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Constraining the transport time of lithogenic sediments to the Okinawa Trough (East China Sea)

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ABSTRACT

The transport time of siliciclastic sediments from their continental sites of formation to their final location of deposition on the seafloor is an important parameter bearing on land–sea interactions, climate variability and understanding of marine sediment record. $^{234}\text{U}/^{238}\text{U}$ activity ratios of the lithogenic fraction from late Quaternary sediment deposited in the Okinawa Trough, East China Sea, were reported in this study. On basis of $^{234}\text{U}/^{238}\text{U}$ activity ratios, the comminution ages and transport times were calculated using recoil loss factors (f_{α}) derived from different equations based on grain size distribution. The transport times were longer (approximately 200 ± 100 kyr) for the Okinawa Trough sediments deposited between 27 and 14 ka, decreased gradually between 14 and 7 ka, and stayed relatively short (<100 kyr) thereafter. Mineralogical, geochemical and isotopic evidences indicate that changes in sediment transport time correspond well with the shift of sediment provenance predominantly from Asia's interior prior to 14 ka to Taiwan Island after 7 ka. This study offers the first and robust constraint on time scale of sediment transport process in East Asia marginal sea, which is constrained by unique sediment source-to-sink transport pattern. The result illustrates the potential of this approach to decipher climate-related changes in the mode of supply of lithogenic sediment to marginal seas. It also highlights current difficulties in obtaining quantitative estimates of comminution age, mostly because of uncertainties in estimating the recoil loss factor.

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1. Introduction

The East China Sea (ECS) links the Eurasian continent and the Pacific Ocean, and it is characterized by broad continental shelf and huge terrigenous sediment input from adjacent rivers. The river-dominated marginal sea witnessed the complex sediment source-to-sink transport and sedimentary environmental changes during the late Quaternary (Li et al., 2014). Particularly, two distinct river systems predominantly determine the sediment source-to-sink process in this region, the Changjiang (Yangtze River), one of the largest river in the world, and the small mountainous rivers, especially those in Taiwan Island (Yang et al., 2015). The sediment transferring from both river systems is thus of great significance to the sedimentary records and chemical evolution in the ECS. In view of this, the sediment provenances, depositional processes and paleoenvironmental changes in the ECS have been extensively investigated over the last decade (Dou et al., 2010a, 2015; Li et al., 2015b). However, the absolute time scale of sediment transport in the ECS and East Asia continental margin, which is critical to the

sediment source-to-sink processes and marine records, remains unknown (Li et al., 2015a).

The timescale of lithogenic sediment cycling is crucial for assessing the long-term carbon burial by erosion, determining the factors controlling the flux of lithogenic material to the ocean, and understanding the stratigraphic evolution of continental margins and marine records. The lithogenic material accumulating in the Okinawa Trough (OT), was mostly derived from the Changjiang and Taiwan Island (Dou et al., 2010a,b, 2016), and has formed continuous and thick sediment strata during the late Quaternary. As one of the major sinks for terrigenous input in the ECS (Qin et al., 1987), the late Quaternary deposition in the OT provides an important archive for investigating the evolution of Changjiang and Taiwan Island river systems in response to climate change and sea level rise (Li et al., 2015a).

U-series nuclides are widely used to constrain the rates of earth surface processes (Bourdon et al., 2003; Chabaux et al., 2008, 2011; Dosseto et al., 2008; Ma et al., 2010; Vigier and Bourdon, 2011; Dosseto, 2015; Dosseto and Schaller, 2016), and a new approach has recently been proposed to estimate the “transport time” of lithogenic particles from their comminution ages (i.e. the time that has elapsed since their formation by weathering) based on their $^{234}\text{U}/^{238}\text{U}$ activity ratios (DePaolo et al.,

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2006, 2012; Maher et al., 2006; Dosseto et al., 2010; Lee et al., 2010; Handley et al., 2013a). This comminution dating approach is primarily based on $^{234}\text{U}/^{238}\text{U}$ disequilibrium resulted from recoil loss in fine-grained sediments. The method has been applied in North Atlantic deep sea sediment (DePaolo et al., 2006), paleo-channel sediment (Dosseto et al., 2010; Lee et al., 2010; Handley et al., 2013a,b) and Antarctica ice core (Aciego et al., 2011), which yielded reasonable time-scale of sediment transfer across various geology settings. However, its theoretical basis, parameter calculation (e.g. recoil loss factor) and external constraint are still under debate. Therefore, further testing and consideration of the methodology are necessary to improve the accuracy of age estimation and to develop its application in future studies.

The goal of this study is to apply the comminution age method to estimate the transport time of lithogenic sediments deposited in OT, and further to compare the results with changes in sediment provenance deduced from a series of mineralogical (Dou et al., 2010b), geochemical (Dou et al., 2010a) and isotopic analysis (Dou et al., 2012, 2016). Besides, this study also makes the first attempt of applying the comminution age method in the river-dominated marginal sea, which will be an important exploration and contribution to the U-series disequilibrium study.

2. Comminution age theory

The determination of comminution age is based on the continuous loss of ^{234}U from the thin outer layer (~30 nm in thickness) of silicate mineral particles, which results from alpha recoil (Kigoshi, 1971). During this process, ^{234}Th is ejected due to ^{238}U alpha-decay and then ^{234}Th decays to ^{234}U with a half-life of only 24 days. Because ^{234}Th ejection from a mineral grain is only possible from this very thin outer layer, the loss of ^{234}U is a function of the surface area to volume of the particle (Vigier and Bourdon, 2011). The continuous loss of ^{234}U results in a measurable decrease in ($^{234}\text{U}/^{238}\text{U}$) (parentheses denote activity ratios throughout this paper) of the entire grain only when the surface area to volume increases to certain extent (approximately at ~50 μm diameter) (DePaolo et al., 2006). The “comminution age” is thus defined as the time elapsed since the sediment grain is smaller than ~50 μm . Once such small particles have been formed by weathering and erosion, their ($^{234}\text{U}/^{238}\text{U}$) ratios start to decrease. If the small particle undergoes no additional abrasion or loss of depleted surface, their ($^{234}\text{U}/^{238}\text{U}$) eventually reaches a steady state value, which is determined by the size and shape of the particles and the roughness of their surface. However, it is argued that preferential loss of ^{234}U relative to ^{238}U may occur via leaching during weathering of the source rock and/or sediment transport process (Eyal and Olander, 1990; Bourdon et al., 2009). However, DePaolo et al. (2006) and Maher et al. (2006) have found that the observed loss of ^{234}U can be caused by α -recoil effects only and does not require preferential leaching of ^{234}U , especially when the leaching occurs in a depth that is not much greater in magnitude than the recoil range. With all these assumptions, the “comminution age” (t_{com}) can be calculated from Eq. (1) (DePaolo et al., 2006):

