Organic carbon isotope ratio (??13C) of Arctic Amerasian Continental shelf sediments

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Organic carbon isotope ratios ($\delta^{13}C$) of Arctic Amerasian Continental shelf sediments

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Abstract Organic matter origins are inferred from carbon isotope ratios ($\delta^{13}C$) in recent continental shelf sediments and major rivers from 465 locations from the north Bering-Chukchi-East Siberian-Beaufort Sea, Arctic Amerasia. Generally, there is a cross-shelf increase in $\delta^{13}C$, which is due to progressive increased contribution seaward of marine-derived organic carbon to surface sediments. This conclusion is supported by the correlations between sediment $\delta^{13}C$, OC/N, and $\delta^{15}N$. The sources of total organic carbon (TOC) to the Amerasian margin sediments are primarily from marine water-column phytoplankton and terrigenous C_3 plants constituted of tundra taiga and angiosperms. In contrast to more temperate regions, the source of TOC from terrigenous C_4 and CAM plants to the study area is probably insignificant because these plants do not exist in the northern high latitudes. The input of carbon to the northern Alaskan shelf sediments from nearshore kelp community (*Laminaria solidungula*) is generally insignificant as indicated by the absence of high sediment $\delta^{13}C$ values (–16.5 to –13.6‰) which are typical of the macrophytes. Our study suggests that the isotopic composition of sediment TOC has potential application in reconstructing temporal changes in delivery and accumulation of organic matter resulting from glacial–interglacial changes in sea level and environments. Furthermore, recycling and advection of the extensive deposits of terrestrially derived organic matter from land, or the wide Amerasian margin, could be a mechanism for elevating total CO_2 and pCO_2 in the Arctic Basin halocline.

Keywords Arctic Amerasian shelf · Sediment carbon and nitrogen isotopes

Introduction

Organic carbon (OC) is a common but generally minor component (<50 mg g^{-1}) of marine sediments (Premuzic et al. 1982; Romankevich 1984). The concentration and composition of marine sediment OC and organic matter (OM) have a potential bearing on understanding marine depositional processes, paleoceanography, pollution, global carbon cycling, and exploration of fossil fuel reserves (Stein 1991; Zahn et al. 1994), and on the cycling of redox-sensitive elements (Finney et al. 1988; Gobeil et al. 1997). Due to the complex nature and variety of OM in marine sediment (Henrichs 1992; Goni and Hedges 1995), tools, such as stable isotopes ($\delta^{13}C$ and $\delta^{15}N$) and biomarkers, are commonly used to determine the sources of OC (Hedges and Mann 1979; Prahl and Muehlhausen 1989; Jasper and Gogosian 1989; Hayes et al. 1989; Yunker et al. 1993; Naidu et al. 1993).

The continental margin of the Arctic Ocean is the world's widest and accounts for 30% of the global ocean shelf area. It is an important sink for total

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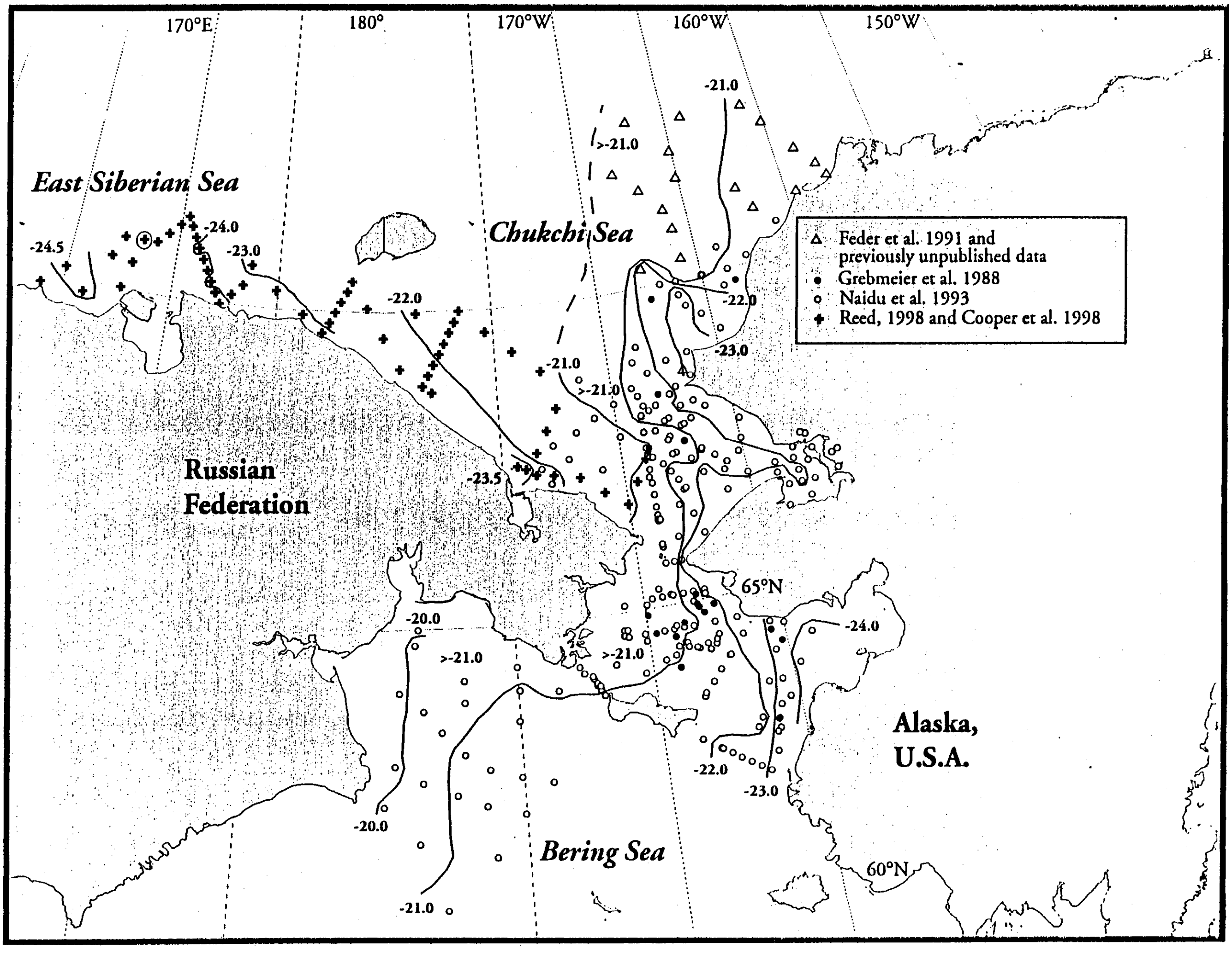
organic carbon (TOC) and a potential source of TOC to the adjacent deep basins (Walsh et al. 1989). Large lateral variations in $\delta^{13}C$ in organic matter of contemporary marine sediments may be anticipated in this area due to regional differences in proportions of marine algal production and terrestrial inputs from rivers, each of which has a different $\delta^{13}C$ composition (Naidu et al. 1993, and references therein). Variations in marine sediment $\delta^{13}C$ may also result from variations in marine OC caused by factors such as phytoplankton growth rate (Descolas-Gros and Fontugne 1990; Nakatsuka et al. 1992; Laws et al. 1995), algal cell size (Fry and Wainright 1991), sediment diagenesis, and $CO_2(aq)$ concentration (Rau et al. 1997). Although considerable data have been gathered on $\delta^{13}C$ in sediment TOC for selected seas marginal of the western Arctic Ocean (Gearing et al. 1977; Grebmeier 1987; Naidu et al. 1993; Grantz et al. 1996; Ruttenburg and Goni 1997; Cooper et al. 1998; Rachold and Hubberton 1999), large data gaps remain, and a coherent regional overview has not been produced. This has hampered clarification of the relative importance of factors which influence the $\delta^{13}C$ of sediment TOC in this region, and evaluation of this marker's potential as a proxy for Arctic paleoceanography especially in context of the $\delta^{13}C$ data reported from the Eurasian margin and basin (Knies and Stein 1998; Fahl and Stein 1999).

