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11. VARIATIONS IN THE ACCUMULATION OF MARINE ORGANIC MATTER AND CARBONATES AT LEG 186 SITES[1]

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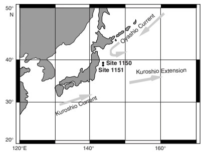
ABSTRACT

Past changes in sea-surface productivity in the Oyashio Current are evaluated on the basis of abundances of biological constituents in sediments from Leg 186 sites. Organic carbon contents at Sites 1150 and 1151 are moderate (0.5 to 1.5 wt%) and have an algal origin as indicated by low C/N ratios (<10) and by carbon isotopic compositions ranging from –23.4‰ to –21.3‰. A decreasing trend in organic carbon contents, carbon isotope ratios, and C/N ratios occurs with depth at both sites, probably as a consequence of diagenetic degradation of organic matter. Mass accumulation rates (MARs) determined for organic carbon and carbonates at Sites 1150 and 1151 show an abrupt increase between ~5 and 7 Ma. Similar results have been reported for sites in the Indian Ocean and the Pacific Ocean for the same time interval. As it has been previously suggested, the observed increase in MAR for both carbonate and organic carbon at Leg 186 sites probably resulted from augmented nutrient supply either from continental sources or from a more vigorous ocean circulation.

INTRODUCTION

Sites 1150 and 1151 are located near the confluence of two important currents in the western Pacific: the Kuroshio and the Oyashio Currents (Fig. F1). The Kuroshio Current is part of the North Pacific gyre, flowing northeast along the southeastern shore of Japan and transporting ~40 Sv (1 Sv = $10^6 m^3/s$). At ~ 35°N, the Kuroshio Current separates

F1. Locations of Leg 186 sites, p. 11.



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from the Japan coast by turning east to form the Kuroshio Extension Current (Fig. **F1**). The Oyashio Current is part of the North Pacific Current, flowing southward from the Bering Sea. High salinities and relatively cold temperatures characteristic of the subarctic Oyashio Current produce a permanent halocline between 100 and 200 m. Relative to the subtropical Kuroshio ocean current, the Oyashio Current shows characteristically higher rates of biological productivity, with daily production ranging from 0.45 to 1.51 g C/m^2 (Taniguchi and Kawamura, 1972). Lower rates of productivity in the oligotrophic Kuroshio Current, ranging from 0.05 to 0.1 g C/m^2, result from the presence of a permanent thermocline that prevents the transport of nutrients from the deep ocean to the surface (Ichimura, 1965).

The convergence between the Kuroshio and the Oyashio Currents presently occurs at ~40°N offshore Japan (Fig. **F1**), and it creates abundant eddies, potentially affecting the climate of Japan. The position where the Kuroshio Current diverges from the Japan coast to form the Kuroshio Extension appears to be persistent (Nishida, 1990). However, occasional swings in the mean position of the current do occur, and they are associated with dramatic changes in the climate of Japan. For example, cooler temperatures in Japan resulted when cold water masses from subarctic waters were able to reach more southerly latitudes as the Kuroshio Current moved south in 1954–1956.

Because of characteristic differences in surface productivity in the Kuroshio and Oyashio Currents, past changes in ocean circulation off eastern Japan can potentially be inferred from variations in sedimentary constituents related to sea-surface productivity. Geochemical proxies for biological accumulation were therefore studied at Leg 186 sites in order to assess changes and trends in primary productivity and to use that information to evaluate changes in ocean circulation in the northwestern Pacific. This research focuses on accumulation of organic matter and carbonate minerals present in the diatomaceous silty clays and claystones extracted during Leg 186. At present, productivity levels at Sites 1150 and 1151 are determined by the Oyashio Current. Thus, a northward shift in the current location of the Kuroshio Current would result in a decrease of organic carbon and carbonate abundances. Conversely, a southward shift of the Oyashio Current would cause increased productivity and would likely be associated with colder climates.

METHODS

Inorganic carbon was determined on board the *JOIDES Resolution* and details of methods are published elsewhere (Sacks, Suyehiro, Acton, et al., 2000). Briefly, ~10 mg of freeze-dried, powdered sediment was reacted with 2-N HCl. The evolved CO_2 was colormetrically titrated using monoethanolamine. Quantification of the evolved CO_2 was based on pure calcite standards. Carbonate content (in weight percent) was calculated from the inorganic carbon (IC) content, assuming that all the carbonate was present as calcite and using the following equation:

$$CaCO_3 \text{ (wt\%)} = IC \text{ (wt\%)} \times 8.332.$$

Total carbon (TC), hydrogen, nitrogen, and sulfur were analyzed using a Carlo Erba 1500 CNS analyzer. About 10 mg of freeze-dried, powdered sediment was combusted with V_2O_5 at 1000°C in a stream of oxygen. Nitrogen oxides were reduced to N_2 gas, and the mixture of evolved gases was separated by gas chromatography. Detection of separated gases was performed on a thermal conductivity detector (TCD) using sulfanilamide as a calibration standard. The amount of organic carbon (C_{org}) was calculated as the difference between total carbon and IC. It is possible, however, that not all carbonates reacted with dilute HCl, resulting in an overestimation of C_{org} values. To verify C_{org} content, freeze-dried, powdered samples of known weight were reacted with vapors of concentrated HCl overnight. The residue was freeze-dried and combusted in a LECO C/S elemental analyzer. Differences in C_{org} content obtained from these two protocols were <5%. For isotopic analyses, aliquots of freeze-dried, decalcified samples were placed in quartz tubes with ~1 g of CuO. The tubes were sealed under vacuum and placed in an oven at 850°C overnight. The evolved CO_2 was cryogenically purified and collected in a pyrex tube. The purified CO_2 was isotopically analyzed on a Finnigan MAT 252, using triphenylamide as an internal standard. Isotopic values are reported in per mil, according to the following standard equation:

$$\delta^{13}C = (R_{sample}/R_{standard} - 1) \times 1000,$$

where $R = {}^{13}C/{}^{12}C$.

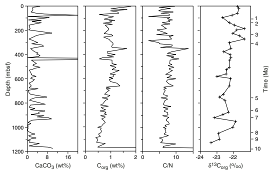
Accuracy (<0.05‰) and precision (<0.15‰) for the isotopic data were established on the basis of repeated analyses of the studied samples and standards of known isotopic composition.

