Enhanced supply of fossil organic carbon to the Okinawa Trough since the last deglaciation

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Received 23 February 2007; revised 28 December 2007; accepted 7 February 2008; published 7 May 2008.

[1] Significantly older ^{14}C ages by 2500–7900 years are found for sedimentary total organic carbon (TOC) when compared to ages of codeposited surface-dwelling foraminifera in the southern Okinawa Trough. This age discrepancy increases with rising sea level since the Last Glacial Maximum. A progressive shift in TOC $\delta^{13}C$ toward more negative values with rising sea level reflects an increasing fractional contribution of terrestrial organics (soil organics, plant debris, and/or fossil organics) to the buried organic pool. Organic matter previously stored on the East China Sea shelf during sea level lowstand and riverine material from Taiwan may be the sources that cause the $\delta^{13}C_{TOC}$ to shift to more terrestrial values. During the Holocene when sea level is above −40 m, $\delta^{13}C_{TOC}$ values stabilize within a narrow range (−22.3 to −22.8‰) while age discrepancies continue to increase and less chemically weathered sediments are deposited. The increase in age discrepancy between TOC and foraminifera in the Holocene may be due to a wetter climate that drove higher rates of physical weathering on Taiwan and greater transport rates of fossil organic C-bearing lithogenic sediment to the ocean. The climate impact on the relative delivery of fossil and nonfossil TOC in depositional settings influenced by fluvial sources should be considered in interpretations of sedimentary C isotope records.

Citation: Kao, S. J., M. H. Dai, K. Y. Wei, N. E. Blair, and W. B. Lyons (2008), Enhanced supply of fossil organic carbon to the Okinawa Trough since the last deglaciation, *Paleoceanography*, *23*, PA2207, doi:10.1029/2007PA001440.

1. Introduction

[2] The burial and exhumation cycle of organic carbon moderates global atmospheric CO_2 and O_2 levels over geologic timescales [*Hedges*, 1992; *Petsch et al.*, 2000]. Recently, small mountainous rivers that drain tectonically active margins, particularly those in the Oceania region, have received attention because of their significant sediment and associated particulate organic carbon (POC) fluxes to the oceans [*Milliman and Syvitski*, 1992; *Lyons et al.*, 2002]. Unlike large rivers that appear to discharge predominantly modern POC [*Hedges et al.*, 1986], small mountainous rivers can discharge abundant fossil POC [*Kao and Liu*, 1996; *Blair et al.*, 2003; *Komada et al.*, 2004; *Leithold et al.*, 2006]. The fossil organic matter exists primarily in the form of an insoluble, unreactive, high-molecular-weight material known as kerogen [*Leithold and Blair*, 2001; *Blair et al.*, 2003]. The estimated flux of fossil POC from rivers (>40 Tg C/a [*Blair et al.*, 2003]) may account for 25% of the global POC burial in marine sediments (160 Tg C/a [*Hedges and Keil*, 1995]), and most of it is likely from the small mountainous rivers of Oceania.

[3] Globally, as much as 90% of riverine sediments are deposited on the continental margin; thus this zone plays a key role in linking the global terrestrial and oceanic carbon cycles [*Thomas et al.*, 2004; *Deng et al.*, 2006]. The East China Sea (ECS) is arguably the most important marginal sea in the western Pacific as it is the interface between the world's largest continent (Asia-Europe) and the Pacific Ocean (Figure 1). Historically, 10% of the global fluvial sedimentary flux has been delivered to the ECS via two large rivers, i.e., the Changjiang and the Huanghe [*Milliman and Meade*, 1983], though recently the contribution from the Huanghe decreased significantly because of upstream water redistribution [*Yu*, 2006]. The annual sediment output from small Taiwanese rivers is ∼184 to 380 Mt/a [*Dadson et al.*, 2003; *Kao et al.*, 2005a], which is comparable to the sediment flux from the largest river, the Changjiang in China (500 Mt/a [*Milliman et al.*, 1985]). These large suspended sediment inputs significantly impact the biogeochemical cycling of carbon and other biogenic elements in the East China Sea and surrounding seas [*Kao and Liu*, 1996; *Lyons et al.*, 2002; *Kao et al.*, 2006a]. Most terrestrial material discharged from the Changjiang and the Huanghe rivers is deposited along the coast of China, a process that began around 7 calendar (cal) ka B.P. [*Liu et al.*, 2004, 2006a, 2006b]. Taiwanese rivers therefore provide a more direct supply of sediment to the southern Okinawa Trough.

[4] The Okinawa Trough (OT) along the outer edge of the ECS continental shelf is the ultimate deposition site for autochthonous organics produced on the shelf and allochthonous organics discharged from rivers in China and Taiwan

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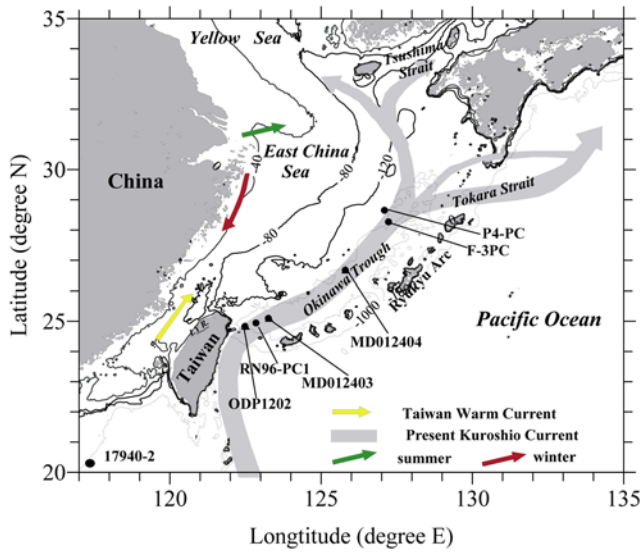

Figure 1. Location map for the International Marine Global Changes Study program (IMAGES VII) core MD012403. Core sites for MD012404, RN96-PC1, ODP1202, P4-PC, F-3PC, and 17940-2 are indicated. The −40, −80, −120, and −1000 m isobaths are shown. The flow path of the Kuroshio Current (gray), Taiwan Warm Current (yellow), summer (green), and winter (red) patterns of Changjiang material transportation are indicated.