$$t_{\text{com}} = -\frac{1}{\lambda_{234}} \ln \left[\frac{A_{\text{meas}} - (1 - f_{\alpha})}{A_0 - (1 - f_{\alpha})} \right] \quad (1)$$

where A_0 is ($^{234}\text{U}/^{238}\text{U}$) of the parent rock, A_{meas} is ($^{234}\text{U}/^{238}\text{U}$) of the sample studied, λ_{234} is the decay constant of ^{234}U , and f_{α} is the recoil loss factor, i.e. the fraction of ^{238}U decay in the sample that results in the ejection of a ^{234}Th atom (Kigoshi, 1971; Maher et al., 2006). The “transport time” of fine-grained particles between sites of their formation from parent rocks to sites of final deposition (e.g. seafloor) can then be calculated by subtracting the depositional age, obtained from core chronology, from the comminution age. Obviously, the calculated “transport time” integrates the storage time of particles in weathering profiles, their transport time in river channels, and residence time in alluvial plains and on the continental shelf (Dosseto et al., 2010).

3. Study area

The Okinawa Trough (Fig. 1a) is a typical back-arc basin of the Ryukyu trench-arc system, bounded by the Ryukyu Ridge and Trench to the south and east, and by the ECS shelf to the north and west. The entire OT is arcuate, convex toward the west Pacific, from Japan to Taiwan. It has a large section of more than 1000 m in depth and the deepest part, near Taiwan Island, is about 2270 m deep. The OT shoals gradually northeastward toward Japan and is underlain by about 1–2 km of sediment (Lee et al., 1980). The OT is a depositional basin with a relatively high rate of sedimentation of primarily terrigenous sediment from the East Asia continent, ECS continental shelf and island arc via the numerous adjacent rivers (Qin et al., 1987). As a passage linking East Asian continent to the west Pacific Ocean, the OT may serve as a sensitive reflection of environmental transition between the ocean and continental settings. The most striking oceanographic feature in the OT is the Kuroshio Current which is the largest western boundary current in the North Pacific Ocean.

Among the numerous rivers entering the ECS, the Changjiang (Yangtze River) is the largest one in East Asian continent. It originates from the Tibet Plateau and its catchment, which is up to $1.8 \times 10^6 \text{ km}^2$ in area, is primarily situated on the Yangtze Craton. Geologically, the Changjiang catchment comprises complex rock types including Archean metamorphic rocks, Jurassic sandstone, Paleozoic carbonate and sedimentary rocks, Mesozoic–Cenozoic igneous and clastic rocks, and Quaternary detrital sediments (Yang et al., 2004). Base on the long-term hydraulic observation, the Changjiang annually delivers about 470 Mt suspended sediments to the ECS (Milliman and Farnsworth, 2011). Major part of the Changjiang-derived sediment is trapped in its estuary and deposited on adjacent ECS shelf (Liu et al., 2007), while the remainder may be transported to the Okinawa Trough, resulting in a thick sedimentary deposit (Qin et al., 1987).

Apart from the large rivers, small rivers in East Asia also play an important role in sedimentation in the ECS, in particular the small mountainous rivers from Taiwan Island (Kao and Milliman, 2008). The island of Taiwan is characterized by its strong tectonic uplift at a rate of 5–10 mm/yr (Shin and Teng, 2001), and high physical erosion rate up to 3–6 mm/yr (Dadson et al., 2003). Together with the frequent typhoon and earthquake events, the rivers in Taiwan discharge about 180 Mt/yr sediment to the surrounding marginal seas, showing one of the highest sediment yields in the world (Kao and Milliman, 2008). The Taiwan river basins are mainly composed of sedimentary rocks and epimetamorphic rocks including sandstone, shale, slate and phyllite, with rare occurrence of acidic rocks. The Zhuoshui (also named Chuoshui) River as the largest one in Taiwan, originates from Central Mountain Range, with an elevation of about 3000 m and the total length of 186 km.

4. Samples and methods

4.1. Sources of river and marine sediment samples

In this study, a total of 24 sediment samples were selected from the piston core DGKS9604 (28°16.64' N, 127°01.43' E, 766 m water depth) taken from the OT in 1996 during the joint Chinese–French DONGHAI Cruise (Fig. 1a). The age model is derived from oxygen isotopic composition and radiocarbon dates measured on planktonic foraminifera *Globigerinoides sacculifer* determined by accelerator mass spectrometry (Yu et al., 2009).

For constraining the sediment transport times of the modern Changjiang River, two suspended sediment samples were collected near Chongqing (CQ) and Nantong (NT), which represent the upstream and estuary (Fig. 1c), respectively. Another two riverine suspended samples from Taiwan Island were collected from the upstream (ZS-1) and estuary (ZS-2) of the Zhuoshui River (Fig. 1d), respectively. All the riverine suspended samples are collected by a 0.45 μm filter. Detailed sample information is shown in Table 1.