In this paper we synthesize published and new data on the sediment $\delta^{13}C$ of the Amerasian shelf of the western Arctic Ocean to gain a better understanding of the sources and distribution of TOC.

Study area

The study area consists of almost one-half of the continental margin of the Amerasian Arctic Ocean encompassing the contiguous north Bering-Chukchi-East Siberian seas, Beaufort Sea, and major river deltas (Figs. 1, 2). Comprehensive descriptions of these regions are provided elsewhere (Sharma 1979; Hood and Calder 1981; Herman 1989; Walsh et al. 1989;

Fig. 1 Locations of sediment samples from the north Bering-Chukchi-East Siberian seas and the distribution pattern of $\delta^{13}C$ in the sediments



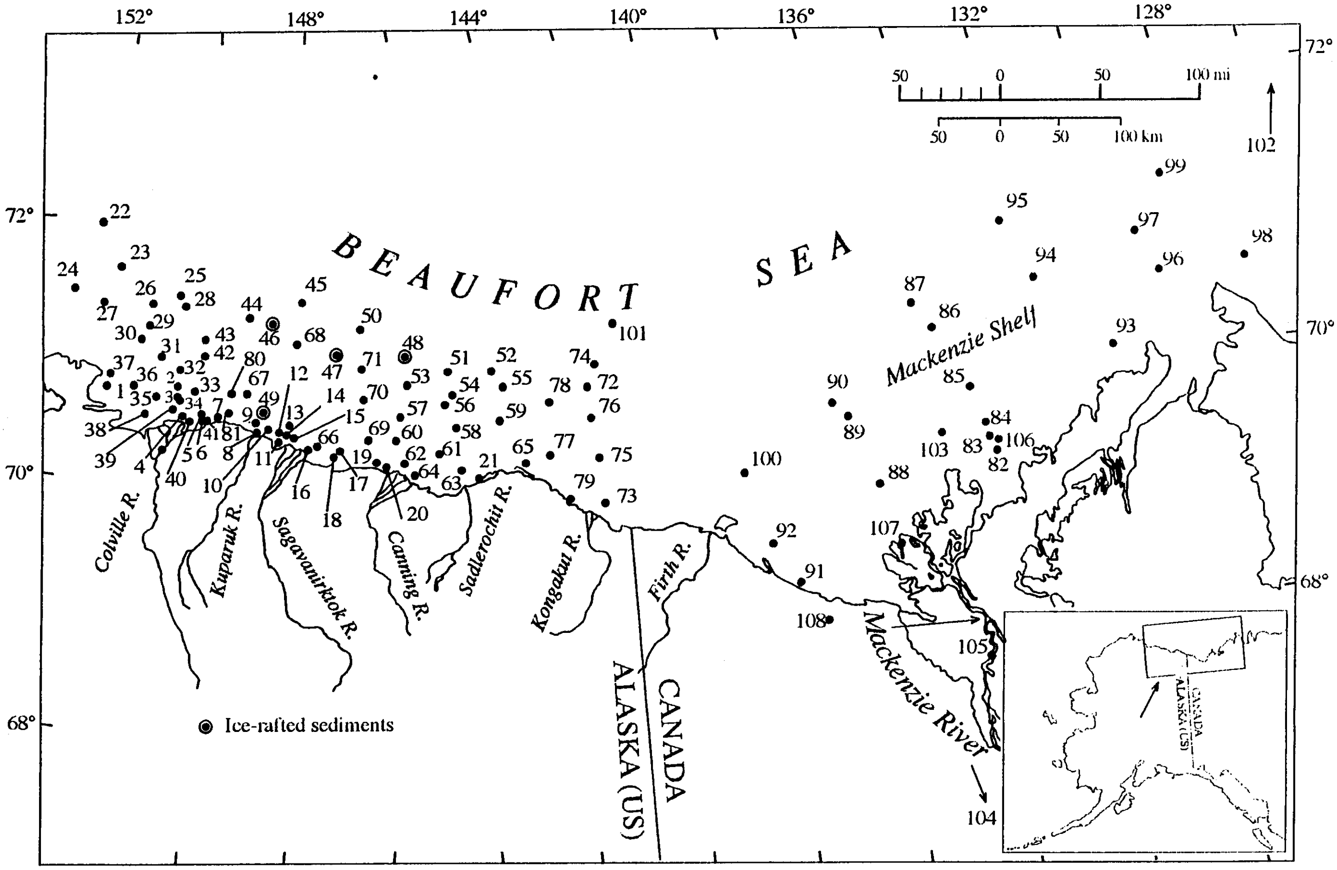


Fig. 2 Locations of sea-floor and sea-ice sediment samples from the Beaufort Sea. See Table 1 for further details on the sample locations

Grebmeier et al. 1995; Macdonald et al. 1998). With the exception of the north Bering Sea, which is a marginal sea of the North Pacific Ocean, the arctic seas are "Mediterranean" water bodies freely exchanging with the interior Arctic Ocean. As a result, the waters tend to be strongly salt stratified, due partly to river runoff and partly to inflowing Pacific water, which is fresher than the Atlantic water found at depth. Sea ice covers the region for 6–9 months of the year. The hydrochemistries of the major river systems affecting the study area are described by Telang et al. (1991), and the sources and fluxes of terrigenous dissolved organic matter in parts of the area and adjacent basin are addressed by Kattner et al. 1999; Opsahl et al. 1999.

Shelf sediments are typically a mosaic of poorly sorted coarse to fine sediment types with mud dominating off the major rivers (Naidu 1988). The transport, deposition, and reworking of sediments is strongly tied to ice, which both rafts the sediment and gouges the bottom. The deltaic processes in the arctic are unique: peak inflow precedes the breakup of coastal ice with the result that sediment-laden river water overflows the sea ice during spring (Dean et al. 1994). Furthermore, there is an absence of the delta-front platform (Naidu and Mowatt 1975). The presence of permafrost and peat have led to some of the world's highest coastal erosion rates, 1–20 m per year (Reimnitz and Barnes 1987), which provides an additional, widely distributed source of terrestrial POC to the nearshore sediments (Naidu et al. 1993; Macdonald et al. 1998).

With the exception of the northwestern Bering Sea and the contiguous southwestern Chukchi Sea, the shelves of the study area have generally low (<50 g C m^{-2} per year) primary production (Subba Rao and Platt 1984; Schell et al. 1984; Macdonald et al. 1998). The hydrography of the Bering Sea is dominated by the Pacific Ocean inflow, and a sharp vertical hydrographic front, extending north–south from the north Bering Sea to south Chukchi Sea. This front separates the nutrient-rich, colder, and more saline Anadyr water in the west from the nutrient-poor, warmer, and less saline Alaska Coastal water in the east (Coachman et al. 1975). Primary productivity in the Bering-Chukchi Sea is ~250–480 g C m^{-2} per year west of the front and 80 g C m^{-2} per year east of the front (cf. Grebmeier et al. 1995). In the study area and most of the Arctic, benthic production is closely coupled to pelagic production. Most of the phytodetritus is ungrazed in the water column and is deposited at the bottom where it supports a rich benthic system (Feder et al. 1994; Grebmeier et al. 1995). Ice-edge production can be substantial (Niebauer and Alexander 1985), but its quantitative importance to the total production has not been established.

Database, materials, and methods

The database for the $\delta^{13}C$ of sediment TOC is drawn from published values (Grebmeier et al. 1988; Feder and Naidu 1991; Feder et al. 1994; Naidu et al. 1993; Baskaran and Naidu 1995; Reed 1997; Cooper et al. 1998) supplemented by new data (Table 1). The data for the Colville, Lena, Mackenzie, and Anadyr rivers/deltas and most of the Beaufort Sea are new, with the exception of 16 sediment samples which were added from Gearing et al. (1977). Additionally, published (Naidu 1985) and new data on $\delta^{15}N$ and OC/N for selected Beaufort Sea sediments are reported here.

All sea-floor sediment samples are from the upper 3-cm oxidized/mixed sediment layer collected by either Van veen grabs, HAPS or gravity cores obtained at 465 locations (Figs. 1, 2; Table 1). Details on sample preparation and methods of analyses of $\delta^{13}C$ and organic carbon (OC) and nitrogen (N) are given by Naidu et al. (1993), Reed (1997), and Cooper et al. (1998). Briefly, after collection, the samples were placed in polyethylene bags and transferred frozen to the laboratory where each sample was thawed and a subsample was treated with 10% HCl to remove carbonate. The CO_3-free sample was washed to free excess acid and salts, oven dried and then finely pulverized. Individual splits of different batches of the powders were analyzed for $\delta^{13}C$ using three functionally similar mass spectrometers (VG Micromass Model 602E, Europa 20/20, and VG SIRA II). The $\delta^{13}C$ values are referenced to the V-PDB standard with a standard error of the $\delta^{13}C$ analysis of ±0.2‰. The OC/N ratios of sediment samples were based on the weight-to-weight percents of OC and N analyzed on splits of the CO_3-free powders, using either a P.E. Model 240B CHN analyzer or a Europa 20/20 isotope ratio mass spectrometer. The precision of the OC and N analyses was better than 10%. A selected number of the powdered samples was also analyzed for stable isotopes of nitrogen ($\delta^{15}N$) using the Europa 20/20 mass spectrometer. Correlations between $\delta^{13}C$, $\delta^{15}N$, and OC were evaluated by regression analyses.