RESULTS AND DISCUSSION

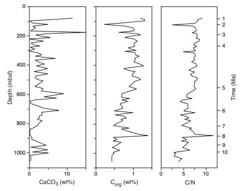
Carbonate Accumulation

Carbonate contents show an irregular distribution with depth in Holes 1150A and 1151A (Figs. F2, F3), ranging from 25 wt% to values below the detection limit of 0.01 wt% (Tables T1, T2). Carbonate contents in ocean sediments depend on a number of factors including (1) productivity of carbonate shell-bearing organisms, (2) variation of the carbonate compensation depth (CCD), (3) dilution by detrital material, and (4) postdepositional (diagenetic) processes (Libes, 1992). The presence of well-preserved calcareous tests of marine organisms throughout the studied sedimentary column at Sites 1150 and 1151 (Sacks, Suyehiro, Acton, et al., 2000) strongly suggests that sedimentation occurred above the CCD. Currently, the North Pacific CCD occurs at a depth of ~3000 m, although it rises to a depth of ~2400 m in coastal and highly productive areas of the North Pacific Ocean (Rea et al., 1995). The water depth of Sites 1150 and 1151 are 2681 m and 2182 m, respectively. It is then possible that variations in carbonate content in the studied sediments are associated with fluctuations in the CCD, given that the modern CCD occurs near the water depth of these sites. However, the reconstructed North Pacific CCD indicates that the modern depth has been the shallowest during the Cenozoic (Rea et al., 1995). Consequently, it appears unlikely that the ~1300- to 1600-m rise of the CCD in the North Pacific Ocean during the last 25 m.y. could have significantly affected carbonate contents in sediments recovered from Sites 1150 and 1151 because these sites were most likely above paleo-CCD.

F2. Carbonate, C_{org}, organic carbon/nitrogen ratio, and $\delta^{13}C_{org}$, Hole 1150A, p. 12.



F3. Carbonate, C_{org}, and organic carbon/nitrogen ratio, Hole 1151A, p. 13.



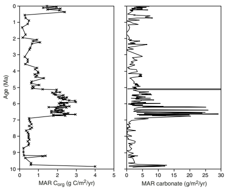
T1. Geochemical results, Site 1150, p. 16.

T2. Geochemical results, Site 1151, p. 17.

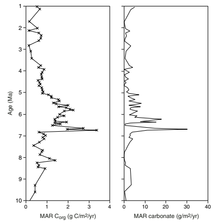
One of the most important diagenetic processes affecting carbonate dissolution is microbial degradation of organic matter (OM) in ocean sediments (Berger et al., 1982). Carbon dioxide formed during the microbial oxidation of OM shifts the carbonate equilibria in pore water, resulting in the dissolution of carbonate minerals (Berger, 1970; Hales and Emerson, 1996). The rate of interstitial carbon dioxide production depends on the level of oxycity in the sediments and on the type of OM being delivered to the ocean bottom (Calvert, 1987). In general, algal OM is more easily degradable than terrestrial OM (Opsahl and Brenner, 1995), potentially resulting in faster rates of carbon dioxide production. Although OM deposited at Sites 1150 and 1151 is mainly marine in origin (see below) and consequently is likely to be readily oxidized, dissolution caused by this oxidation was probably minimum as indicated by the good preservation of foraminiferal tests in the sediments (Sacks, Suyehiro, Acton, et al., 2000). However, some carbonates at Sites 1150 and 1151 correspond to dolomite (Sacks, Suyehiro, Acton, et al., 2000), strongly suggesting a diagenetic origin.

Mass accumulation rates (MARs) of carbonates are a better representation of the delivery of carbonates to the ocean floor because they account for dilution effects caused by changing sedimentation rates. The age model employed to estimate MAR for carbonates at Leg 186 sites is based on a combination of biostratigraphic events and magnetic polarity (Sacks, Suyehiro, Acton, et al., this volume). Estimations of MAR for carbonates vary from values as high as 30 $g/m^2/yr$ to almost negligible values (Figs. F4, F5). Notably, high MAR values occur in sediments accumulated between 5 and 7 Ma. Although diagenetic formation of carbonates can potentially augment MAR, increased delivery of calcareous tests from planktonic organisms seems a more plausible cause for the observed high MAR values between 5 and 7 Ma. An independent evaluation of the possibility that increased productivity of planktonic organisms produced the observed high MAR values between 5 and 7 Ma at Sites 1150 and 1151 can be achieved by studying accumulation of organic carbon matter, which is another proxy for primary sea-surface productivity.

F4. MAR, Hole 1150A, p. 14.



F5. MAR, Hole 1151A, p. 15.



Organic Matter Accumulation

Organic matter contents in sediments recovered at Sites 1150 and 1151 show relatively moderate values, ranging from ~0.5 to 1.5 wt%. A decreasing trend in organic carbon abundance occurs with depth at both sites (Figs. F2, F3). Although organic carbon contents of hemipelagic sediments are generally related to productivity, changes in sedimentation rates can strongly affect organic carbon contents by diluting accumulating OM. Mass accumulation rates of organic carbon, however, account for changing sedimentation rates, thereby better representing the delivery of OM to the sediments. MAR for organic carbon in Holes 1150A and 1151A ranges from ~ 0.2 to 3 g $C/m^2/yr$ (Figs. F4, F5). Relatively high values occurred in sediments deposited between 5 and 6.8 Ma and between 0.5 Ma and recent in Hole 1150A (Fig. F4). Increased delivery of OM between 5 and 7 Ma is also evident in Hole 1151A (Fig. F5). However, the significant increase in OM accumulation at Site 1150 during the last 0.5 m.y. is unresolved in the record of Hole 1151A.

Increased delivery of organic carbon between 5 and 7 Ma can be explained by (1) enhanced input of terrigenous OM, (2) increased productivity of marine organisms, (3) increased accumulation rates that enhanced the preservation of OM by reducing the exposure time to oxic conditions, or (4) formation or enhancement of anoxic conditions at the sediment/water interface (Emerson and Hedges, 1988). An evaluation of the source of OM and the sedimentary conditions during OM accumulation assists with the examination of these possibilities.