(Figure 1). The Kuroshio Current (KC) plays a major role in dispersing sediments and organics on the ECS shelf and into the Okinawa Trough. The KC enters the OT from the south flowing to the northeast and ultimately bifurcating into the Yellow Sea and the Sea of Japan (Figure 1). In this paper, correlations between climate-influenced terrestrial inputs, sea level fluctuations, and organic carbon burial are presented. The supply of fossil organic matter from Taiwanese rivers to the Okinawa Trough appears to be enhanced in the Holocene. We separate total organic carbon into two components, fossil organic carbon (FOC) and contemporaneously deposited organic carbon (COC), by using ^{14}C activity in total organic carbon and a bimodal mixing model. We conclude that climate change apparently affects the supply and burial of fossil organics in the oceanic realm, particularly in seas surrounding high-standing islands.

2. Materials and Methods

[5] The total organic carbon (TOC) and its isotopic compositions ($^{13}C/^{12}C$ and $^{14}C/^{12}C$) are reported here from a 20 m piston core that penetrated down to the Last Glacial Maximum (LGM) in the southern Okinawa Trough. The giant piston core MD012403 was taken onboard R/V *Marion Dufresne* on 28 May 2001 at 25.07°N, 123.28°E (Figure 1) from a water depth of 1420 m during WEPAMA cruise of the International Marine Global Changes Study program (IMAGES VII) [*Bassinot et al.*, 2002].

[6] All samples were rinsed with Milli-Q water to remove salts and were freeze-dried. Details of the sample preservation and pretreatments have been reported by *Kao et al.* [2005b]. For organic carbon isotope analysis, 1 N HCl was applied for 16 h to remove carbonates. This procedure would not have removed recalcitrant authigenic carbonates such as siderite. The samples were separated from the supernatant via centrifugation and then freeze-dried [*Kao et al.*, 2003]. Some loss of organic carbon (OC) may have ensued as a result of rinsing and centrifugation, though for samples of this age and diagenetic state the fraction will be small. The supernatant solution was analyzed for acid extractable calcium (Ca_A) (section 2). The acidified sediments were measured for carbon isotopic composition by using a Carlo-Erba Flash EA 1112 elemental analyzer connected to a Thermo Finnigan Delta^{plus} Advantage isotope ratio mass spectrometer. Carbon isotopic compositions are presented in the standard δ notation with respect to Peedee belemnite. U.S. Geological Survey standard 40, which has a certified $\delta^{13}C$ value of −26.24‰, was used as a working standard. The accuracy and precision of carbon isotopic determination are better than 0.2‰.

[7] Aluminum (Al), total calcium (Ca_T), acid extractable Ca (Ca_A), potassium (K), and sodium (Na), which were analyzed by using an ICP-OES (Optima 3200DV, Perkin-Elmer^{TM} Instruments, Waltham, Massachusetts, United States), were utilized to assess the state of chemical weathering to elucidate changes in the sources of lithogenic particles. The total digestion method has been reported by *Hsu et al.* [2003]. According to *Nesbitt and Young* [1989] the chemistry of typical weathering of upper crustal rocks can be approximated by the weathering of feldspar and volcanic glass, which make up 75% of labile material in the upper crust. Much of the chemical variation resulting from weathering may be expressed in the system $Al_2O_3 - (CaO^* - Na_2O) - K_2O$, where CaO* represents the Ca in the silicate fraction only. Therefore, the chemical index of alteration (CIA) can be defined as

$$\mathrm{CIA} = \left[\frac{\mathrm{Al}}{\mathrm{Al} + \mathrm{Ca}^* + \mathrm{K} + \mathrm{Na}}\right] 100. \qquad (1)$$

[8] A correction to the measured Ca content for the presence of Ca in carbonates was made (i.e., $Ca^* = Ca_T - Ca_A$). These chemical signatures of lithogenic particles on land are ultimately transferred to the sedimentary record which provides a useful tool for monitoring source area weathering conditions, and therefore they are widely used to interpret the weathering history of modern and ancient sediments [e.g., *Nesbitt and Young*, 1982; *Wronkiewicz and Condie*, 1987]. For example, high CIA values reflect removal of labile cations (Ca, Na, and K) relative to stable residual constituents (Al) during weathering, while low CIA values indicate the near absence of chemical alteration and might reflect cool and arid conditions. An assumption is made that the chemistry of the particles is not significantly altered by marine processes during transport and burial.

