate unique and unambiguously interpreted results. The comparison of the data on the mean carbon residence time in the fractions shows that the increasing time leads to an increase in the *heavier* nitrogen in the organic matter (Fig. 4). Evidently, this is related to the fact that, in the course of the biogeochemical transformation of the nitrogen compounds, the content of the *heavier* nitrogen in the organic matter grows due to the emission of lighter

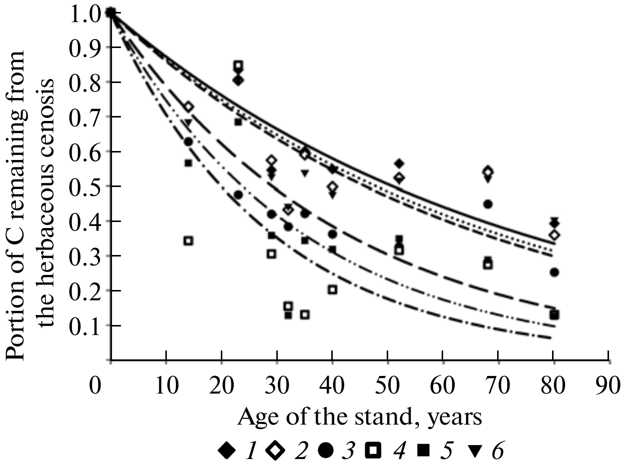

Fig. 3. Proportion of carbon in the soil organic matter entering from the different biocenoses.

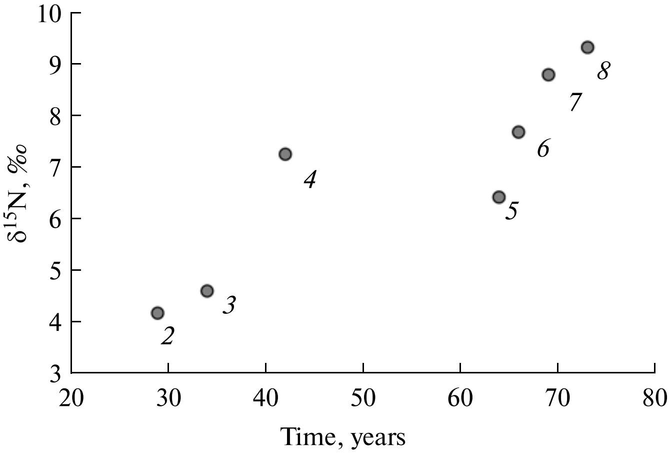

Fig. 4. Changes in the nitrogen isotope composition as related to the carbon residence time in the organic matter.

isotopes with volatile compounds as a result of the kinetic effect of the isotope fractionation.

CONCLUSIONS

The results obtained demonstrate the unique possibilities of using the methods of stable carbon isotope geochemistry in investigating the biological carbon cycle. Under certain biocenotic conditions, this method permits one to divide and quantify the carbon sources for soil organic matter. The possibilities of this method enlarge to a great extent as the granulodensimetric fractionation of the soil material is applied. The latter allows dividing the organic matter according to its activity in the biological carbon cycle and helps, using the data of the isotope analysis, to determine the structure of the soil carbon pool and its functional possibilities. The latter procedure is especially important.

For the upper part of the humus horizon (0–15 cm) of the light-textured solonetzic chestnut soil with the content of organic carbon of 0.7%, the carbon residence time in the organic matter of the fractions isolated varies within a wide range: 30–75 years. According to the time of the carbon residence, two groups of organic matter are distinguished: the free organic matter (30–40 years) and the organic matter sorbed by the mineral matrix (65–75 years).

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