Organic carbon/nitrogen (C/N) ratios can be used to assess OM sources because terrestrial OM typically shows C/N ratios >15, whereas marine algal OM exhibits C/N ratios <10 (Tyson, 1995). This difference in C/N ratios is due to the abundant presence of N-depleted compounds in vascular plants, such as lignin and cellulose. C/N ratios in sediments at Sites 1150 and 1151 range from 2.5 to 12, indicating a predominantly marine source for the accumulated OM (Figs. F2, F3). Original C/N ratios, however, can be affected by diagenetic factors. For example, a decreasing trend in C/N ratios is observed at Sites 1150 and 1151. This trend is probably the result of degradation of organic matter as suggested by a correlative decreasing trend in organic carbon content with depth (Figs. F2, F3). Another diagenetic process affecting C/N ratios is the incorporation of ammonia produced during OM degradation in the interlayer positions of clay minerals. This process decreases C/N ratios of bulk nitrogen determinations (Müller, 1977), and it can be identified when C/N ratios are lower than those existing in marine organisms. Whereas photosynthetic algae typically exhibit C/N ratios of ~4 (Redfield et al., 1982), some intervals at Sites 1150 and 1151 exhibit C/N ratios of <4 (Figs. F2, F3), thereby indicating the diagenetic incorporation of ammonia in clay minerals.

Although C/N ratios at Leg 186 sites fall within those reported for marine algae, some variability in these ratios exists in sediments from Holes 1150A and 1151A, probably as a result of different degrees of OM degradation. Depending on pore water conditions and the nature of OM, it is possible that some intervals have experienced a larger degree of microbial degradation that produced various degrees of OM alteration. For example, sedimentation rates can potentially control the degree of OM degradation by determining the residence time of OM in oxic zones. At high sedimentation rates, settling OM is rapidly buried and removed from the water/sediment interface, where a significant degree of degradation occurs (Müller and Suess, 1979). No apparent correlation exists, however, between sedimentation rates and C/N ratios at Sites 1150 and 1151 (Figs. F3, F4). Alternatively, changes in the quantity of terrestrial OM delivered to the ocean floor can also account for the observed C/N variability at Sites 1150 and 1151, due to the presumed resistance of terrestrial OM to microbial degradation.

Another parameter used to distinguish terrestrial and marine OM sources is the carbon isotopic composition of OM ($\delta^{13}C_{org}$). Whereas terrestrial vascular plants use atmospheric CO_2 as a carbon source and result in $\delta^{13}C_{org}$ values of about –27‰ ± 3‰, algae assimilate dissolved CO_2 or bicarbonate and result in $\delta^{13}C_{org}$ values of about –21‰ ± 2‰ (O'Leary, 1988). This difference in $\delta^{13}C_{org}$ values helps to evaluate the relative contribution of terrestrial and marine OM (e.g., Prahl et al., 1994). Values of $\delta^{13}C_{org}$ in Hole 1150A sediments range from –23.4‰ to –21.3‰ (Fig. F2), suggesting a contribution of predominantly marine OM to the sediments. This suggestion is in agreement with the marine origin for the accumulated OM inferred from C/N ratios. Values of $\delta^{13}C_{org}$ show a decreasing trend with depth at Site 1150. This trend parallels that of organic carbon contents and C/N ratios, thereby indicating that microbial degradation of sedimentary OM produced a preferential loss of compounds enriched in ^{13}C.

Superimposed on the decreasing trend, $\delta^{13}C_{org}$ values show a positive excursion between 3 and 4 Ma, followed by a negative shift (Fig. F2). These excursions may be the result of changes in the dissolved CO_2 that produced variations in marine algae $\delta^{13}C_{org}$ values (Rau et al., 1992; Hayes, 1993). In today's oceans, there is a latitudinal gradient in the isotopic composition of marine algae, showing relatively low values near the poles and relatively high values in the tropical oceans. This isotopic gradient appears to obey the latitudinal variation in sea-surface temperatures that control CO_2 solubility (Rau et al.,1989). Thus, variations in sea-surface temperatures during the last 4 m.y. could explain the observed $\delta^{13}C_{org}$ trend at Site 1150. In addition to temperature, productivity influences the isotopic composition of marine algae. At oligotrophic conditions, algae can maximize the preferential discrimination of ^{13}C during photosynthesis, resulting in relatively low $\delta^{13}C_{org}$ values. In contrast, high productivity typical of upwelling zones causes algae to show relatively high $\delta^{13}C_{org}$ values (e.g., Pankost et al., 1999). The possibility that changes in productivity produced the observed excursions in $\delta^{13}C_{org}$ values at Leg 186 sites seems unlikely, as indicated by the lack of correlation between $\delta^{13}C_{org}$ values and elevated C_{org} contents and organic carbon accumulation rates (Fig. F2). In fact, the presence of diatomaceous sediments at Sites 1150 and 1151 suggests relatively high primary productivity. Consequently, changes in carbon isotope ratios are most likely due to changes in solubility of CO_2 controlled by temperature fluctuations.

Paleoceanographic Implications

Results at Leg 186 sites indicate that primary productivity has been relatively high, as manifested by the presence of diatomaceous sediments throughout the studied cores and by relatively high organic carbon and carbonate contents. There was, however, an even higher productivity event between 7 and 5.3 Ma, as indicated by high MAR values for carbonates and organic carbon. Similar results of enhanced productivity for this time period have been obtained for a number of records in the Indian Ocean and the Pacific Ocean (e.g., Peterson et al., 1992; Dickens and Owen, 1996; Seisser, 1995). Because the residence time of limiting marine nutrients (nitrate and phosphate) is shorter than the 2-to 4-m.y.-long algal bloom interpreted for the Pacific and Indian Oceans, a mechanism is needed to explain both augmented surface productivity and the sequestering of nutrients in sediments. The invoked mechanisms include an external (continental) delivery of nutrients and an internal (oceanic) supply of nutrients from the deep ocean. Delaney and Filippelli (1994) and Rea et al. (1995), among others, argue that this period of enhanced productivity was the result of higher input of nutrients resulting from weathering of the Himalayan Mountains. Under this scenario, the Himalayas reached a height sufficient to trap storm tracks from the Indian Ocean at ~8 Ma, producing an intensification of the Asian monsoon and the consequent net increased transfer of nutrients from land to oceans (Filippelli, 1997). An additional supply of nutrients to the world's oceans could come from the South American continent as a result of the main uplift of the Andes Mountains, which also occurred in the late Miocene (Gregory-Wodzicki, 2000).