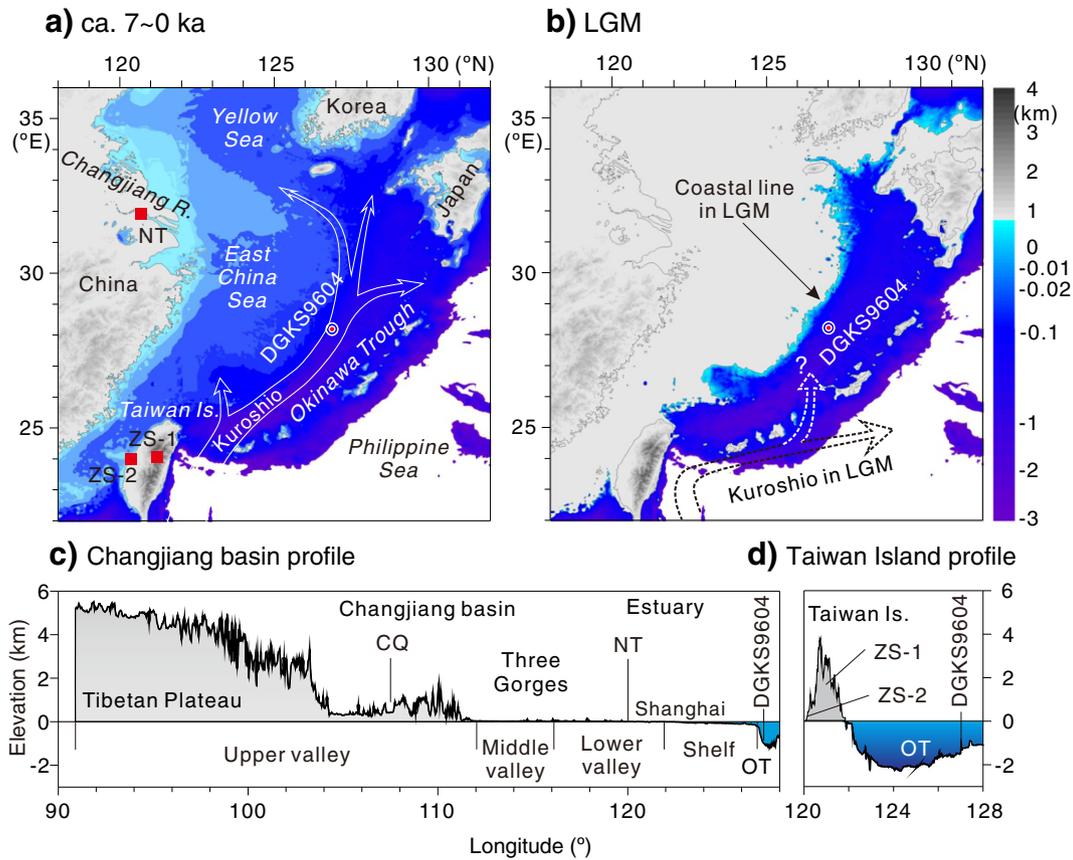


Fig. 1. Map of the East China Sea showing the location of core DGKS9604, the coastline and oceanic circulation at present (a) and during the last glacial maximum (LGM) (b). The LGM coastline is based on the 120 m isobaths (Ujiié and Ujiié, 1999). The major current system in the East China Sea and pathway of the Kuroshio Current is modified from Ichikawa and Beardsley (2002). Transects of the Changjiang catchment to Okinawa Trough (OT) and Taiwan Island to OT are shown in panel (c) and (d), correspondingly. The Changjiang (CQ and NT) and Zhuoshui River (ZS-1 and ZS-2) samples are also indicated in panel (a), (c) and (d).

4.2. $^{234}\text{U}/^{238}\text{U}$ measurements

The river and marine sediment samples were leached with 1.5 N HCl for 30 min at room temperature to remove carbonate and authigenic phases, following the method by DePaolo et al. (2006). After centrifugation, about 0.1 g of the residue was used for grain size analysis, and 0.2 g was dried and ground for the measurement of uranium isotopes. The powdered samples were digested in mixed acids ($\text{HClO}_4\text{-HF-HNO}_3$), before U separation on UTEVA® resin (98% column recovery for U; the procedure blank and column blank are 0.5% and 0.003% of the samples, correspondingly). The measurement of ($^{234}\text{U}/^{238}\text{U}$) ratios was carried out on a Multi-Collector ICP-MS (Nu 021; Nu Instruments Ltd., UK) at the Pacific Centre for Isotopic and Geochemical Research, University of British Columbia, following the procedure described in Andersen et al. (2004). We used standard bracketing with CRM-145B in which ($^{234}\text{U}/^{238}\text{U}$) was verified against Plešovice zircons (Sláma et al., 2008). The $\delta^{234}\text{U}$ value obtained for CRM-145B relative to these zircons ($-37.47 \pm 1.24\%$ (2 standard deviations, $n = 5$)) was within the error of $\delta^{234}\text{U}$ reported for CRM-145B ($-36.50 \pm 0.14\%$ (Andersen et al., 2004)). The latter value was thus used as reference for the calculation of activity ratios in the samples. The external reproducibility of the measurements was estimated by analyzing BCR-2 ($\delta^{234}\text{U} = 4.13 \pm 2.29\%$ (2 standard deviations, 54 measurements)).

4.3. Determination of grain size distributions

The grain size distribution of the sediment samples was measured on a Malvern Mastersizer 3000 particle size analyzer at the University of British Columbia, while grain size distribution of the river samples

was measured on a Beckman Coulter LS230 particle size analyzer at the State Key Laboratory of Marine Geology, Tongji University.

5. Calculation of comminution ages and transport time from ($^{234}\text{U}/^{238}\text{U}$)

The accuracy of comminution ages calculated with Eq. (1) depends on the validity of the assumptions made regarding A_0 and all f_{α} . One assumption is that in the absence of large crustal fluid flow, the U-series decay chain of continental rocks is generally in secular equilibrium (Vigier and Bourdon, 2011). Supporting this view, a systematic investigation of ($^{234}\text{U}/^{238}\text{U}$) in rocks taken from a glacial outwash yielded values very close to secular equilibrium (1.00 ± 0.01), independent of lithology (DePaolo et al., 2012). However, Handley et al. (2013b) found significant ^{234}U depletion in sedimentary rock samples, raising the possibility that the assumption of secular equilibrium in parent rocks may not always be correct. While A_0 could potentially be estimated when the source rock of sediment can be clearly identified, in most instances this is very difficult and thus A_0 is often assumed to be 1 (DePaolo et al., 2006, 2012; Maher et al., 2006; Dosseto et al., 2010; Lee et al., 2010; Handley et al., 2013a). Determining A_0 for the lithogenic sediments from the Changjiang River, ECS shelf and OT as well is particularly difficult considering the very complicated and diverse provenance rock types in the large Changjiang drainage basin, and complex sediment source-to-sink transport in the continental margin. Taking this into account, $A_0 = 1$ is used in this study to provide a general estimation of sediment transport time for sediment from the Changjiang and Taiwan Rivers, as well as OT.