Results

The OC/N, $\delta^{13}C$, and $\delta^{15}N$ values of sediment TOC of the Beaufort Sea floor, bed loads of the Colville, Mackenzie, Lena, and Anady rivers/deltas and sea-ice samples are presented in Table 1. In this table the OC/N values for selected locations (for those stations with no corresponding $\delta^{15}N$ data) are from Naidu (1985), the $\delta^{13}C$ for stations with map codes 50–65 are from Gearing et al. (1977) and the rest of the data are new. The distributional patterns of all available sediment $\delta^{13}C$ data for the study area (Figs. 1, 3) show notable trends. Within the north Bering Sea there is generally an east to west cross-shelf progressive increase in $\delta^{13}C$ from <–24.0 to –20‰ (Fig. 1). This trend continues in the Chukchi Sea with the exception that within the central–west Chukchi Sea region there is a reversal in the trend and relatively lower (–22‰) $\delta^{13}C$ values along the Russian nearshore. Another pattern is the decrease northwestward in $\delta^{13}C$ from –22 to –24.5‰ from the northwest Chukchi Sea to the East Siberian Sea (Fig. 1). In Beaufort Sea (Fig. 3) a broad successive seaward increase is observed in $\delta^{13}C$ from <–26.5 to >–23.0‰ from the Colville-Kuparak Delta complex to the continental margin edge. Generally, this trend extends in the contiguous Alaskan shelf in the east; however, in the latter inshore, the lightest $\delta^{13}C$ values (<–26.5‰, which are associated with the Colville-Kuparak prodelta) are absent. In the Canadian Beaufort Sea, there is also a broad progressive seaward increase in $\delta^{13}C$ from <–26.5‰ in the Mackenzie Delta to –24.5‰ at the Mackenzie Shelf edge (Fig. 3). The $\delta^{13}C$ of the sea-ice sediments of Beaufort Sea (map codes 46–49) vary from –25.3 to –19.3‰ (Table 1), whereas the sediments of the Colville, Mackenzie, Anadyr, and Lena rivers/deltas have typically lower $\delta^{13}C$ (<–26.5‰). For the Beaufort Sea, $\delta^{13}C$ values covary strongly with $\delta^{15}N$ and exhibit a negative correlation with OC/N (Fig. 4).

Discussion

For sediments, the factors controlling $\delta^{13}C$ values are the mean $\delta^{13}C$ value of the TOC assemblage initially accumulating at the sea floor and subsequent modification of this value during diagenesis. We discount diagenesis as a major factor for our study area, because there appears to be little difference between the $\delta^{13}C$ of settling POC (based on analyses of sediment trap samples) and sea-floor surficial sediment TOC of the Chukchi Sea and Beaufort Sea shelves (Minagawa et al. 1991; Naidu et al. 1993; Baskaran and Naidu 1995).

The overall cross-shelf seaward increases in $\delta^{13}C$ (Figs. 1, 3) are most simply explained as predominantly the product of mixing of TOC derived from two end-member sources – marine phytodetritus and terrestrial C_3 plants which consist chiefly of tundra angiosperms and taiga. Presumably terrestrial C_4 and CAM plants do not contribute carbon to the sediments investigated because these plants do not thrive in the northern high latitudes impacting our study area (Ehleringer 1979). In the northern Alaskan arctic shelf there is an additional potential source of carbon to the sediment TOC. Here, Dunton and Schell (1987) have reported presence in the nearshore of isolated boulder deposits that support a kelp community

Table 1 Locations of sediment samples from sea-floor and sea-ice of the Beaufort Sea, Colville, Mackenzie, Anadyr, and Lena rivers, and OC/N, δ^{13}C and δ^{15}N of the samples