Although enhanced weathering explains the observed increase in MAR, another possible scenario involves increased nutrient supply that resulted from a more vigorous oceanic circulation, which transferred nutrients to the Pacific and Indian Oceans from other oceanic localities (Dickens and Owen, 1996; Hermoyian and Owen, 2001). The implications for each mechanism are different. Enhanced continental delivery of nutrients to the oceans promoting elevated productivity would result in a net drawdown of atmospheric carbon dioxide as marine organic carbon and carbonate burial increased. In contrast, a redistribution of nutrients caused by an intensification of ocean circulation would result in enhanced productivity in some areas but in decreased productivity in others, with no detectable change in atmospheric CO_2 concentrations. Although results from Sites 1150 and 1151 cannot resolve any of these mechanisms, they do indicate that the productivity event of the late Miocene and early Pliocene also occurred in the northwestern Pacific, and it is consistent with the suggestion that this productivity event resulted from the establishment of the modern deep-water circulation that brings nutrients to the Oyashio Current (Rea et al., 1995).

CONCLUSIONS

Results from this study indicate that the sedimentary organic matter at Sites 1150 and 1151 comes from marine photosynthetic organisms, as suggested by low C/N ratios and values of $\delta^{13}C_{org}$ typical of algae. However, diagenetic degradation of organic matter has affected the values for C/N and $\delta^{13}C_{org}$, producing a decreasing trend of these two geochemical parameters with depth. Superimposed on this diagenetic trend, $\delta^{13}C_{org}$ values show ~1.5‰ fluctuations related to changes in the solubility of CO_2 in seawater.

Moderate values for organic carbon derived from primary producers coupled with the persistent presence of diatomaceous sediments at Leg 186 sites suggests relatively high rates of primary productivity. This condition is characteristic of the Oyashio Current, thereby suggesting that the Kuroshio Current rarely reached the latitude of Sites 1150 and 1151 throughout the Cenozoic. The relative high rates of primary productivity in the Oyashio Current significantly increased between 5 and 7 Ma, as indicated by higher MAR for carbonate and organic carbon at Leg 186 sites.

The results presented in this study are in agreement with previous reports for increased surface productivity during the late Miocene and early Pliocene. However, MAR values at Leg 186 sites could be somewhat affected by diagenetic formation of carbonate minerals and lower degradation rates of organic carbon. The extent of increase in MAR values is large enough that it is unlikely to be solely the result of diagenesis. Therefore, the interpretation that increased MAR values between 5 and 7 Ma are the result of eutrophic conditions in the Oyashio Current implies an ocean-wide increase in the delivery of nutrients. However, the source of these nutrients remains to be determined.

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Figure F1. Location of Leg 186. Diagram shows the mean position of the Oyashio and the Kuroshio Currents and the meanders that occurred in 1954 and 1959.

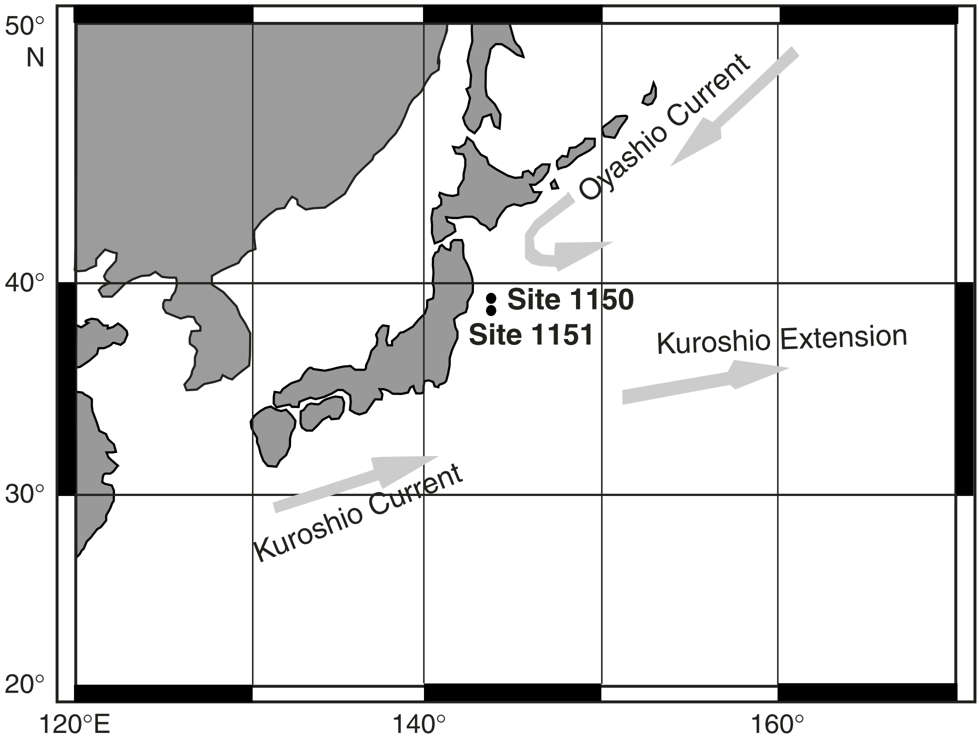

Figure F2. Carbonate abundance, organic carbon content (C_{org}), ratio of organic carbon and nitrogen, and carbon isotopic composition ($\delta^{13}C_{org}$) of organic carbon in sediments in Hole 1150A.