The recoil loss factor (f_{α}) is also difficult to accurately estimate because it depends on the size, shape and surface roughness of mineral

Table 1
Grain size distribution and ($^{234}\text{U}/^{238}\text{U}$) in core DGKS9604 sediments from the Okinawa Trough and in the suspended particles from the Changjiang and Zhuoshui Rivers.

Depositional time (ka)	Mean grain size ^a (μm)	Sand ^b (%)	Silt ^b (%)	Clay ^b (%)	($^{234}\text{U}/^{238}\text{U}$)
0.1	11.9	3	58	39	1.028 \pm 0.001
1.1	11.1	2	59	39	0.988 \pm 0.001
1.6	11.1	2	60	38	1.004 \pm 0.001
4.1	11.0	2	59	39	1.017 \pm 0.002
4.6	10.9	2	60	38	1.040 \pm 0.002
7.1	12.2	2	59	39	0.988 \pm 0.002
7.6	14.4	3	63	34	0.973 \pm 0.002
8.2	12.6	3	59	38	0.979 \pm 0.001
8.7	12.6	3	58	39	0.987 \pm 0.001
10.2	14.1	4	61	35	1.011 \pm 0.001
11.3	13.6	4	58	38	0.962 \pm 0.002
11.8	13.0	3	55	42	0.942 \pm 0.001
13.0	13.1	4	58	38	0.960 \pm 0.002
13.7	12.9	3	64	33	0.942 \pm 0.001
13.9	13.5	3	60	37	0.945 \pm 0.002
15.8	11.0	1	65	34	0.922 \pm 0.003
17.1	10.1	1	62	37	0.940 \pm 0.002
18.5	9.7	1	59	40	0.919 \pm 0.002
20.2	10.9	2	59	39	0.917 \pm 0.002
21.7	11.7	2	61	37	0.923 \pm 0.002
22.6	11.5	2	61	37	0.918 \pm 0.002
23.3	10.1	1	59	40	0.928 \pm 0.001
26.2	12.3	3	57	40	0.922 \pm 0.001
27.3	12.8	3	61	36	0.924 \pm 0.002
CQ ^c	9.9	0	59	41	0.979 \pm 0.002
NT ^c	8.8	0	59	41	0.941 \pm 0.001
ZS-1 ^c	11.1	1	74	25	1.002 \pm 0.002
ZS-2 ^c	15.4	1	80	19	0.960 \pm 0.001

^a Detailed grain size distribution is available in appendix Table A. 1.

^b Relative proportion of sand ($d > 62.5 \mu\text{m}$), silt ($3.9 \mu\text{m} < d < 62.5 \mu\text{m}$), and clay ($< 3.9 \mu\text{m}$).

^c The Changjiang River samples CQ (29.56°N, 106.59°E) and NT (31.96°N, 120.83°E) were collected from the upper and lower reaches, respectively. Samples ZS-1 (23.97°N, 121.11°E) and ZS-2 (23.82°N, 120.21°E) were respectively taken from the upper and lower Zhuoshui River.

grains. Different approaches have been proposed to estimate f_{α} based on (a) the distribution of grain size in the sample for U isotope analysis, and the assumptions for changes in surface roughness and grain aspect ratio as a function of grain size (DePaolo et al., 2006; Maher et al., 2006; Dosseto et al., 2010; Lee et al., 2010; Handley et al., 2013a); (b) surface area measurements (e.g., Brunauer–Emmett–Teller (BET)) with fractal correction to account for the significant size difference between the adsorbed gas molecules used for surface area measurement and the recoil length scale of alpha particles (Olley et al., 1997; Aciego et al., 2011; Handley et al., 2013a); and (c) ($^{234}\text{U}/^{238}\text{U}$) measured in samples approaching a steady state (i.e. older than 500 ka) (Lee et al., 2010).

In the present study, we estimate f_{α} using two independent methods based on grain size distribution. The method proposed by DePaolo et al. (2006) is based on Eq. (2):

$$f_{\alpha} = \int_{L/2}^{r_{\max}} X(r)\beta(r)\lambda_s(r) \frac{3}{4} \left(\frac{L}{r} - \frac{L^3}{12r^3} \right) dr. \quad (2)$$

Where $X(r)$ is the volume fraction of different particle size intervals (dr) present in the sample analyzed for U isotopic ratio, $\beta(r)$ is their aspect ratio, $\lambda_s(r)$ is their surface roughness factor, and L is the α -recoil length generally taken to be 30 nm. Following previous studies (DePaolo et al., 2006; Dosseto et al., 2010; Handley et al., 2013a), we also assume that: (a) the aspect ratio ($\beta(r)$) of particles with $r < 25 \mu\text{m}$ increases linearly with decreasing grain size, from 1 for the largest grains to 10 for the smallest; and (b) the surface roughness $\lambda_s(r)$ increases linearly with grain size from 1 for the smaller grains to 2 (DePaolo et al., 2006) or 17 (Handley et al., 2013a) for the larger

grain size. Thereby, a range of f_{α} and comminution age for each of our samples can be obtained on the basis of the above assumptions.

Another approach to evaluate f_{α} assumes a constant surface roughness (λ_s) and a constant dimensionless grain shape factor K (Cartwright, 1962) over the entire grain size spectrum (Lee et al., 2010):

$$f_{\alpha} = \frac{LK}{4d} \lambda_s. \quad (3)$$

Where d is the grain diameter and L is the recoil distance. Freshly crushed silicate minerals generally have a relatively constant $\lambda_s = 7$ over a wide range of grain size (White and Peterson, 1990) and this value was used for Eq. (3). $K = 6$ is the grain shape factor for spheres, but for silicate minerals with mean particle sizes ranging from 0.2 to 10 μm , $K = 14$ –18 (Cartwright, 1962). We thus calculated f_{α} with Eq. (3) using a constant $\lambda_s = 7$ and $K = 6$ or 18 to obtain a range of comminution ages and transport times for our samples.