Map code	Original sample no.	Longitude °W	Latitude °N	Water depth (m)	OC/N	δ^{13}C (‰)	δ^{15}N (‰)
Continental margin							
1	7E	152°04'	70°44'	3.3	15.4	–25.7	2.7
2	6D	150°29'	70°45'	18.4	11.3	–25.7	3.0
3	6C	150°32'	70°40'	16.1	10.5	–25.6	6.3
4	6B	150°25'	70°33'	5.5	9.5	–26.3	1.8
5	6A	140°58'	70°32'	3.6	12.6	–26.4	2.4
6	6G	149°54'	70°31'	2.1	12.9	–26.3	3.7
7	SL	149°39'	70°33'	2.4	12.9	–26.1	2.0
8	5F	148°50'	70°27'	1.5	17.7	–26.9	8.4
9	5A	148°46'	70°30'	11.4	10.9	–26.3	3.3
10	5(10)	148°30'	70°27'	8.2	8.8	–25.7	3.7
11	WPB	148°23'	70°21'	2.5	8.4	–25.4	3.0
12[a]	5(5)	148°18'	70°26'	6.7			
13	5G	148°05'	70°29'	9.3	9.6	–25.9	5.6
14	5(1)	148°04'	70°25'	5.8	7.7	–24.5	5.7
15[a]	5(2)	148°07'	70°25'	5.8			
16	4A	147°40'	70°19'	4.5	8.4	–25.9	4.0
17	3B	147°02'	70°18'	4.2	7.3	–25.3	2.3
18	3A	147°06'	70°17'	6.2	11.8	–26.0	2.9
19[a]	2E	146°12'	70°13'	7.4		–25.3	
20	2F	146°03'	70°10'	1.9	11.2	–26.1	3.5
21	1D	144°06'	70°05'	6.0		–25.4	
22	BSS 97	152°41'	70°53'	1500	8.0	–24.3	
23	BBS 95	152°00'	71°35'	545	9.0	–24.3	
24	BSS 84	153°00'	71°24'	75		–24.7	
25	GLA71 85	150°34'	71°22'	1053	12.0	–24.9	
26	GLA71 74	151°13'	71°20'	93		–24.0	
27	BSS 85	152°21'	71°19'	50	8.0	–24.2	
28	BSS 93	150°30'	71°20'	580		–24.7	
29	GLA71 72	151°14'	71°11'	47	9.0	–23.9	
30[a]	GLA71 71	151°22'	71°04'	21			
31	73ABP-9	150°56'	70°57'	17	8.0	–25.2	
32	AJT71-35	150°30'	70°52'	20	10.0	–25.2	
33	AJT71-36	150°08'	70°45'	17		–25.9	
34	AJT71-33	150°30'	70°41'	16	10.0	–26.0	
35	AJT72-43	151°00'	70°40'	14		–25.5	
36	AJT72-44	151°30'	70°45'	12	6.0	–25.9	
37	AJT72-45	152°00'	70°48'	7	5.0	–26.1	
38	HB71-5	151°12'	70°32'	3	16.0	–26.3	
39	HB71-3	150°37'	70°35'	3	17.0	–26.2	
40	HB71-6	149°58'	70°30'	3	16.0	–26.7	
41	SL877-25	149°52'	70°31'	3	21.0	–26.7	
42	GLA71-78	149°59'	70°58'	29	9.0	–24.8	
43	BSS-88	150°00'	71°05'	30	9.0	–24.9	
44	GLA71-58	149°03'	71°15'	1011	7.0	–23.3	
45	GLA71-21	147°55'	71°23'	2200	6.0	–22.1	
46	72ABP-34IR	146°31'	71°14'	995		–24.5	
47	GLA71-19IR	149°04'	71°00'	365		–21.5	
48	GLA71-7IR	145°35'	71°00'	509		–19.4	
49	72ABP-41IR	148°42'	70°35'	19		–25.3	
50	GR1	146°36'	71°11'			–22.6	
51	GR2	144°40'	70°52'			–22.6	
52	GR3	143°42'	70°52'			–23.3	
53	GR4	145°35'	70°46'			–23.6	
54	GR5	144°38'	70°42'			–23.5	
55	GR6	143°30'	70°45'			–23.4	
56	GR7	144°44'	70°38'			–23.9	
57	GR8	144°45'	70°33'			–23.8	
58	GR9	144°32'	70°28'			–24.3	
59	GR10	143°06'	70°30'			–23.4	
60	GR11	145°50'	70°23'			–25.0	
61	GR12	144°51'	70°17'			–24.3	
62	GR13	145°40'	70°13'			–25.5	
63	GR14	144°23'	70°08'			–25.0	
64	GR15	145°40'	70°06'			–24.8	
65	GR16	143°05'	70°11'			–21.9	