[9] To develop a geochronology of the core, ^{14}C measurements were made on foraminifera from 12 depth intervals. More than 300 specimens of the most abundant planktonic foraminifera (*Globigerinoides ruber* and *Orbulina universa*) were picked exhaustively from the >250 μm size fraction,

Table 1. Conventional ^{14}C Ages for Foraminifera and Total Organic Carbon Samples as a Function of Depth in MD012403[a]

Depth, m	^{14}C Age Foraminifera, years B.P.	Weight, mg	$\delta^{13}C$ (Foraminifera), ‰	^{14}C age TOC, years B.P.	Age Differences (TOC-Foraminafera)	pC_{oc}[b], %	Calibration Age, cal years B.P.
0.26–0.27	797[c]	NA	NA	6,987 ± 35	6,190	46	NA
0.31–0.32	1,048 ± 45	6.571	2.25				588
2.41–2.42	3,244 ± 45	7.135	2.12	11,139 ± 50	7,895 ± 95	37	3,026
3.71–3.72	4,432 ± 45	7.397	1.82				4,570
4.91–4.92	6,164 ± 55	8.306	2.55	13,557 ± 60	7,390 ± 115	40	6,559
6.01–6.02	8,211 ± 50	6.915	1.60				8,725
7.61–7.62	9,553 ± 55	6.904	1.67				10,372
9.51–9.52[d]	9,557 ± 70	2.906	1.42	13,989 ± 60	4,436 ± 130	57	10,373
10.60–10.62	10,111 ± 55	6.938	1.60				11,040
11.41–11.42	10,312 ± 45	5.329	2.03				11,275
13.51–13.52	11,325 ± 55	5.495	2.09	15,159 ± 70	3,834 ± 125	62	12,872
16.71–16.72	15,867 ± 65	4.541	0.91	18,368 ± 90	2,501 ± 155	73	18,763
19.61–19.62	17,936 ± 75	6.804	0.38				20,705
19.91–19.92	18,000 ± 110	6.696	1.03	20,660 ± 110	2,660 ± 220	71	20,775

[a]The $\delta^{13}C$ (foraminifera) was obtained during ^{14}C measurement and used for ^{13}C correction. Species are a mixture of *Globigerinoides ruber*, *Globigerinoides sacculifer*, *Globigerinoides conglobatus*, *Globigerina ququilateralis*, and *Orbulina universa*. NA indicates not available.
[b]Percent contemporaneous OC, defined as TOC formed near the time of deposition (see section 2).
[c]Extrapolated foraminifera age (see text).
[d]Problematic because of insufficient amount of specimen, yet, real age should fall within a small range without significant influence.

which is relatively less mobile after deposition. Problems in dating that may result from varying abundance, bioturbation, and resuspension should thus be attenuated [see *Mollenhauer et al.*, 2005]. The same approach was applied for ^{14}C measurements in Ocean Drilling Program (ODP) core 1202 nearby [*Wei et al.*, 2005]. The ^{14}C ages thus obtained were consistent with trends of planktonic foraminiferal $\delta^{18}O$ values and coccolith $U_{37}^{K'}$ parameters that vary synchronously with ice volume changes [*Wei et al.*, 2005; *Zhao et al.*, 2005].

[10] Sediment from seven depth intervals was decarbonated with 0.1 N HCl for ^{14}C measurements (Table 1). The samples were analyzed by the Rafter Radiocarbon Laboratory, National Isotope Centre of the Institute of Geological and Nuclear Sciences in New Zealand. Conventional ^{14}C ages are presented here. Those ages were corrected for isotopic fractionations that had occurred during the formation of the dated material or during sample processing and measurement [*Stuiver and Polach*, 1977]. Thus, despite the different $\delta^{13}C$ values of organic matter and carbonates, this convention allows for presumably direct comparisons between the radiocarbon contents of foraminifera and TOC, and accordingly, apparent age differences can be calculated. Conventional ^{14}C ages were converted to "calendar ages" by employing the CALIB 5.0 program by M. Stuiver et al. (available at http://calib.qub.ac.uk/calib) and a 400 year surface ocean reservoir correction [*Bard*, 1988].

[11] TOC and total sulfur (TS) were measured using an HORIBA model EMIA-220V C/S analyzer operating at 1350°C. Additional TOC and TS data from *Kao et al.* [2005b] and $\delta^{13}C_{TOC}$ values from the core MD012404 [*Kao et al.*, 2006b, Figure 1] were pooled with the results from this study for discussion. For a comparison between sites in the trough, three samples from the core MD012404 were analyzed for TOC ^{14}C content. The age model for core MD012404 has been detailed by *Chang et al.* [2005], and apparent ^{14}C age offsets between TOC and foraminifera were determined.

[12] The fractional contributions of FOC and more modern organics to sedimentary TOC were estimated using a bimodal mixing model. The primary assumption made is that the TOC is dominated by two pools with distinctive radiocarbon compositions. In systems that are fed by small, high-sediment-yielding watersheds, this is not an unreasonable assumption [*Leithold et al.*, 2006]. The fossil organics consist of dead carbon (no ^{14}C activity), while the modern organic C is defined as TOC formed near the time of deposition (contemporaneous OC); therefore, the "modern" component ages at the same rate as codeposited foraminifera and is assumed to have the same ^{14}C age as the codeposited foraminifera. Since the ^{14}C activity in buried TOC is a mixture of ^{14}C in contemporaneous OC and fossil carbon (dead carbon), the age offset between TOC (apparent age) and foraminifera (absolute age) at any interval can thus be used to calculate the relative contribution of the contemporaneous OC (pC_{oc}, percent contemporaneous OC) by using the equation

$$pC_{oc}(\%) = 100\{\exp[-(\text{age offset} + (2001 - 1950)/1.03)/8033]\}. \quad (2)$$

[13] Equation (2) is modified following the equation for "percent modern" by *Stuiver and Polach* [1977]. The year of age measurement is 2001. Accordingly, we calculated the pC_{oc} values for those samples possessing the necessary ^{14}C measurements and then estimated pC_{oc} values for all other samples in between by linear interpolation. The pC_{oc} values were thus used to resolve the TOC into the FOC and COC components. Unfortunately, there is insufficient data to perform the same calculation for MD012404 (section 3).