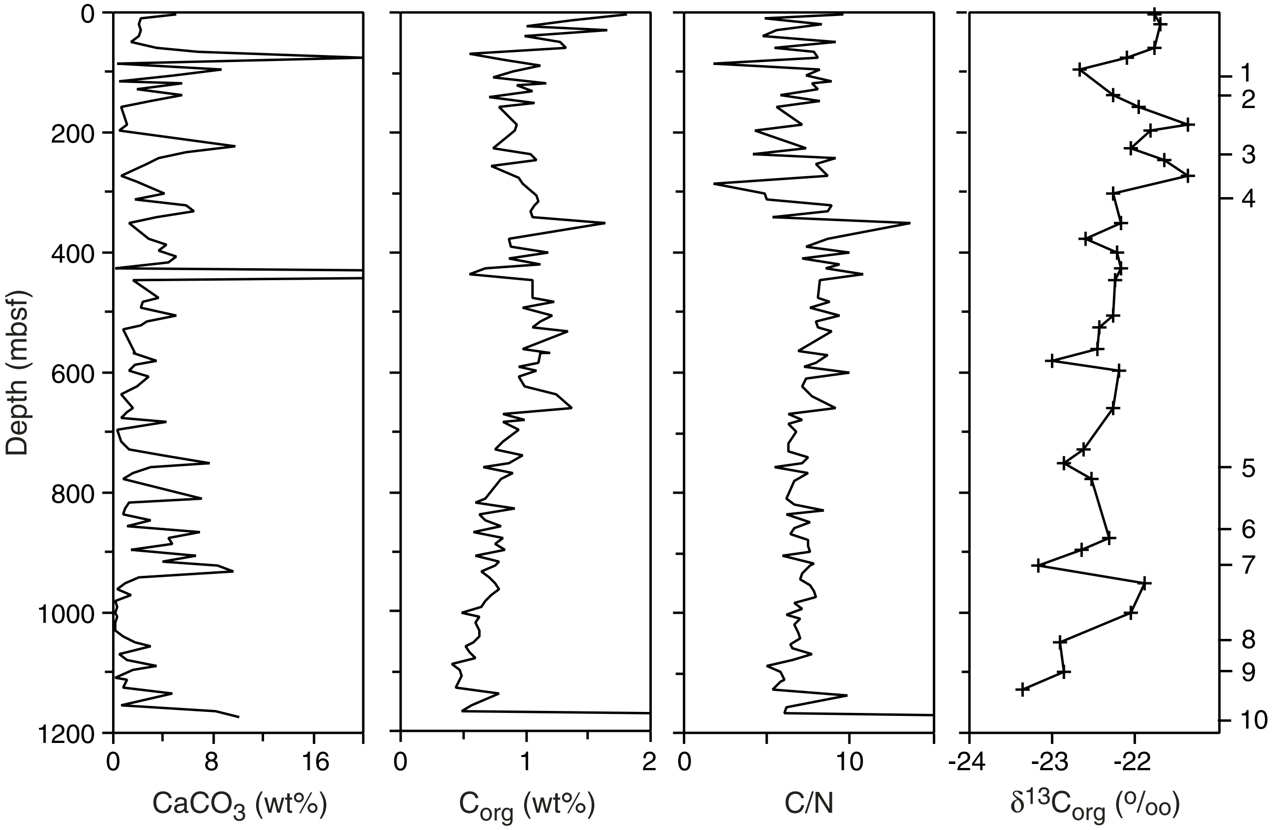

Figure F3. Carbonate abundance, organic carbon content (C_{org}), and ratio of organic carbon and nitrogen in sediments in Hole 1151A.

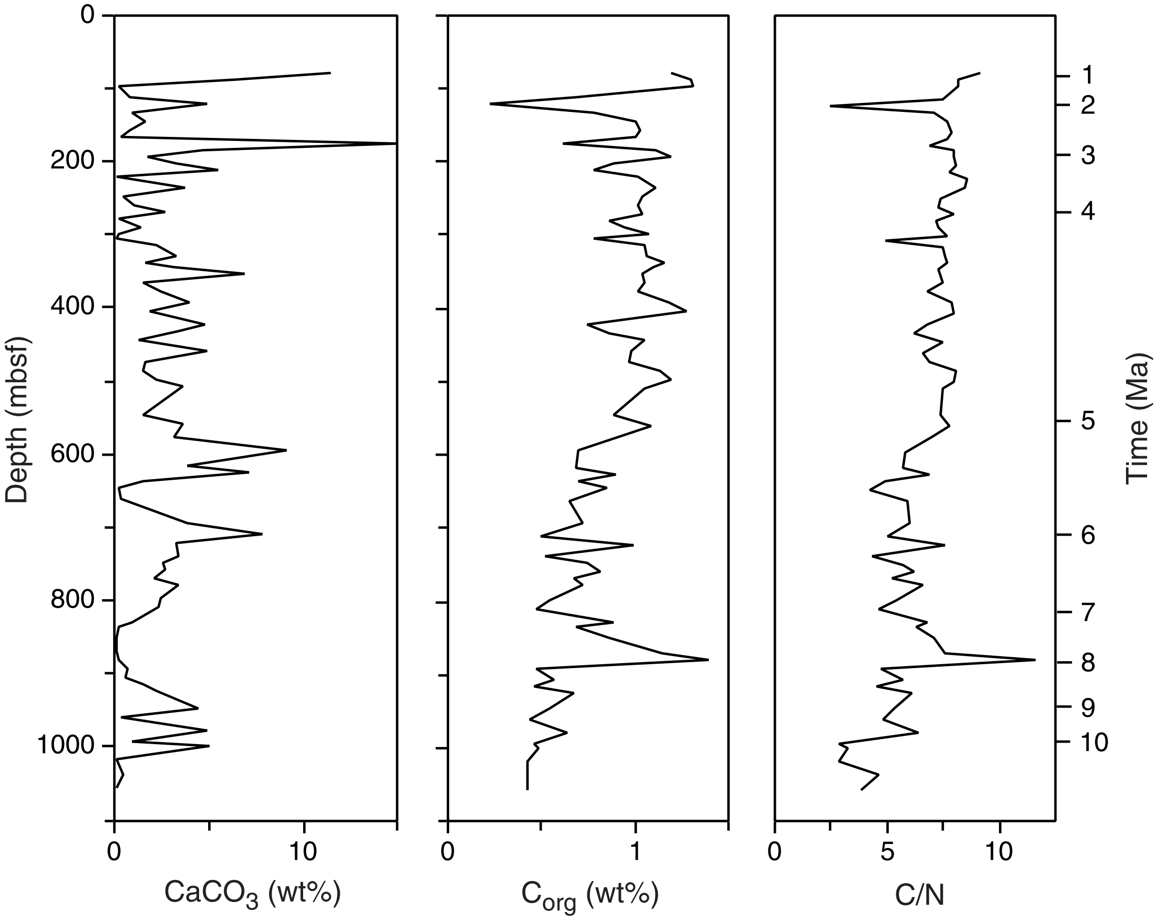

Figure F4. Mass accumulation rates (MAR) for organic carbon and carbonates for sediments in Hole 1150A.