66	GLA71-25	147°31'	70°21'	26	9.1	–25.5	4.2
67	GLA71-63	149°00'	70°43'	26	8.9	–25.1	4.0

Table 1 (Continued)

Map code	Original sample no.	Longitude °W	Latitude °N	Water depth (m)	OC/N	δ^{13}C (‰)	δ^{15}N (‰)
68	GLA71-30	147°58'	70°05'	93	9.6	–25.3	
69	GLA77-31	146°26'	70°23'	28	9.9	–25.7	2.0
70	GLA77-32	146°31'	70°41'	42	12.4	–29.4	4.6
71	GLA77-33	146°35'	70°55'	66	8.0	–24.1	6.1
72	GLA77-9	141°39'	70°42'	640	7.8	–24.5	4.7
73	GLA77-12	141°29'	69°50'	22	9.7	–26.0	2.6
74	GLA77-8	141°30'	70°51'	2048	7.4	–24.2	5.0
75	GLA77-15	141°33'	70°11'	54	8.7	–25.7	3.3
76	GLA77-16	141°37'	70°28'	118	9.7	–25.8	2.5
77	GLA77-17	142°33'	70°14'	51	9.7	–26.1	2.2
78	GLA77-18	142°30'	70°37'	80	7.6	–24.9	4.2
79	BL-10	142°13'	69°52'	3	10.6	–26.0	2.3
80	71AJT-26	149°23'	70°43'	17		–26.8	
81	71AJT-40	149°23'	70°34'	10	14.1	–26.6	1.8
82	G4ST3	133°21'	69°47'	10	11.8	–27.6	2.3
83	G5ST4	133°25'	69°53'	16	10.7	–26.5	2.2
84	G8ST5	133°26'	70°01'	28	8.5	–26.0	4.6
85	G14ST7	133°39'	69°16'	50	7.7	–25.4	4.6
86	G11ST9	134°10'	70°47'	61	7.2	–24.8	5.5
87	G10ST10	134°36'	70°42'	205	8.7	–25.8	3.0
88	G35ST13	135°50'	70°57'	15	9.6	–26.1	1.8
89	G36ST16	136°15'	69°49'	48	8.1	–25.8	2.8
90	G37ST17	136°28'	70°11'	59	8.0	–25.7	4.0
91	G33ST19	137°46'	69°05'	16	9.9	–26.4	2.5
92	G31ST23	138°08'	69°23'	69	8.5	–25.7	6.8
93	G19ST29	130°57'	70°18'	17	9.2	–25.9	3.6
94[a]	G15ST22	131°42'	70°54'	50			
95	G15ST35	132°20'	71°25'	460	7.6	–24.9	5.3
96	G23ST39	128°55'	70°42'	22	11.3	–26.4	2.5
97	G26ST41	129°16'	70°57'	32	9.4	–25.1	3.9
98	G22SS1(47)	127°14'	70°43'	238	8.0	–24.6	5.1
99	G28SS2	128°34'	71°28'	167	8.1	–24.9	4.8
100	G30SS4	138°34'	69°56'	255	8.0	–25.3	4.4
101	ST44	141°24'	71°22'	2790	7.4	–25.5	6.3
102	9070L014	126°29'	73°12'	112	7.2	–23.1	8.1
	T3-1	141°51'	74°19'	3637	6	–24.2	
	T3-2	142°35'	74°35'	3650	7	–24.4	
River/delta							
Colville:						–26.9	
Mackenzie:							
103	M28	134°22'	70°01'		8.3	–26.4	2.8
104	M22	134°04'	67°35'		13.1	–26.5	1.4
105	M26	134°08'	68°22'		9.5	–26.8	2.13
106	M14	133°18'	69°52'		8.4	–26.5	2.7
107	M46	135°37'	69°19'		9.2	–27.2	1.9
108	M39	137°12'	68°52'		7.9	–27.7	1.9
Lena		Longitude °E					
	4	129°06'	72°01'	18.5	57.6	–26.9	3.0
	5	129°30'	71°49'	5.3	14.2	–27.2	2.2
	11	132°11'	71°26'	1.7	25.7	–26.0	5.6
	15	130°28'	71°37'	15	52.7	–26.0	4.9
	16	130°05'	71°37'	16	78.4	–25.5	6.3
Anadyr:							
	WD95WP2	177°50'	64°44'	10		–26.7	
	WD95WP4	176°49'	64°48'	3		–26.6	
	WD95WP8	176°58'	64°40'	7		–26.7	
	WD95WP11	176°28'	64°45'	9		–27.2	
	WD95WP15	177°42'	64°38'	8		–27.0	
	WD95WP17	178°15'	64°38'	3		–27.1	
	WD95WP21	177°48'	64°25'	3		–24.9	
	WD95WP25	177°53'	64°37'	9		–26.8	