3. Results

[14] The $\delta^{13}C$ values for foraminifera, as well as the measured ^{14}C ages of both foraminifera and sedimentary TOC, are shown in Table 1. The temporal variations in

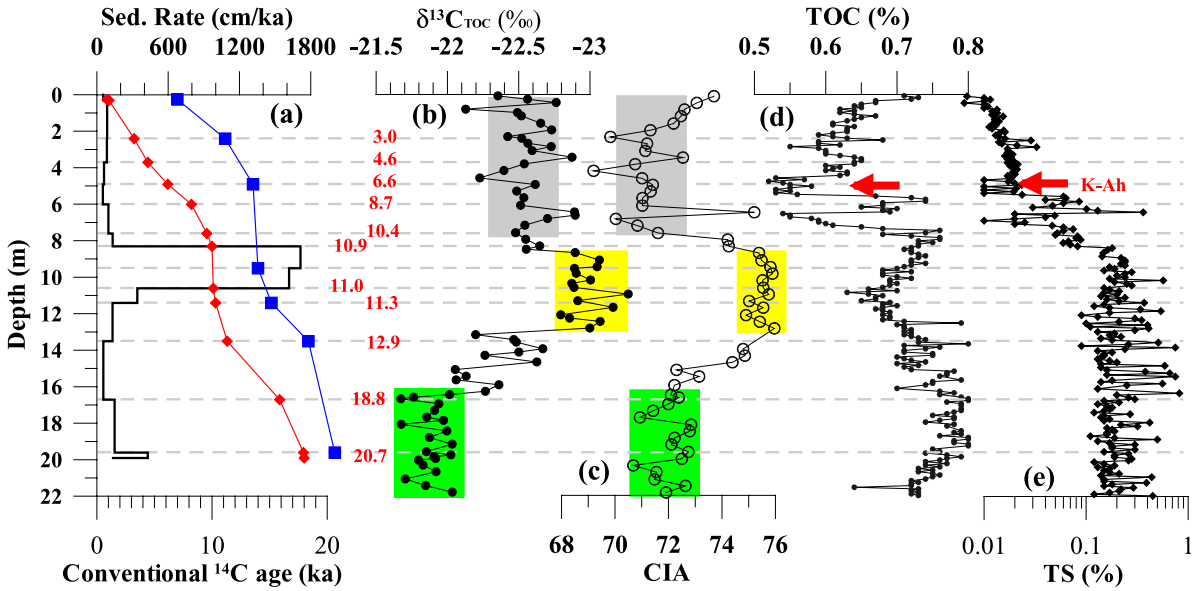

Figure 2. Depth profiles for (a) sedimentation rate showing ^{14}C age of surface dwelling foraminifera (red diamonds) and ^{14}C age of total organic carbon (blue squares), (b) $\delta^{13}C_{TOC}$, (c) chemical index of alteration (CIA) (see text), (d) total organic carbon content (TOC) (%), and (e) content of total sulfur (TS) (%). Dashed lines and red numbers represent calibrated calendar age for foraminifera samples. The red arrows indicate K-Ah volcanic layers. Gray, yellow, and green fields in Figures 2b and 2c are the same as in Figure 4 (see text).

foraminiferal $\delta^{13}C$ values are consistent with those for the nearby ODP1202 core [*Wei et al.*, 2005]. Agreement between the two data sets suggests that the carbon isotopic compositions, including those of ^{14}C, have not been altered by dissolution/precipitation. Barring extensive lateral transport, the foraminiferal ^{14}C ages should thus represent the absolute ages of their corresponding depth intervals.

[15] The oldest ^{14}C date of 18.0 ka (20.7 cal ka B.P.) marks the LGM at ~19.9 m (line with red diamonds, Figure 2a). The age for the near-surface layer (0.26 m, Table 1) was determined by linear interpolation between the age at 0.31 m and a modern age for the surface. Calculated sedimentation rates of different time intervals ranged from 50 to 200 cm/ka (with one exception around 9 m, see Figure 2a). Sedimentation rates are lower in the LGM and are higher between 8.7 and 12.9 cal ka B.P. (>100 cm/ka) when sea level rose rapidly (Figure 3a). An interval with extremely rapid sedimentation (~1700 cm/ka) occurred between 10.9 and 11.0 cal ka B.P. (8.31–10.61 m), which marks the beginning of the Holocene. Summer monsoonal precipitation reached the maximum at latitude 25°N in southern China at this time [*Yuan et al.*, 2004]. Nearby core RN96-PC1 [see *Ujiié and Ujiié*, 1999, Figure 1] reveals, for the same timing, a significant increase in the proportion of warm water foraminiferal species, which implies that the inflow of the surface Kuroshio Current intensified [*Xu and Oda*, 1999; *Ujiié and Ujiié*, 1999; *Ujiié et al.*, 2003], transporting greater heat and moisture to East Asia in the Holocene. Sea level effects on surface and deep circulation have been examined with a 3-D model by *Kao et al.* [2006c], in which they found an enhancement of the KC throughflow and deepwater ventilation in the Okinawa Trough in the Holocene. Oxygen supply to the deep water should have been elevated in response. Independent evidence shows that a transformation from foliation to anomalous sedimentary magnetic fabric, which indicates an enhancement of deep-water circulation, has occurred as the KC has intensified since the Holocene [*Kao et al.*, 2005b].

[16] Apparent ^{14}C ages measured for codeposited TOC range from 6987 to 20,660 years B.P. (Table 1 and Figure 2a). Large age offsets between TOC and foraminifera are evident throughout the core. The minimum offset of 2.5 ka (Table 1) occurs during the LGM, whereas the maximum offset of 7.9 ka appears in the Holocene. Age offsets between TOC and foraminifera have been reported by *Mollenhauer et al.* [2005] in four depositional environments under different climatic conditions with various aeolian and fluvial inputs. However, their maximum age offset never exceeded 3.0 ka, and age discrepancies barely changed with sea level fluctuates. In contrast, the age differences increase in our core in concert with the rising sea level since the Last Glacial Maximum (Figure 3c).