Notwithstanding the uncertainties raised by the basic premises of the approach above, we calculate the transport times of the lithogenic sediments deposited in the OT from their ($^{234}\text{U}/^{238}\text{U}$), assuming that $A_0 = 1$, and estimate f_{α} using the two approaches described above. The goal of this study is to investigate whether there are coherent changes in the calculated transport times of lithogenic particles, and whether the transport times are consistent with changes in sediment provenance reconstructed in earlier studies (Dou et al., 2010a,b).

6. Results

6.1. Uranium isotopic ratios

The ($^{234}\text{U}/^{238}\text{U}$) ratios of the OT sediment range from 0.917 to 1.040, with a clear temporal trend (Fig. 2a). The ratios are relatively low and uniform between 27 and 14 ka with a mean value of 0.924, increase gradually between 14 and 7 ka, and stay high and closer to secular equilibrium after 7 ka. Some samples from the upper section of the core have ($^{234}\text{U}/^{238}\text{U}$) values > 1 , which cannot be directly interpreted in terms of comminution ages.

The ($^{234}\text{U}/^{238}\text{U}$) ratios of the Changjiang River (China) sediments range from 0.979 \pm 0.002 at CQ to 0.941 \pm 0.001 at NT, while the ($^{234}\text{U}/^{238}\text{U}$) ratios for the Zhuoshui River (Taiwan) sediments range from 1.002 \pm 0.002 at ZS-1 to 0.960 \pm 0.001 at ZS-2 (Table 1). Both rivers show decreasing ($^{234}\text{U}/^{238}\text{U}$) ratios toward the lower reaches.

6.2. Sediment grain size and f_{α}

The sediment samples from the OT are dominated by silt ($3.9 \mu\text{m} < d < 62.5 \mu\text{m}$) and clay ($< 3.9 \mu\text{m}$) fractions, which account for 60 \pm 7% and 37 \pm 5% correspondingly. The sand content ($d > 62.5 \mu\text{m}$) is low throughout the core (2 \pm 2%), and the downcore variations of mean grain size is also small (Table 1; Fig. 3a, b). Similarly, f_{α} yields small changes with core depth although the absolute values of f_{α} estimated by different assumptions vary significantly (Table 2; Fig. 3c).

The river particles from the Changjiang have similar proportions of silt (59%) and clay (41%) as the OT sediments, without sand-size material, while the samples collected from the Zhuoshui River consist of more silt (74–80%) and less clay (19–25%) (Table 1). f_{α} estimates for the Changjiang River particles are higher, while those for the Zhuoshui River particles are slightly lower compared to the OT sediments.

6.3. Calculated transport times

Transport times of the river and marine sediments derived from f_{α} with two different methods (Eqs. (2) and (3)) are listed in Table 2. The estimated transport times for the OT sediment are longer during the last glacial period and gradually decrease through the last

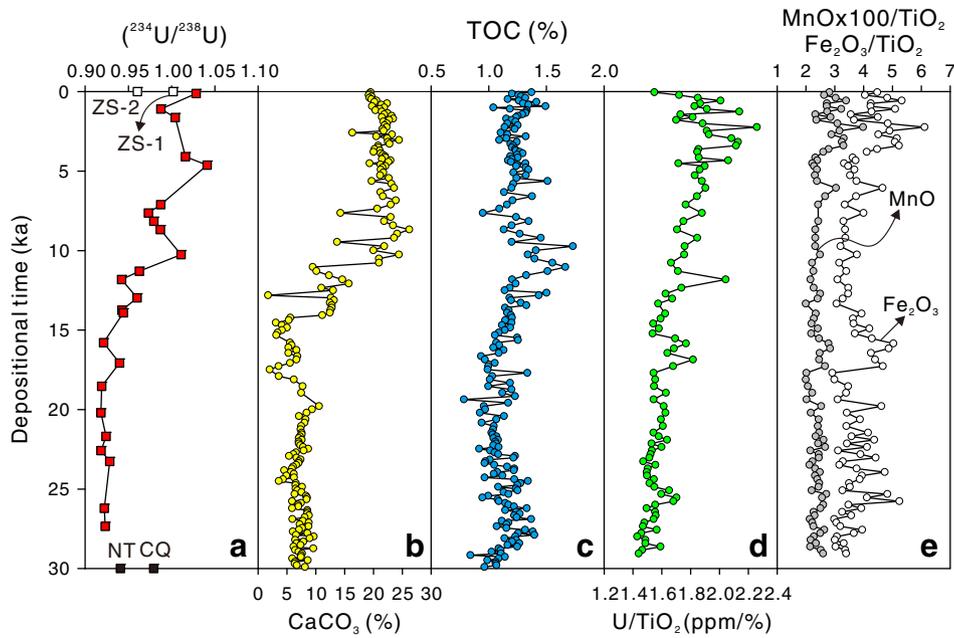


Fig. 2. Downcore profiles of ($^{234}\text{U}/^{238}\text{U}$) (a), CaCO_3 (b), total organic carbon (TOC) (c), U/TiO_2 (d), MnO/TiO_2 and $\text{Fe}_2\text{O}_3/\text{TiO}_2$ (e) in core DGKS9604. CaCO_3 data is from Dou et al. (2010a). TOC data is from Dou (2010), U, TiO_2 , Fe_2O_3 and MnO data (1 N HCl residue) are from Dou et al. (2012). Note that the ($^{234}\text{U}/^{238}\text{U}$) of the Changjiang and Zhuoshui river sediments are presented at the 30 ka and 0 ka level (a), correspondingly.

deglaciation, mirroring the downcore variability of ($^{234}\text{U}/^{238}\text{U}$). The range of transport time obtained with Eq. (3) for K of 6–18 is overall larger than that obtained with Eq. (2) for λ_{max} of 2–17 (Fig. 4a).

The estimated transport times for the suspended particles collected in the Changjiang and Zhuoshui Rivers increase downstream. Particles from the upper Changjiang (CQ) yield a transport time of 21–67 ka, compared to 66–253 ka for particles from the lower Changjiang (NT). Likewise, the transport time for suspended particles from the upper Zhuoshui (ZS-1) is near zero, while the lower Zhuoshui (ZS-2) sample yields transport times of 82–344 kyr, showing a larger range than the counterpart of the Changjiang (Table 2).