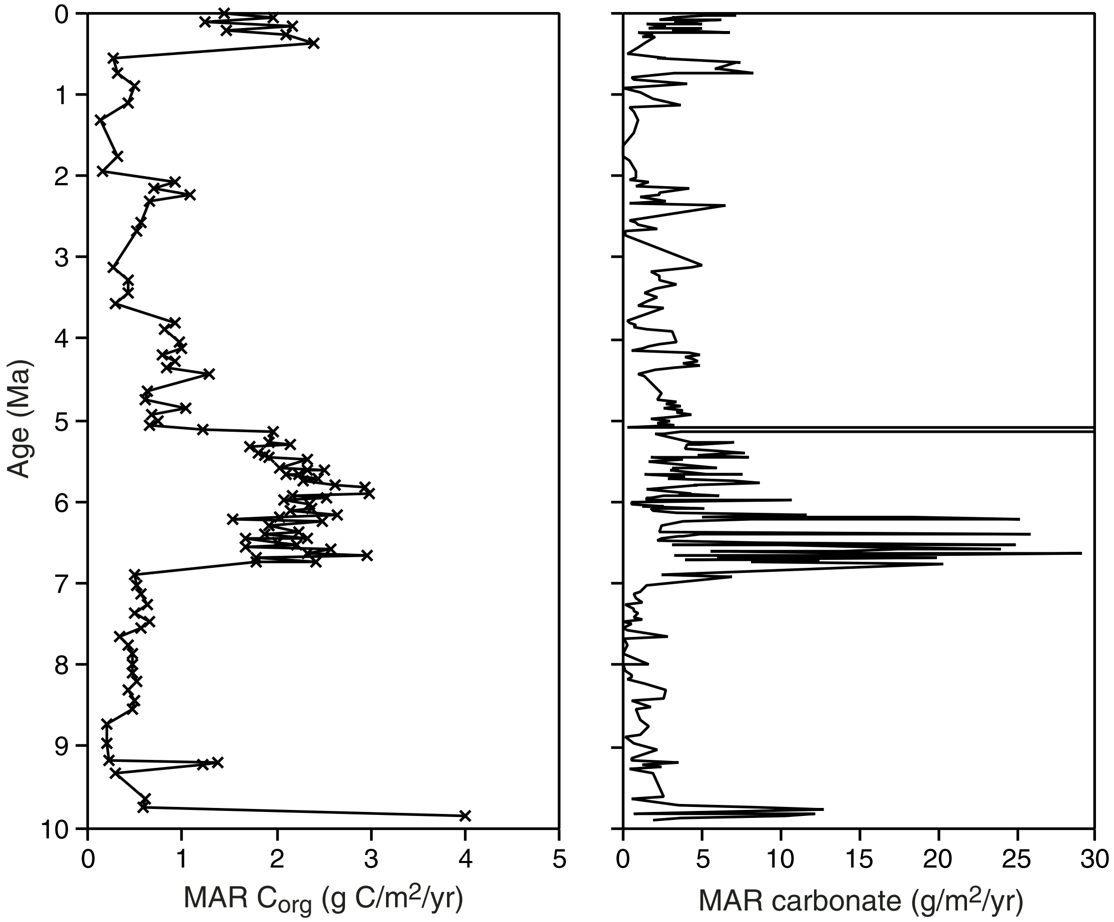

Figure F5. Mass accumulation rates (MAR) for organic carbon and carbonates for sediments in Hole 1151A.

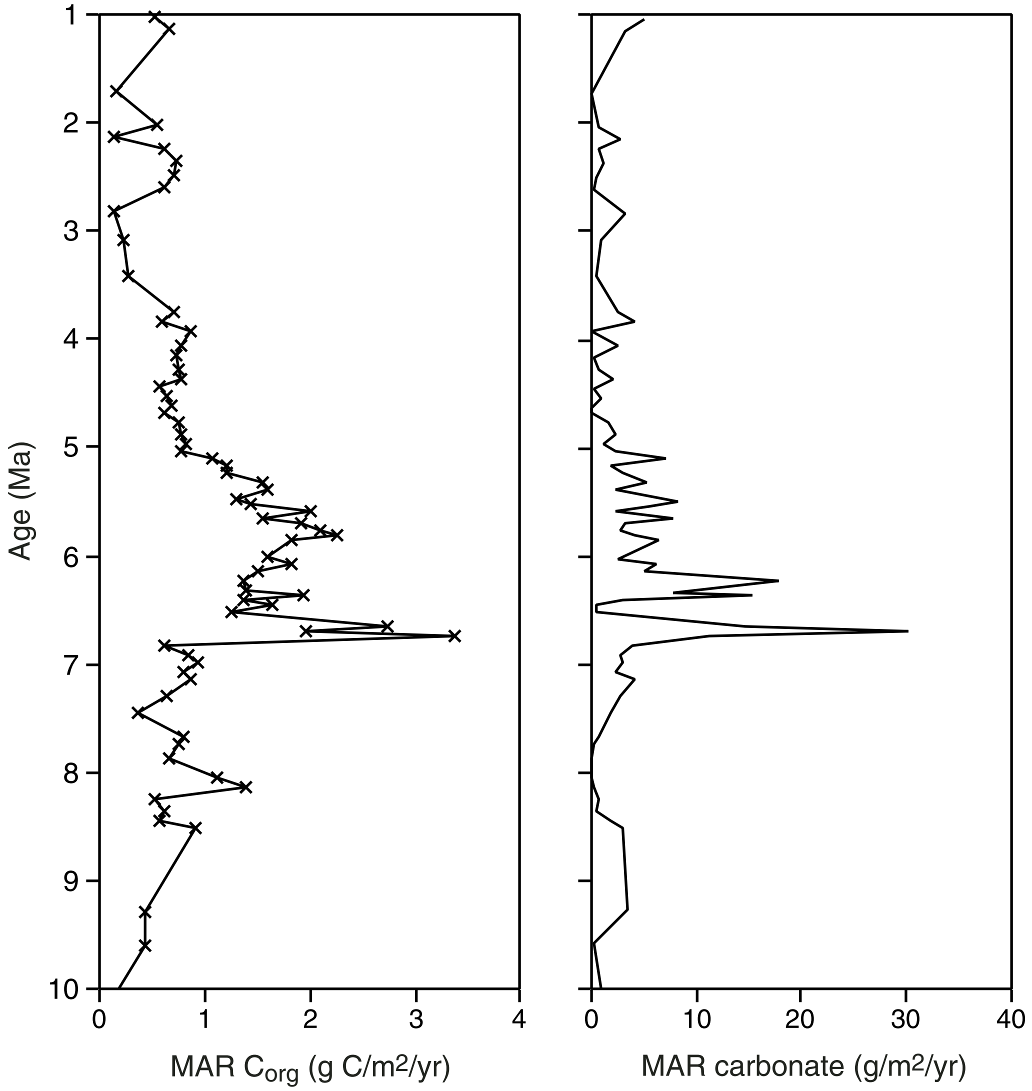

Table T1. Geochemical results for sediments at Site 1150.