[a] Concentrations of OC and N are very low; thus, no credible δ^{13}C and δ^{15}N values could be obtained

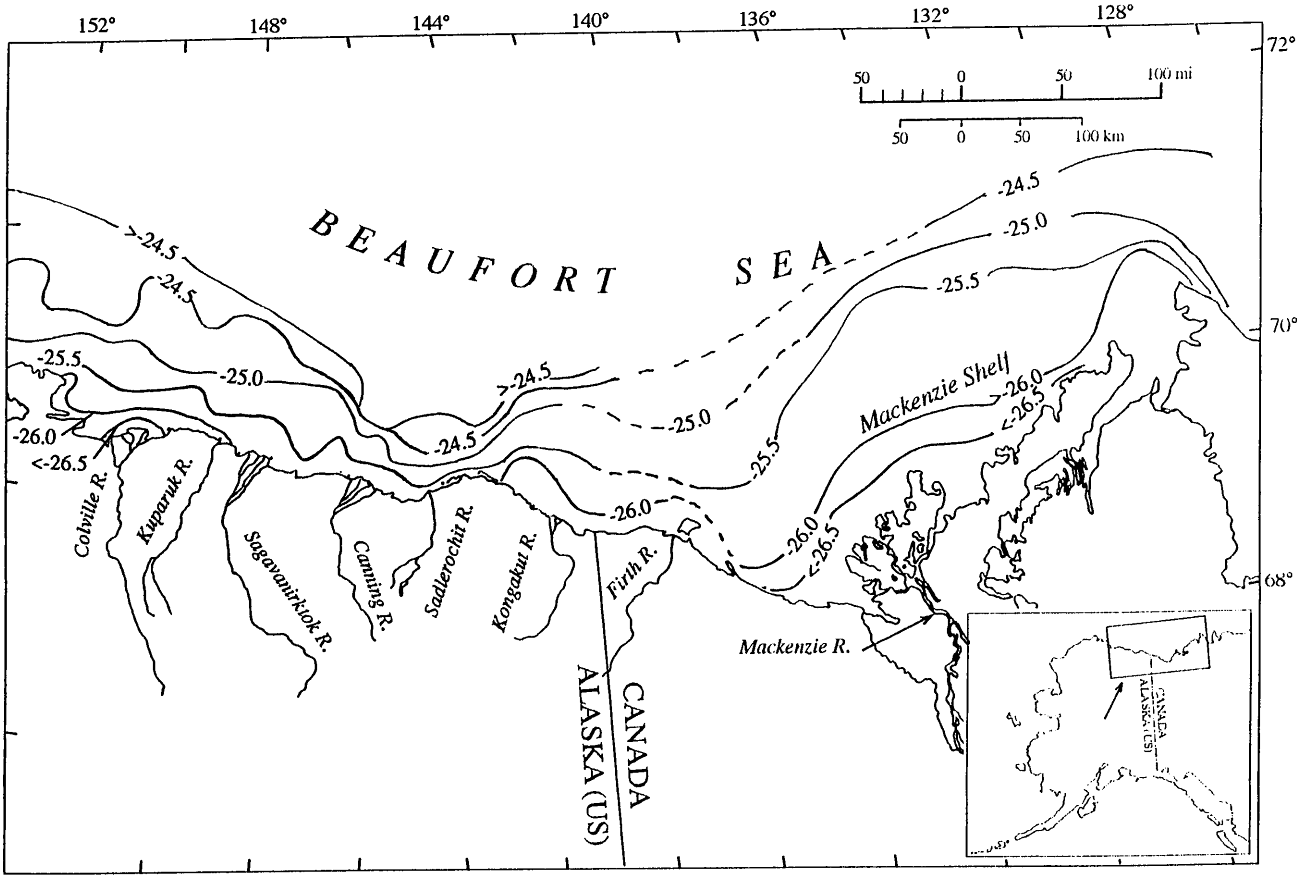


Fig. 3 Distribution pattern of $\delta^{13}C$ of sediments in Beaufort Sea

(*Laminaria solidungula*). The kelp carbon has $\delta^{13}C$ values of –13.6 to –16.5‰. That this source for carbon is not important is indicated by the lack of heavy $\delta^{13}C$ values in all the sediments investigated. In the northern Bering-Chukchi Sea the $\delta^{13}C$ of the terrestrial end member is –27‰ and the marine end member is –21‰ (see Table 1 for Anadyr River data; Naidu et al. 1993). The increase westward in $\delta^{13}C$ in the northern Bering-Chukchi Sea reflects an overall westward increase in marine POC that is commensurate with the progressively higher phytoplankton production and deposition of phytodetritus in nutrient-rich waters in the west (Naidu et al. 1993). Considering the light $\delta^{13}C$ values of the Anadyr River sediments (Table 1) and the relatively heavy $\delta^{13}C$ values in Gulf of Anadyr (Fig. 1), we infer that relatively little TOC from the river is admixed in the Gulf sediments. The contiguous band of $\delta^{13}C$ enriched sediments extending from the west Bering Strait to the northwest coast of Alaska is consistent with the northward transport of predominantly marine-derived POM from the northwestern Bering to northern Chukchi Sea margin (Walsh et al. 1989; Feder et al. 1994). As is discussed later, this northward transport of POM does not continue beyond the margin into the deeper Amerasian Basin. Westward from Bering Strait and the central Chukchi Sea the influence of terrestrial carbon to sediment TOC appears to increase in importance as suggested by the relatively lighter $\delta^{13}C$ values adjacent to the Chukotka coast of the Russian far east (Fig. 1). Likewise, the East Siberian Sea shelf sediments are strongly impacted throughout by inputs of terrestrial TOC as indicated by the relatively low $\delta^{13}C$ values there; however, the progressive decrease westward in shelf sediment $\delta^{13}C$ (Fig. 1) connotes increased fluvial inputs and transport of terrestrial TOC from the west.