7. Discussions

7.1. Uncertainties of comminution age calculation in the OT sediment

7.1.1. ($^{234}\text{U}/^{238}\text{U}$) in marine sediments

It is notable that some OT sediments show ($^{234}\text{U}/^{238}\text{U}$) > 1, which thus yield several negative comminution ages (Table 2). These abnormal ($^{234}\text{U}/^{238}\text{U}$) values are probably a result of the non-conservative behavior of uranium in seawater. Uranium with ($^{234}\text{U}/^{238}\text{U}$) close to that of seawater (1.14) is expected to be present in the biogenic (carbonates) and authigenic (organic matter, Fe–Mn oxyhydroxides and clays)

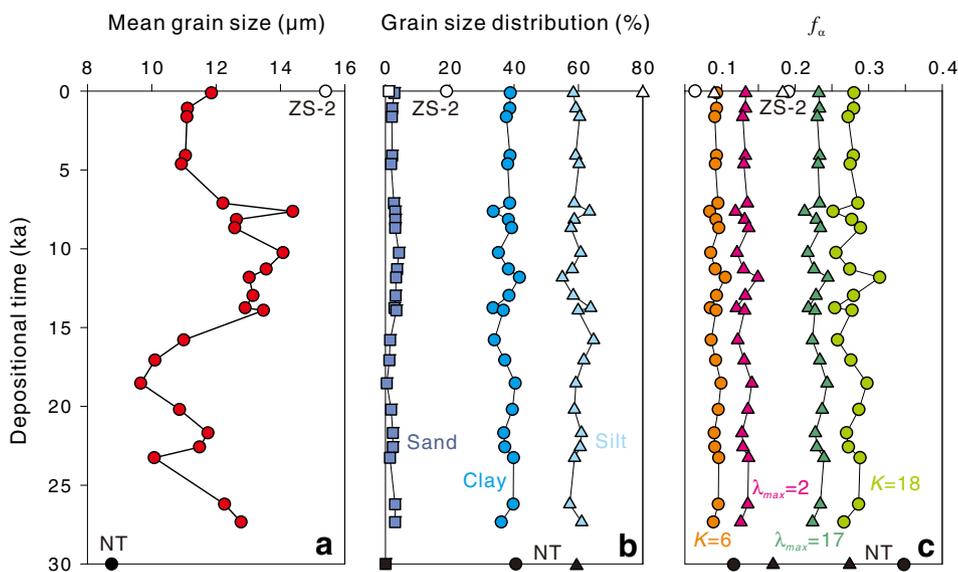


Fig. 3. Comparison of mean grain size (a), grain size distribution (b) and f_{α} (c) for the core DGKS9604 sediments and two river sediments from the Changjiang (NT; in black at 30 ka level) and Zhuoshui River (ZS-2; in white at the 0 ka level). f_{α} is calculated using Eq. (2) assuming that the surface roughness factor (λ_s) varies from 1 to 2 (DePaolo et al., 2006) and from 1 to 17 (Handley et al., 2013a) with increasing grain size, and using Eq. (3) with a constant surface roughness factor (λ_s) of 7 and grain shape factor (K) varying from 6 to 18 (Cartwright, 1962; Lee et al., 2010).

Table 2
Recoil loss factors (f_{α}) and transport times (T_{trans}) calculated with different equations. See the text for the detailed equations and assumptions.

Depositional time	f_{α}^a	f_{α}^a	T_{trans}	T_{trans}	f_{α}^b	f_{α}^b	T_{trans}	T_{trans}
(ka)	$\lambda_{max} = 17$	$\lambda_{max} = 2$	$\lambda_{max} = 17$	$\lambda_{max} = 2$	$K = 18$	$K = 6$	$K = 18$	$K = 6$
			(kyr)	(kyr)			(kyr)	(kyr)
<i>Core DGKS9604 sediments</i>								
0.1	0.232	0.133	-40	-67	0.280	0.093	-34	-92
1.1	0.233	0.133	18	32	0.280	0.093	14	47
1.6	0.230	0.129	-8	-13	0.272	0.091	-7	-18
4.1	0.233	0.132	-28	-46	0.279	0.093	-24	-62
4.6	0.231	0.130	-61	-100	0.275	0.092	-53	-133
7.1	0.233	0.135	12	26	0.285	0.095	8	41
7.6	0.213	0.119	39	82	0.252	0.084	32	127
8.2	0.229	0.131	26	54	0.277	0.092	20	84
8.7	0.234	0.137	11	26	0.289	0.096	7	42
10.2	0.217	0.121	-28	-41	0.256	0.085	-25	-53
11.3	0.226	0.130	55	113	0.274	0.091	42	181
11.8	0.244	0.149	83	161	0.315	0.105	60	269
13.0	0.228	0.132	55	115	0.279	0.093	42	186
13.7	0.218	0.120	95	216	0.254	0.085	77	388
13.9	0.227	0.131	84	177	0.278	0.093	64	303
15.8	0.224	0.122	136	347	0.258	0.086	112	841
17.1	0.234	0.131	88	200	0.276	0.092	70	357
18.5	0.244	0.141	125	283	0.298	0.099	94	581
20.2	0.237	0.136	132	311	0.287	0.096	100	687
21.7	0.228	0.128	124	303	0.270	0.090	97	656
22.6	0.230	0.129	135	337	0.273	0.091	105	821
23.3	0.239	0.137	103	241	0.288	0.096	78	464
26.2	0.234	0.136	116	274	0.286	0.095	86	568
27.3	0.224	0.126	119	297	0.266	0.089	91	650
<i>River samples</i>								
CQ	0.276	0.177	28	44	0.362	0.121	21	67
NT	0.274	0.170	86	151	0.348	0.116	66	253
ZS-1	0.205	0.105	-3	-6	0.222	0.074	-3	-8
ZS-2	0.183	0.090	86	206	0.191	0.064	82	344

^a f_{α} calculated from Eq. (2).

^b f_{α} calculated from Eq. (3).

phases of marine sediment. The carbonate content in the core DGKS9604 sediment is overall low (5–10%) prior to 14 ka, and gradually increases to 20–25% after 7 ka (Fig. 2b). The uranium content of biogenic calcite is low, about 20–30 ppb (Russell et al., 1994). Since biogenic

calcite readily dissolves in 1.5 N HCl, this small pool of uranium should have been effectively removed by our sample pre-treatment.