Core, section, interval (cm)	Depth (mbsf)	$CaCO_3$ (wt%)	C_{org} (wt%)	C/N	$\delta^{13}C_{org}$ (‰)
186-1150A-					
1H-1, 89–91	0.89	2.25			
1H-3, 87–88	3.87	5.04	1.82	9.58	–21.764
2H-3, 87–88	11.57	2.18	1.37	4.89	
3H-3, 82–83	21.02	2.06	1.00	8.33	–21.697
4H-3, 71–72	30.41	2.26	1.66	5.53	
5H-3, 99–100	40.19	2.11	0.99	4.71	
6H-3, 95–96	49.65	1.49	1.27	9.07	
7H-3, 89–90	59.09	3.39	1.32	5.50	–21.768
8H-3, 69–70	68.39	6.72	0.55	7.86	
9H-3, 62–63	77.82	21.90	0.81	8.10	–22.105
10H-3, 69–0	86.56	0.28	1.12	1.84	
11H-3, 93–94	97.13	8.72	0.90	8.18	-22.675
12H-3, 95–96	106.3	4.83	0.74	7.40	
13X-3, 88–89	116.1	0.41	1.16	8.92	
14X-3, 110–111	120.5	5.53	0.93	7.75	
15X-3, 72–73	129.7	1.94	1.05	8.08	
16X-3, 73–74	139.3	5.50	0.70	5.83	–22.26
17X-3, 97–98	149.2	2.94	1.07	8.23	
18X-3, 73–74	158.5	0.64	0.78	5.57	–21.957
21X-3, 72–73	187.3	1.06	0.93	7.15	–21.366
22X-3, 110–111	197.4	0.46	0.91	4.33	–21.828
25X-3, 89–90	226.1	9.77	0.74	7.40	–22.049
26X-3, 68–69	235.6	5.89	1.04	4.16	
27X-3, 72–73	245.2	3.69	1.09	9.08	–21.665
28X-3, 72–73	254.8	2.67	0.72	8.00	
30X-3, 78–79	274.2	0.61	0.95	8.64	–21.371
31X-4, 89–90	284.7	1.93	0.98	1.85	
33X-3, 113–114	303.5	4.13	1.08	4.91	–22.273
34X-3, 139–140	313.5	1.68	1.10	5.00	
35X-3, 50–51	321.6	5.83	1.07	8.92	
36X-3, 97–98	332.1	6.46	1.04	8.67	
37X-3, 85–86	341.9	3.53	1.06	5.30	
38X-3, 58–59	351.2	1.26	1.64	13.67	–22.165
41X-1, 79–80	377.2	2.91	0.86	8.60	–22.599
42X-3, 49–50	389.5	4.28	0.88	7.33	
43X-3, 119–120	399.8	3.57	1.19	9.92	–22.226
44X-3, 64–65	408.8	5.00	0.86	7.17	
45X-3, 134–135	419.1	4.38	1.12	9.33	
46X-3, 78–79	427.9	0.13	0.68	8.50	–22.173
47X-2, 54–55	436.1	71.80	0.54	10.80	
48X-3, 72–74	447.4	1.55	1.06	8.15	–22.253
51X-3, 86–87	476.5	3.62	1.05	8.08	
52X-CC, 12–13	482.4	2.37	1.23	8.79	
53X-CC, 21–22	492.2	2.27	0.98	7.54	
54X-4, 27–28	506.5	5.11	1.21	9.31	–22.28
55X-3, 145–146	515.7	2.61	1.12	8.00	
56X-3, 23–24	524.1	2.21	1.05	8.08	–22.435
57X-1, 26–28	530.9	0.78	1.34	8.93	
60X-3, 12–14	562.6	1.55	0.97	6.93	–22.47
61X-1, 23–24	569.3	1.71	1.20	8.57	
61X-1, 24–26	569.3	1.51	1.12	8.62	
62X--3, 107–108	582.8	3.42	1.11	7.93	–23.009
63X-1, 97–98	589.4	1.77	0.94	7.23	
64X-1, 97–98	599.0	1.23	1.09	9.91	–22.188
65X-1, 97–98	608.7	2.88	0.95	7.31	
66X-5, 114–115	624.4	1.92	0.99	7.07	
68X-1, 146–147	638.0	0.56	1.24	7.75	
70X-3, 129–131	659.7	1.53	1.37	9.13	–22.274
71X-3, 73–75	668.5	1.07	0.82	6.31	
72X-3, 123–125	678.6	0.67	0.99	7.07	
73X-1, 61–63	684.6	4.28	0.82	6.31	
74X-3, 40–41	697.1	0.26	0.95	6.79	
75X-3, 98–99	707.3	0.57	0.96	7.38	
76X-3, 135–136	717.4	0.62	0.82	6.31	
186-1150B-					
1R-3, 83–84	707.0	1.29	0.68	5.67	
2R-1, 26–27	710.0	0.27	0.64	5.82	
3R-1, 30–31	719.6	0.39	0.79	6.58	–22.85
4R-1, 39–40	729.4	1.27	0.76	6.33	
5R-1, 50–52	739.1	3.97	0.97	7.46	
6R-3, 39–41	751.7	7.71	0.86	7.17	–22.534
7R-1, 30–31	758.2	2.92	0.66	5.50	
8R-1, 62–63	768.2	1.61	0.89	7.42	
9R-2, 63–64	779.3	0.82	0.80	6.67	
12R-3, 143–144	810.3	7.04	0.68	6.18	
13R-2, 122–123	818.4	1.19	0.60	6.67	
14R-3, 79–80	829.1	0.94	0.92	8.36	
15R-2, 16–17	836.7	0.74	0.62	6.20	
16R-3, 26–27	847.9	2.96	0.68	7.56	
17R-3, 138–139	858.6	1.09	0.80	6.67	
18R-3, 110–111	867.9	6.94	0.58	6.44	
19R-3, 81–82	877.3	4.45	0.82	7.45	
20R-3, 124–125	887.3	4.65	0.75	7.50	
21R-3, 79–80	896.5	1.42	0.84	7.64	
22R-3, 113–114	905.6	6.65	0.59	5.90	
23R-3, 137–138	916.2	3.86	0.78	7.80	
24R-1, 27–28	921.8	8.36	0.76	7.60	
25R-3, 36–37	934.4	9.53	0.64	7.11	
26R-2, 137–138	943.5	2.11	0.70	7.00	
27R-2, 69–71	952.5	1.02	0.76	7.60	
28R-3, 106–107	963.9	0.29	0.78	7.80	
29R-3, 84–85	973.0	1.34	0.72	8.00	
30R-3, 25–26	982.4	0.12	0.67	6.70	
31R-3, 4–6	991.6	0.36	0.64	7.11	
32R-3, 53–54	1002	0.13	0.49	6.13	
33R-1, 131–132	1009	0.36	0.63	7.00	
34R-1, 107–108	1019	0.14	0.6	6.67	
35R-3, 105–107	1031	0.14	0.62	6.89	
36R-1, 98–99	1038	0.27			
36R-3, 91–92	1041	0.81	0.63	7.00	–22.91
37R-3, 32–33	1050	1.69	0.58	6.44	
38R-1, 128–129	1058	3.02	0.52	6.50	
39R-3, 62–63	1069	0.53	0.54	7.71	–22.872
41R-1, 43–44	1086	2.33			
42R-1, 11–12	1095	0.44			
42R-5, 109–110	1102	5.82			
43R-3, 101–102	1109	0.20	0.48	6.00	–23.353
44R-1, 54–55	1115	1.10	0.47	5.88	
45R-3, 59–60	1127	0.85	0.43	5.38	
46R-3, 48–49	1137	4.80	0.79	9.88	
49R-3, 36–37	1166	8.13	0.48	6.00	