In the Beaufort Sea the cross-shelf seaward increases in $\delta^{13}C$ (Figs. 1, 3) are, once again, most simply explained in context by the mixing of two end members (terrestrial and marine). The $\delta^{13}C$ value of the terrestrial end member for the Beaufort Sea TOC is approximately –27‰ based on the data for the Colville and Mackenzie River/Delta sediments and coastal peat samples (Table 1; Schell 1983; Minigawa et al. 1991; Ruttenberg and Goni 1997). This value is the same or close to that of the terrestrial end member of the northern Bering–south Chukchi seas (Naidu et al. 1993) and the Lena River/Delta (Table 1; Rachold and Hubberton 1999). Although an exact measurement of the $\delta^{13}C$ composition for the marine end member for the Beaufort Sea is not available, we estimate it to be approximately –24‰. This value is deduced from a per-trophic-level interpolation by 1‰ of the mean $\delta^{13}C$ value of the dominant zooplankton

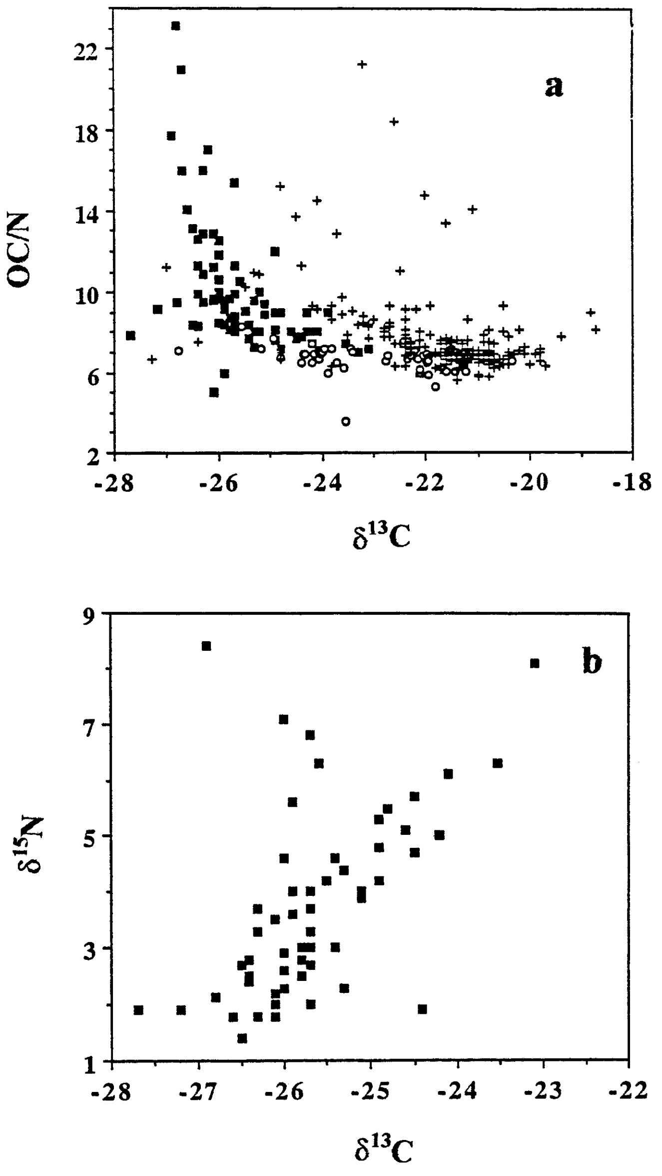


Fig. 4 **a** Correlations between δ^{13}C and OC/N based on all sediment samples analyzed from the study area, and **b** correlations between δ^{13}C and δ^{15}N of sediments based on data from the Beaufort Sea. Bering-Chukchi Sea data depicted by *open circles* are after Reed (1997) and data depicted by *crosses* are after Naidu et al. (1993). The *filled box plots* are from this study

primary consumers of the above region (Schell et al. 1998; see also Rau et al. 1983) and from δ^{13}C of sediment trap POC (Minigawa et al. 1991). The δ^{13}C values listed in Table 1 suggest that the Beaufort shelf sediments contain a predominance of terrestrial TOC, especially off the Colville and the Mackenzie rivers (Fig. 3). In contrast, the outer margin of the shelf has a significant portion of marine TOC. An estimate of the relative proportions of terrestrial and marine-derived TOC at any location can be made based on an isotopic mixing equation (Shultz and Calder 1976). Our interpretation that the Beaufort Sea surface sediments are strongly impacted by terrestrial POM, is consistent with biomarker data, elemental ratios, and δ^{13}C and OC budgets for the region (Yunker et al. 1993; Ruttenberg and Goni 1997; Macdonald et al. 1998). The relation between δ^{13}C and OC/N and the strong covariance between δ^{13}C and δ^{15}N (Fig. 4) strongly support the foregoing interpretation of the δ^{13}C data (cf. Naidu et al. 1993). Furthermore, the high OC/N (>15) and low δ^{15}N values (1–3‰) associated with our fluvial/delta sediment samples, and the most negative δ^{13}C (Table 1), are typical of terrestrial TOC (Bordowsky 1965). Likewise, the relatively low OC/N (<6) and high δ^{15}N (8‰), which generally correspond to our less negative δ^{13}C values, are typical of marine POC.