The total organic carbon (TOC) content in the core sediments ranges between 1.0% and 1.5%, without a clear trend with depth (Fig. 2c). The U/TiO₂ ratios (ppm%; Fig. 2d) in the core sediments decrease toward the lower part of this core, and remain below the ratio reported for the upper continental crust (0.18 ppm%; (Taylor and McLennan, 1995)). Thus, the level of authigenic U in OT sediment appears to be very low, which is expected considering very high sedimentation rate of lithogenic material at the study site, about 29 cm/kyr (Dou et al., 2010a). Adsorption of U from bottom water or pore water on these Fe–Mn oxyhydroxides could contribute authigenic U to the sediments. However, a significant addition of authigenic U by this mechanism is not supported by the downcore variability of U/TiO₂ (Fig. 2e), which does not increase concomitantly with MnO/TiO₂ and Fe₂O₃/TiO₂ in the core (Fig. 2f). Secondary clay minerals could also have high (²³⁴U/²³⁸U) ratios relative to their amalgamated pellet grain size, and if present in abundance, could offset calculations based on grain size (Handley et al., 2013a). As indicated by ZS-1 (²³⁴U/²³⁸U = 1.002 ± 0.002), the Taiwan riverine sediments are closer to 1 according to the comminution theory, and thus a small amount of dissolved U trapped by clays may shift the (²³⁴U/²³⁸U) to more than 1. On this account, the ²³⁴U excess in the upper core sediments is most likely to be associated with the clay fraction in Taiwan riverine sediment.

Overall, the concentration of authigenic U appears too low to significantly affect the (²³⁴U/²³⁸U) of the OT sediment, largely because the much larger addition of lithogenic uranium overwhelms the much slower accumulation rate of authigenic uranium. However, four of the Holocene sediment samples yield (²³⁴U/²³⁸U) significantly higher than 1 (Fig. 2a), which suggests significant contribution from authigenic phases. Hence, it is doubtful that the simple acid leaching (1.5 N HCl, 30 min at room temperature) provided by DePaolo et al. (2006) may not be as applicative as be used for north Atlantic marine sediment, and adds more ambiguities to the interpretation of these results. An more appropriate and targeted chemical leaching method to remove authigenic U fraction should therefore be treated with great caution (Suresh et al., 2014; Martin et al., 2015).

7.1.2. (²³⁴U/²³⁸U) of parent rocks–A₀

Taiwan Island has a steep topography and is subjected to a rapid tectonic uplift, resulting in high erosion rates of up to 3–6 mm/yr (Dadson et al., 2003). As a result, the small mountainous rivers in Taiwan show

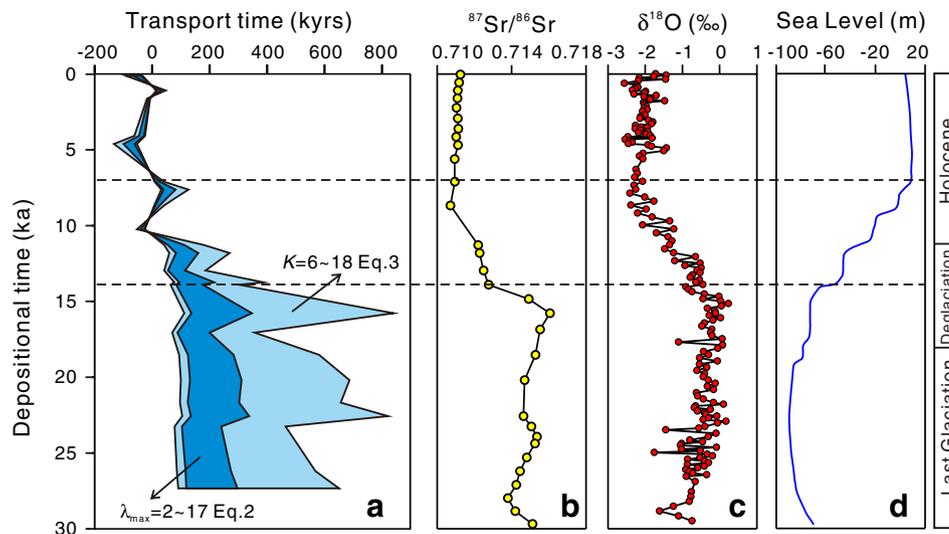


Fig. 4. (a) Transport time of the lithogenic fraction in core DGKS9604 sediments. Range of transport times estimated for DGKS9604 are indicated in dark blue (Eq. (2)) and light blue (Eq. (3)). (b) ⁸⁷Sr/⁸⁶Sr ratios in the lithogenic fraction (Dou et al., 2012); (c) oxygen isotopic ratios in *Globigerinoides sacculifer* (Yu et al., 2009); (d) sea level variability over the last 30 kyr (modified after Lambeck et al. (2002)). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

by far the highest sediment yield in the world (Kao and Milliman, 2008). The ($^{234}\text{U}/^{238}\text{U}$) for sediment from the upper Zhuoshui is 1.002 ± 0.002 , revealing that the sediment is quite fresh and still in the state of secular equilibrium. However, it is hard to evaluate the A_0 for the Changjiang sediment due to the large river basin and complex provenance lithology. The river sediment from upper Changjiang shows ($^{234}\text{U}/^{238}\text{U}$) ratio of 0.979 ± 0.002 , which is close to the state of secular equilibrium. On the other hand, Handley et al. (2013a) indicates that the t_{com} variation derived from f_a uncertainty is much larger than that derived from A_0 uncertainty based on a Monte Carlo statistics. On this account, the assumption of $A_0 = 1$ for the Changjiang sediment may still yield a reasonable calculation of transport time.

7.1.3. Recoil loss factor (f_α) and calculated transport time

The three-fold range in f_α calculated with Eqs. (2) and (3) (Fig. 3c) suggests that without a better means of constraining this important parameter, estimation of comminution ages will remain a large uncertainty. Considering the variations of f_α between the last glaciation and Holocene from core DGKS9604, however, these estimates still provide useful information. Notwithstanding the large uncertainties in absolute comminution ages, they do indicate a clear change from longer transport times during the last glacial period to much shorter transport times through the last deglaciation and Holocene (Fig. 4a; Table 2). This is in large part because of the relatively constant grain size (and presumably f_α) with depth at our study site, and the sharp contrast in the origins and transport pathways of lithogenic sediments deposited in OT over the last glacial/interglacial climatic cycle.