[17] The $\delta^{13}C_{TOC}$ values (Figure 2b) vary from −23.1 to −21.6‰ with a mean value of −22.2 ± 0.3‰ (n = 45), most likely reflecting a mixture of terrestrial (a typical end-member value of ~−26‰ [e.g., *Wada et al.*, 1987]) and marine organics (with a typical end-member of ~−20.0‰ [*Goericke and Fry*, 1994]). In Taiwan, the terrestrial end-member is ~−25.5‰, representing a mixture of predominantly C3 vascular plants, soil, and kerogen [*Kao and Liu*, 2000]. Contributions from C4 plants ($\delta^{13}C$ ~ −9 to −13‰) cannot be discounted entirely, especially during the LGM when the climate was drier on Taiwan [*Wei et al.*, 2003]. However, the offshore trend in $\delta^{13}C$ between cores is most

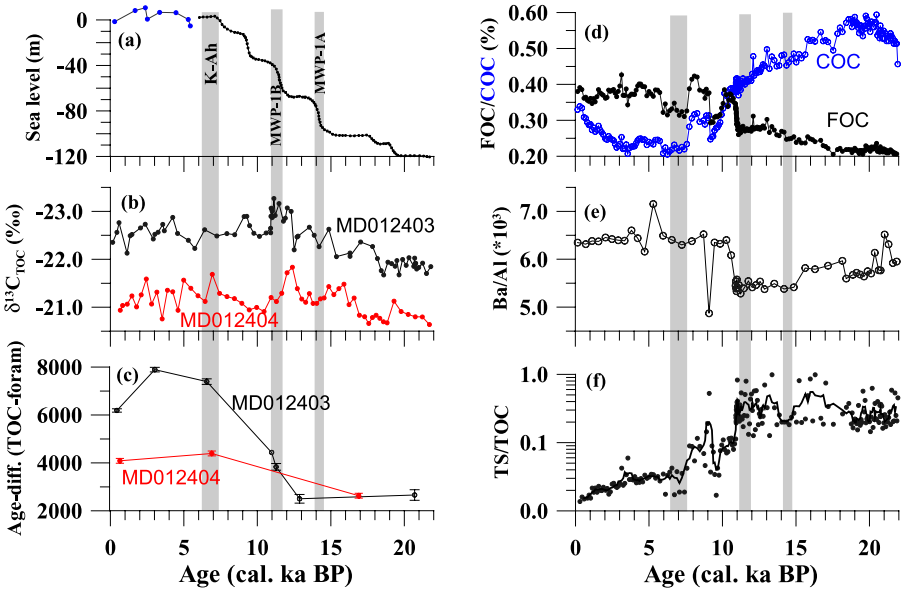

Figure 3. (a) Sea level curve. Blue dotted line and black curve represent data digitized from *Saito et al.* [1998] and *Liu et al.* [2006b], respectively. Gray columns through six panels represent MWP-1A, MWP-1B, and K-Ah volcanic ash layers, respectively. (b) Down-core trends of $\delta^{13}C_{TOC}$ value for MD012403 (solid black circles) and MD012404 (solid red circles) (data from *Kao et al.* [2006b]). (c) Age difference between total organic carbon and foraminifera of selected samples from MD012403 (black circles) and MD012404 (solid red circles). Error bars are also shown. (d) Fossil organic carbon (FOC) (solid black circles) and contemporaneous organic carbon (COC) (blue circles) contents (in %) calculated based on bimodal mixing (see text). (e) Bulk Ba/Al ratios from *Kao et al.* [2005b]. (f) Ratio of TS to TOC. TS data are taken from *Kao et al.* [2005b].

consistent with dilution of terrestrial C3-derived material by marine organics and thus will be treated as such.

[18] The $\delta^{13}C_{TOC}$ increase by 1.5‰ upward in the core reflects an increase of the fractional contribution from the terrestrial end-member. It reaches the minimum at the beginning of the Holocene and then increases by ~0.5‰ afterward and finally stays within a narrow range from −22.3 to −22.8‰. The $\delta^{13}C_{TOC}$ value for the surface sample agrees well with those values found in modern surface sediments in the southern Okinawa Trough [*Kao et al.*, 2003]. Within the time interval of sea level progression, several steps of rapid sea level rise (Figure 3a) are observed. According to *Fairbanks* [1989], the meltwater pulse-1A and meltwater pulse-1B (MWP-1A and MWP-1B in Figure 3a) events correspond to two periods of rapid sea level rise in the East China Sea that may have created rapid land area submergence. In core MD012403, one $\delta^{13}C_{TOC}$ peak is found corresponding to the MWP-1B (Figure 3b). In core MD012404, two $\delta^{13}C_{TOC}$ peaks appear just prior to the two meltwater pulse events. Similarly, a core taken from the slope on the Sunda Shelf in the South China Sea reveals a signal of enhanced terrestrial organic input prior to MWP-1A [*Steinke et al.*, 2003].

[19] CIA values fall into three distinguishable zones (Figure 2c). There are relatively low values around the LGM, a high plateau around 13 cal ka B.P., followed later by low but variable values. The CIA transition pattern prior to the Holocene (below 8 m) coincides with that of $\delta^{13}C_{TOC}$.