As discussed previously, terrestrial C_4 plants do not contribute carbon to the TOC of margin sediments of the Amerasian arctic. This is in contrast to the recent findings for some tropical and subtropical margins (e.g., northwestern Gulf of Mexico) where terrestrial TOC in sediments has a significant fraction ($\geq$50%) of carbon derived from C_4 plants (Goni et al. 1998).

The role of sea ice as a source (land vs marine) and agent of transport of TOC to the Amerasian continental margin and basin has not been quantitatively assessed, except for the Canadian Beaufort shelf (Macdonald et al. 1998). A mean δ^{13}C value of –23.6‰ (n=13, SD±0.18) has been measured for TOC entrained in sea ice along a north Chukchi Sea–North Pole Arctic Ocean Section (Cooper et al. 1998). This value is significantly lower than the Pacific-derived TOC deposited on the margin of the north Chukchi Sea (i.e., ca. –21‰; Fig. 1). The relatively lower δ^{13}C value (–25‰) of the Amerasian Basin abyssal seafloor sediments located beneath the sea-ice samples along the section (Grantz et al. 1996) and the δ^{13}C of approximately –24.3‰ of sediment samples collected from the adjacent central basin bottom (T3-1 and T3-2; Table 1) suggest no major transport of Pacific-derived, marine-enriched TOC to the western Amerasian Basin. Stein et al. (1994) also concluded that in general the sediments of the ice covered Arctic Eurasian Basin contiguous to our study area are dominated by terrigenous material. In the Alaskan Beaufort Sea shelf off the Colville River where the δ^{13}C of ice-entrained TOC is less than –24.5‰ (map code nos. 46 and 49), transport of terrigenous TOC by sea ice appears relatively important. In the outer margin, however, the transport by ice of marine TOC is dominant, as reflected by the –19.4 to –21.5‰ δ^{13}C values there (map code nos. 47 and 48). We note that the TOC in ice-rafted sediments of the outer margin of our study area and a few samples in the Arctic Ocean Section (Cooper et al. 1998) have δ^{13}C values that are significantly higher than the value estimated for the marine end member (–24‰). We attribute this difference to be a result of mixing of ice-entrained TOC derived from ice algae, which have been reported in the high arctic to have δ^{13}C values of approximately

–18‰ (Hobson and Welch 1992; Hobson et al. 1995). The increasing trend in the $\delta^{13}C$ of ice-rafted TOC from the shelf to the outer margin generally matches the seaward changes in bottom sediment $\delta^{13}C$ (Fig. 3).

The $\delta^{13}C$ value of POC estimated for the Beaufort Sea marine end-member (–24‰) is significantly less than that reported by Naidu et al. (1993) for the northern Bering-Chukchi Sea (i.e., –21‰). These regional differences are also manifested in the $\delta^{13}C$ of taxa zooplanktons of the Beaufort and Bering-Chukchi seas (Schell et al. 1998). The cause for this regional difference, which could be any of the factors indicated previously, has not been determined.

Recently, considerable interest has been focused on the sources and processes leading to the elevated ΣCO_2, pCO_2, DOC, and DOM in the upper halocline of the Arctic Basin (Walsh 1995; Sambrotto 1996; Opsahl et al. 1999). The elevated values have been explained by several possible factors, such as high respiration rates, high solubility of CO_2 in cold waters, in situ biological production of DOC, and horizontal advection of terrigenous DOM from the extensive arctic shelves to the basin (Walsh 1995; Opsahl et al. 1999; Kattner et al. 1999; Wheeler 1998). A major implication of our and several other recent studies (Macdonald et al. 1998; Rachold and Hubberten 1999; Stein et al. 1999) is that the sediment deposits of the Arctic Ocean margin contain high concentrations of land-derived TOC. It would thus seem that there is a sufficient pool of terrigenous organic matter in this margin that could be recycled and potentially enrich ΣCO_2 in the Arctic Basin halocline. This suggestion is consistent with recent preliminary studies (Macdonald et al. 1998; Kattner et al. 1999; Opsahl et al. 1999) and the concept that the ocean margins are a significant source of OM to the deep basin (Bauer and Druffel 1998). Further investigations are needed to clarify the role of biogeochemical cycling in sediments, sea ice, the water column, and terrestrial systems on the budget of organic matter in the Arctic Ocean.

Conclusion

In the Arctic Amerasian continental shelf and adjacent margins, the $\delta^{13}C$, $\delta^{15}N$, and OC/N composition of the organic matter of surficial sediments provides an estimate of the relative proportion of terrigenous and marine-derived organic matter. There is generally an across-the-shelf gradient in sediment $\delta^{13}C$ throughout the study area which reflects progressive seaward dilution of terrigenous organic matter by marine organic matter, a conclusion consistent with results of previous investigations. The study indicates that the $\delta^{13}C$ composition of TOC in sediments has potential paleoceanographic applications. Stratigraphic changes in $\delta^{13}C$ in conjunction with other parameters of sediment organic matter, such as hydrogen index, OC/N, and biomarkers (Stein et al. 1994; Fahl and Stein 1999), can provide useful information on temporal changes in the sources of organic matter (marine vs land derived) resulting from glacial–interglacial changes in sea levels, circulation, and river inflow. The high concentrations of terrigenous TOC on the Amerasian margin are a potential source of the elevated ΣCO_2 in the Arctic Basin halocline.

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