Table T2. Geochemical results for sediments at Site 1151.

Core, section, interval (cm)	Depth (mbsf)	$CaCO_3$ (wt%)	C_{org} (wt%)	C/N
186-1151A-				
2R-1, 95–96	78.95	11.47	1.19	9.18
3R-3, 63–64	87.33	6.39	1.30	8.15
4R-3, 71–72	97.01	0.25	1.31	8.19
6R-1, 64–65	113.24	0.79	0.68	7.50
7R-1, 90–91	123.10	4.90	0.22	2.47
8R-1, 98–99	132.78	0.90	0.78	7.11
9R-3, 81–82	145.21	1.57	1.00	7.71
10R-5, 80–81	157.80	0.76	1.03	7.91
11R-5, 112–113	167.72	0.37	1.00	7.66
12R-5, 60–61	176.80	15.13	0.62	6.93
13R-3, 90–91	183.70	4.63	1.11	7.96
14R-3, 51–52	193.01	1.76	1.19	7.92
15R-3, 68–69	202.88	3.22	0.88	8.04
16R-3, 72–73	212.62	5.53	0.78	7.77
17R-3, 68–69	222.18	0.13	1.02	8.54
18R-5, 91–92	235.01	3.70	1.11	8.51
20R-1, 91–92	248.21	0.44	1.04	7.41
21R-3, 88–89	260.78	1.07	1.02	7.29
22R-3, 91–92	270.51	2.67	1.04	8.00
23R-3, 63–64	279.83	0.23	0.86	7.18
24R-3, 48–50	289.38	1.41	0.95	7.31
25R-3, 112–114	299.72	0.18	1.08	7.70
26R-1, 4–6	305.34	0.13	0.78	4.90
27R-1, 7–9	314.97	2.22	1.05	7.53
28R-3, 61–63	328.21	3.25	1.06	7.57
29R-3, 52–54	337.72	1.60	1.16	7.72
30R-1, 48–50	344.38	3.19	1.10	7.32
31R-1, 10–12	353.70	6.91	1.04	7.43
32R-1, 134–136	364.54	1.53	1.05	7.47
33R-3, 93–95	376.83	2.46	1.01	6.76
35R-1, 25–27	392.35	3.99	1.18	7.88
36R-3, 54–56	404.94	1.85	1.28	7.98
38R-1, 135–136	421.95	4.76	0.75	6.80
39R-3, 34–35	433.34	3.31	0.86	6.16
40R-4, 21–22	444.61	1.22	1.05	7.52
42R-1, 23–25	459.43	4.89	0.98	6.55
43R-3, 105–106	472.85	1.61	0.97	6.90
44R-5, 127–128	485.67	1.52	1.13	8.05
46R-1, 55–56	498.15	2.18	1.19	7.92
47R-1, 145–146	508.65	3.64	1.05	7.52
51R-1, 17–18	545.87	1.45	0.89	7.39
52R-3, 111–112	559.51	3.65	1.09	7.80
54R-1, 75–76	575.45	3.10	0.91	6.98
56R-1, 67–68	594.57	9.15	0.70	5.85
58R-3, 83–84	617.03	3.86	0.69	5.72
59R-3, 28–29	626.18	7.15	0.90	6.94
60R-3, 35–36	635.85	1.54	0.70	4.97
61R-3, 79–81	645.89	0.25	0.85	4.25
63R-1, 28–30	661.48	0.30	0.65	5.95
66R-3, 12–13	692.92	3.89	0.72	6.03
68R-1, 120–121	710.20	7.79	0.50	5.04
69R-3, 90–92	722.20	3.28	0.99	7.58
71R-1, 21–22	738.21	3.39	0.52	4.36
72R-1, 58–59	748.28	2.51	0.75	5.75
73R-1, 23–24	757.53	2.66	0.81	6.24
74R-1, 122–123	768.12	2.05	0.67	5.19
75R-1, 88–89	777.38	3.41	0.72	6.55
77R-1, 128–129	796.98	2.39	0.54	5.44
78R-3, 120–122	809.60	2.37	0.47	4.66
80R-3, 125–126	828.85	0.87	0.89	6.81
81R-1, 84–86	835.04	0.24	0.69	6.28
82R-5, 24–26	850.04	0.12	0.86	7.13
84R-5, 114–115	870.24	0.10	1.14	7.58
85R-5, 94–95	879.74	0.16	1.39	11.59
87R-1, 32–33	892.22	0.68	0.47	4.68
88R-3, 83–84	905.43	0.52	0.57	5.68
89R-3, 143–144	915.63	1.53	0.46	4.56
90R-3, 29–30	924.09	2.18	0.67	6.07
92R-5, 25–26	945.98	4.37	0.54	5.35
94R-1, 87–88	960.17	0.36	0.44	4.86
96R-1, 11–13	978.71	4.88	0.64	6.44
97R-3, 130–132	992.50	0.94	0.46	2.86
98R-1, 99–100	998.79	4.98	0.48	3.21
100R-1, 69–70	1017.69	0.10	0.43	2.85
102R-1, 110–111	1037.30	0.45	0.42	4.63
104R-1, 72–73	1056.32	0.13	0.42	3.86