[20] TOC contents (Figure 2d) vary between 0.5 and 0.8%. Mean TOC concentrations are higher in pre-Holocene deposits (0.72 ± 0.4) relative to Holocene sediments (0.64 ± 0.5). Low TOC concentrations are found at ~6.6 cal ka B.P. and ~9 cal ka B.P. (Figure 2d), which corresponds to periods that exhibit relatively high values of bulk magnetic susceptibility [*Kao et al.*, 2005b]. Significant reductions in TOC content are also observed around 6.8 to 6.3 cal ka B.P. in P4-PC and F-3PC at the northern trough (core locations shown in Figure 1) [*Oguri and Matsumoto*, 2000], and this TOC depletion zone was caused by dilution by the Kikai-Akahoya (K-Ah) volcanic ash (from Kikai-Akahoya eruption in Japan). The age of the K-Ah layer (6.3 ka B.P.) is consistent with the ^{14}C chronology of our core, adding confidence to our age determination. The cause for the TOC depletion around 9 cal ka B.P. in our core is unknown, and no related events have been reported from nearby cores.

[21] TS contents vary by over an order of magnitude from ~0.01 to ~0.5% (Figure 2e). Variable but persistently higher TS contents (mostly >0.1%) occur below 8 m, where framboidal pyrite was frequently identified. A progressive decrease in TS contents started ~10 cal ka B.P. and continues to the present. Similar low TS values in the middle Okinawa Trough have been interpreted to have resulted from low rates of organic matter oxidation via sulfate reduction [*Kao et al.*, 2006b]. Alternatively, low TS storage may reflect reoxidation of sulfide minerals via bioturbation and/or physical processes [*Aller and Blair*, 1996; *Sommerfield et al.*, 2001]. Either situation might be

associated with enhanced deepwater circulation due to Kuroshio inflow intensification [*Kao et al.*, 2006b]. Sulfate reduction rates could also be lowered if the reactivity of deposited TOC decreased [*Berner*, 1984].

4. Discussion

[22] Climate and sea level changes influence the delivery of terrestrial material to the ocean, and the magnitude of change may have been influenced by either freshwater fluxes and/or distance between land and deposition site. The average sedimentation rates for ODP1202 [*Wei et al.*, 2005], MD012403, MD012404 [*Chang et al.*, 2005], and P4-PC (or F-3PC) [*Oguri and Matsumoto*, 2000], from the south to the north (Figure 1), during the last 20 ka are ~4, ~1, ~0.5, and ~0.2 m/ka, respectively. The decreasing trend in sedimentation rate with increasing distance from Taiwan suggests that fluvial material from Taiwanese rivers is a significant source of the sediment deposited in the Okinawa Trough during the past 20 ka. For all five cores, higher sedimentation rates occur during the deglaciated period from ~10 to 15 cal ka B.P. In core ODP1202, which is the closest to Taiwan, the sedimentation rate peaks during transgression, and the sedimentation in the Holocene is significantly higher than that during the Last Glacial Maximum. The overall temporal and spatial sedimentation patterns in the Okinawa Trough reveal that sediment delivery to the Okinawa Trough was not enhanced in response to the lowering of sea level and the exposure of nearshore sediments. Instead, sedimentation increased during the more humid Holocene, presumably as a result of a greater fluvial sediment transport. Wet conditions contribute to landsliding, gullying, and bank erosion [*Trustrum et al.*, 1999; *Hartshorn et al.*, 2002]. Fluvial sediment transport increases nearly exponentially with increasing water discharge [e.g., *Kao and Liu*, 2001; *Kao et al.*, 2005a].

[23] The mean TOC content in ODP1202 in the Holocene period is 0.58 ± 0.2% [*Chang et al.*, 2005], which is comparable to that observed in MD012403 (0.67 ± 0.1%) and MD012404 (0.64 ± 0.1%) within the same time interval. The mean TOC content is similar to the baseline value of 0.5% for particulate organic carbon reported for the Lanyang River (Figure 1) during high-volume floods [*Kao and Liu*, 1996] and for other high sediment-yielding, small mountainous rivers [*Komada et al.*, 2004; *Leithold et al.*, 2006]. Much of the particulate OC from such systems has been attributed to fossil POC derived from the mass wasting of bedrock [*Leithold et al.*, 2006]. Accordingly, much of the TOC delivered to the Okinawa Trough could be composed of a significant quantity of FOC from the mountain belt in Taiwan. The importance of FOC should be relatively higher in the southern trough because of the proximity to source; that is, higher FOC concentrations are expected for the more proximal MD012403 site relative to MD012404.

[24] Sea level rise–induced shelf submergence may have played an important role in regulating the supplies of lithogenic particles and organics from the East China Sea shelf to the deep sea. By using sedimentary $\delta^{13}C_{TOC}$ values of the two cores from the northern trough (P4-PC and F-3PC, locations noted in Figure 1), *Oguri and Matsumoto* [2000] argued for offshore transport of terrestrial organic matter that was induced by rapid shelf submergence during early deglaciation. Sedimentation rates were enhanced at the northern Okinawa Trough during the transgression stage. To determine if this phenomenon occurred in the middle and southern Okinawa Trough, sea level curves from *Saito et al.* [1998] and *Liu et al.* [2006a] and $\delta^{13}C_{TOC}$ data were examined.

[25] Sea level reached its present level around 8 cal ka B.P. (Figure 3a). From that time on, the circulation pattern resembles that of today [*Kao et al.*, 2006c]. Notably, the $\delta^{13}C_{TOC}$ values do not trend toward the terrestrial end-member at lowstand (Figure 3b) even though the landmass extended closer to the Okinawa Trough. As indicated in section 4, fluvial sediment discharge to the ocean may have been relatively low during the LGM as revealed by sedimentation rates in the Okinawa Trough, and this may be the result of decreased erosion rates and/or increased sediment trapping within the watershed.

[26] The down-core trend of $\delta^{13}C_{TOC}$ presented here resembles that reported by *Kienast et al.* [2001] for the northern shelf of the South China Sea (core 17940-2, see Figure 1). This trend was unexplained, but changes due to variations in pCO_2 over time or the influx of terrestrial plant organic matter appear to have been ruled out [*Kienast et al.*, 2001]. We hypothesize that even though such factors as variations in pCO_2 and precipitation over the glacial-interglacial transition may have influenced the marine and terrestrial plant end-member $\delta^{13}C$ values [*Kienast et al.*, 2001; *Gagen et al.*, 2006], the dominant source of the changes in $\delta^{13}C_{TOC}$ is variability in the relative flux of FOC delivered to the seabed. If true, this observation serves as a serious cautionary note for those attempting to use organic carbon $\delta^{13}C$ values as indicators of pCO_2 or marine/terrestrial plant OC ratios in regions influenced by fluvial inputs.

[27] The low CIA values during the LGM are attributable to a dry and cold climate on land, whereas the high CIA values during the period of rapid sea level rise may result from the remobilization of weathered sediments stored in coastal plains or deltas. Our high CIA values (74–76) during deglaciation are similar to the mean value of 74 reported for the Changjiang River [*Li et al.*, 1984]. During the warmer and more humid Holocene [*Yuan et al.*, 2004], CIA values drop to an unexpectedly low level similar to that in the LGM (Figure 2c). The low CIA values in the humid Holocene may reflect a predominance of physical over chemical weathering on land because of the highly erodible nature of the landscape, which results in rapid denudation rates [*Kao and Liu*, 1996; *Carey et al.*, 2005; *Lyons et al.*, 2005; *Selvaraj and Chen*, 2006]. Elevated contributions of less weathered materials in the Holocene from Taiwan to the deposition site of core MD012403 can lower CIA values and amplify the age offsets between TOC and foraminifera because of the addition of FOC. Yet, enhanced inputs of the low-CIA material have only a small effect on $\delta^{13}C_{TOC}$ values because the FOC in the fluvial material from Taiwan has a $\delta^{13}C_{TOC}$ value of −25.5‰ [*Kao and Liu*, 2000], resembling the typical terrestrial end-member. Lower age offsets at low sea level conditions are attributable to the

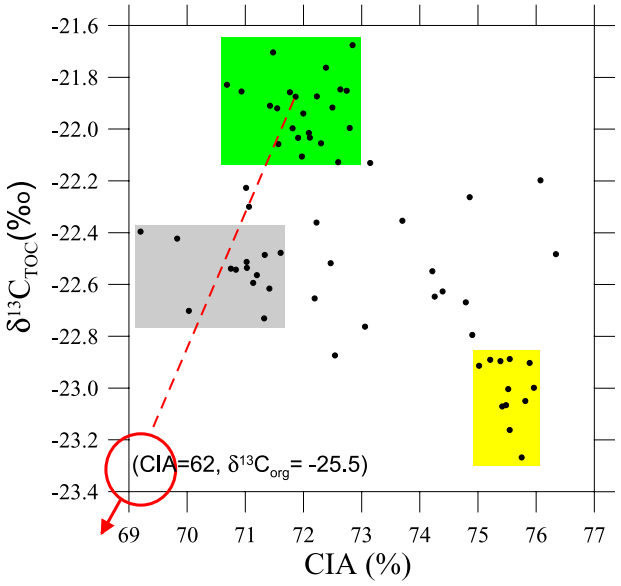

Figure 4. The scatterplot for $\delta^{13}C_{TOC}$ versus CIA values from MD012403. Yellow, gray, and green areas represent mixed terrestrial and marine organic carbon (OC) sources associated with highly weathered sediments, a more predominantly fossil OC source with less weathered sediments, and a marine-dominated OC source associated with intermediately weathered material, respectively. A potential end-member source derived from Taiwanese rivers with a CIA value of 62 and a $\delta^{13}C_{TOC}$ value of −25.5‰ is marked by a red circle. The dashed line is for the proposed possible mixing trajectory between the marine and riverine sources.

relatively lower supply of fossil organics and/or terrestrial organics that are preaged on land. Age offsets of the three selected samples from the middle trough core MD012404 show a similar trend to that revealed in MD012403 (Figure 3c), though the age offsets are smaller. Distance from source evidently attenuates the effect of FOC input. The temporal variations of FOC and COC calculated for the core MD012403 are shown in Figure 3d. The estimated FOC is ∼0.2% on a sediment dry weight basis (∼30% of the TOC) during the LGM and increases to ∼0.4% (∼60% of the TOC) during sea level highstand. The COC shows the opposite trend. The increase in FOC burial can be directly attributed to greater erosion rates on land in Taiwan during wetter periods, resulting in a higher flux of sedimentary rock organic carbon to the Okinawa Trough. The drivers behind the COC trend are less obvious. As indicated by Ba/Al ratios (Figure 3e) [*Kao et al.*, 2005b], primary production was high in the Holocene possibly because of enhanced upwelling driven by intensified Kuroshio inflow to the southern Okinawa Trough. The lower COC content of the sediments at that time suggests that much of the primary production was not making it to the seabed or was not preserved. Increased deepwater oxygen supply may have caused greater consumption of marine organics in the water column [*Kao et al.*, 2005b, 2006c]. The TS/TOC ratios (Figure 3f) are generally consistent with the relative COC/FOC inputs (Figure 3d). High COC/FOC ratios argue for greater metabolizing ability of the deposited OC and thus greater availability for sulfate reduction when sea level was low.

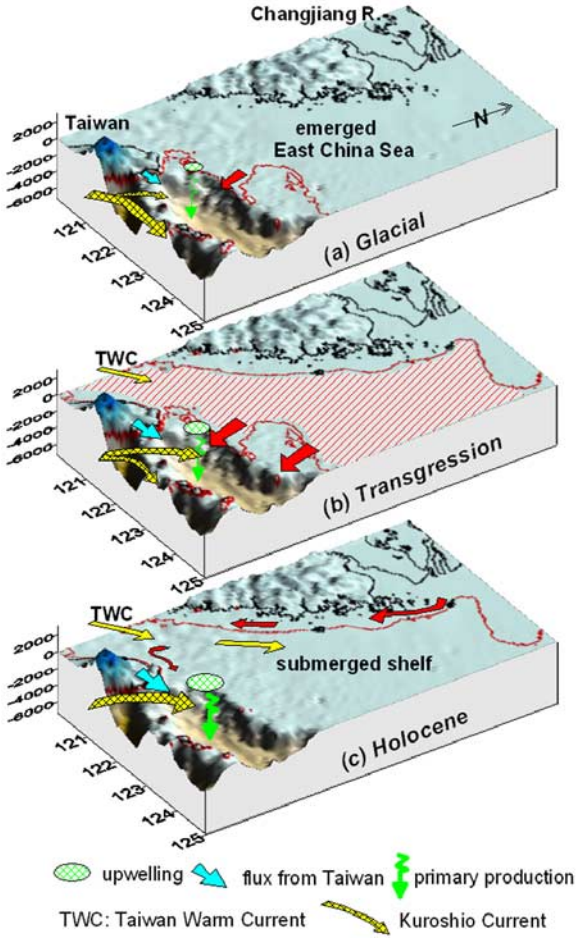

Figure 5. Schematic diagrams for processes and conditions in different time periods. (a) The Kuroshio Current (KC) was weak, and the material flux from Taiwan was low during the glacial period. The sea level was at −120 m. Primary production was low, but the burial of marine organics was relatively high because of weakened oxygen supply to the deep water. (b) Sea level rose from −120 to −40 m during the transgressive period. The submerged coastal region served as a significant source of weathered sediment and terrestrial organic matter (shaded area). KC-induced upwelling and the Taiwan Warm Current (TWC) were weaker. Material flux from Taiwan was moderate. (c) During the Holocene the sea level was above −40 m. Terrestrial material from the Changjiang River was confined near the China coast because of the formation of the TWC. Material flux from Taiwan was enhanced because of the wet climate on land. Upwelling and the KC were intensified. Despite higher surface production, the burial efficiency of fresh organics was low because of enhanced deepwater circulation.

[28] Considering the spatial and temporal dimensions at play in this study, it is prudent to consider whether there are potential TOC pools in the system with ^{14}C contents intermediate to those of the COC and FOC. Some evidence of such a component is seen in the scatterplot of $\delta^{13}C_{TOC}$ versus CIA (Figure 4). The field of data denoted by green represents moderately weathered lithogenic particles and predominantly marine TOC emplaced during low sea level (Figure 5a). The gray field corresponds to the Holocene during which FOC and less chemically weathered sediments that were most likely sourced from Taiwanese rivers provided a greater contribution to the deposition site (see Figure 4). The yellow field represents a mixture of marine organic matter with an additional input from a source with more negative $\delta^{13}C_{TOC}$ and higher CIA values. This additional source is most consistent with a weathered soil component that contains terrestrial plant-derived OC. Given its relative importance during the trangressive period, this material may be alluvia or shallow water sediments that were remobilized (Figure 5b). Though we have no independent evidence, it is possible that the organics in this fraction may be significantly aged. If aged soil OC was an important component during the transgressive period, the FOC and COC have been overestimated for that time.

[29] The supply and transport patterns for sediment and TOC to the southern Okinawa Trough for the glacial, transgression, and Holocene periods are illustrated in Figure 5. When sea level was low (Figure 5a), the seabed of the East China Sea emerged. The material flux from Taiwan was low because of the dry climate and the resulting low erosion and sediment transport rates. The upwelling intensity and primary production may have been low because of a weakened Kuroshio Current, though the burial efficiency of marine organics may have been higher because of the attenuated oxygen supply to the deep water. During transgression (Figure 5b), sea level rise and the newly formed Taiwan Warm Current swept weathered inorganics and terrestrial plant-derived organics from alluvial soils into the Okinawa Trough. Primary production was enhanced gradually because of the intensification of the KC. Sediment flux and the supply of fossil organics from Taiwan increased during the transition into a wetter and humid climate. In the Holocene (Figure 5c), the Taiwan Warm Current blocked delivery of material from the Changjiang and mainland rivers. The KC further intensified. Primary production may have been the highest among the three periods, yet this was not reflected by the burial of marine organics, possibly as a result of enhanced oxygen supply to the deep water. During this period, the wet climate on Taiwan provided the largest supply of fossil organics and unweathered sediment from bedrock.

5. Conclusion

[30] Significant ^{14}C age discrepancies (2500–7900 years) were found between sedimentary total organic carbon and codeposited surface-dwelling foraminifera in the southern Okinawa Trough. The age difference increases with rising sea level and is attributed to an increase in the input of fossil organic C from Taiwan to the Okinawa Trough during the wetter climate of the Holocene. The enhanced contribution of fossil organics from sedimentary rock erosion was accompanied by sediment exhibiting a low index of chemical weathering, a direct consequence of physical weathering. During sea level transgression, weathered materials, such as those from low-lying coastal deposits, contribute more to sedimentary organics in the trough. Variations in the reburial flux of fossil OC appear to be an important part of the climate-impacted carbon biogeochemical cycle and should be considered when reconstructing paleodepositional environments influenced by riverine inputs.

[31] **Acknowledgments.** This study was supported by grants from the National Science Council of Taiwan (NSC 95-2611-M-001-004 to S.J.K. and NSC95-2611-M-002-019 to K.Y.W.). We acknowledge the crew and scientists on board R/V *Marion Dufresne* for retrieving IMAGES core MD 012403. Two anonymous reviewers are gratefully acknowledged.

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