The transport times calculated with Eqs. (2) and (3) overlap but the range is significantly smaller for Eq. (2) (Fig. 4a). The largest and smallest grain shape factors (K) used in Eq. (3) thus appear inconsistent with a reasonable range of parameters for Eq. (2). This may be because the grain size distribution of OT sediment is relatively consistent and thus, the grain shape factors (K) vary only in a small range. Furthermore, $K = 6$ for sphere shape grain yields the largest deviation for the calculated transport time, suggesting that the sphere grain model is not applicable for the OT sediment.

Overall, the calculation results suggest that the range of transport times for glacial lithogenic sediments estimated by Eq. (2) (approximately 200 ± 100 kyr) may provide our best estimates. On the other hand, while clearly shorter (<100 kyr), the transport time of lithogenic particles accumulated in the Holocene section is too short to be quantified by the method as presently developed.

7.2. Transport time variations induced by changes of sediment provenance

7.2.1. Provenance changes of the OT sediment since 27 ka based on previous studies

The changes of sediment provenances and paleoenvironment in the OT have been extensively investigated over the last decade (Yang et al., 2015). Generally, the sediment provenances changed gradually in the OT during the last 27 kyr. The OT sediment deposited in the last glacial period was mainly derived from East Asian continent and/or the ECS continental shelf through the Changjiang River transport, while the Taiwan Island became the main sediment source of the OT with the strengthening of Kuroshio Current during the Holocene. This conclusion has been verified by various lines of evidence from rare earth elements (Dou et al., 2010a), Sr (Fig. 4b) and Nd isotopes (Dou et al., 2012; Li et al., 2015b), and clay mineralogy (Diekmann et al., 2008; Dou et al., 2010b; Wang et al., 2015). During the last glacial maximum, the temperature is much colder (Fig. 4c) and the sea level was about 120–135 m lower than today in East Asia marginal sea and the continental shelf was largely exposed (Figs. 1b and 4d). The paleo-Changjiang discharged a large volume of fine-grained sediments from the East Asia continent into the mid-outer ECS shelf, and finally to the OT (Yang et al., 2015). At the same time, the main stream of the Kuroshio Current might have shifted eastward (Ujiié et al., 1991) (Fig. 1b), limiting the supply of sediment

sourced from Taiwan Island. During the period of deglaciation and early Holocene (ca. 14–7 ka), the temperature increased sharply (Fig. 4c) and the ECS coastline and paleo-Changjiang River mouth retreated gradually with the continuously rising sea level (Figs. 1a and 4d), while the major Kuroshio Current shifted westward over the OT (Ichikawa and Beardsley, 2002). Consequently, the sediment contribution from the Changjiang gradually declined because of increased trapping of the river sediment in the estuary and on the inner shelf (Yang et al., 2015), while the sediment supply from Taiwan Island gradually increased with the strengthening of the Kuroshio Current.

7.2.2. Provenance changes of the OT sediments evidenced by ($^{234}\text{U}/^{238}\text{U}$) and transport time

The change of sediment provenance in the OT is consistent with the gradual increase of ($^{234}\text{U}/^{238}\text{U}$) in the core sediments (Fig. 2a). Regardless of possible effect of preferential release of ^{234}U into solution during alteration/dissolution, when a small mineral grain is produced by erosion, it begins to leak ^{234}Th to its surroundings as a result of α -recoil (Bourdon et al., 2003), and the ($^{234}\text{U}/^{238}\text{U}$) starts to decrease. Thus, value of ($^{234}\text{U}/^{238}\text{U}$) in sediment provides a qualitative measurement of the time since the small grain was produced, which has been coined by DePaolo et al. (2006) as “comminution age”. In our case, the ($^{234}\text{U}/^{238}\text{U}$) ratios are relatively low and uniform (0.924 on average) between 27 and 14 ka, suggesting a longer sediment transport process. In comparison, the ($^{234}\text{U}/^{238}\text{U}$) ratios increase gradually between 14 and 7 ka, and stay high and closer to secular equilibrium after 7 ka, which indicates short sediment transport times. The variations of ($^{234}\text{U}/^{238}\text{U}$) in the core DGKS9604 sediments correspond well with the changes of Sr isotopic composition (Fig. 4b) which is a robust evidence for sediment provenance alteration.

The ($^{234}\text{U}/^{238}\text{U}$) for the modern Changjiang is 0.979 in the upstream (CQ) and 0.941 in estuarine samples (NT), while the ($^{234}\text{U}/^{238}\text{U}$) from the Zhuoshui River are 1.002 and 0.960 for upper and lower reaches, respectively. The higher ($^{234}\text{U}/^{238}\text{U}$) for the upstream Changjiang sediment (CQ) is probably because of the upper Changjiang basin being featured by active tectonics, high relief and steep river channel (Fig. 1c), which results in fast erosion and rapid transport of small particles after their formation. In contrast, the relatively lower ($^{234}\text{U}/^{238}\text{U}$) ratio in the Changjiang estuary (NT) indicates a longer transport time from the upper valley to the estuary. The lower Changjiang reaches is characterized by well-developed flood plains and numerous lakes and basins in the mid-lower valley (Fig. 1c), which may effectively trap the particles derived from the upper reaches, and thereby increase their residence times in the mid-lower Changjiang valley. In comparison, the ($^{234}\text{U}/^{238}\text{U}$) ratio in the Zhuoshui River sediments is overall larger than in the Changjiang samples, suggesting short comminution ages in Taiwan Island. Taiwan Island has a rugged topography and high denudation rates (Dadson et al., 2003). Short river courses, steep relief and high monsoon rainfalls account for the fast sediment transfer from Taiwan Island to the sea (Kao and Milliman, 2008).

The ($^{234}\text{U}/^{238}\text{U}$) variation in the OT sediments, which represents the qualitative constraint on time scale of sediment transport based on comminution age theory, provides independent and robust evidence to verify the change of sediment source from East Asia continent and/or ECS shelf in the last glaciation to Taiwan Island during the Holocene. The time scale of sediment source-to-sink transport process can be quantitatively achieved with a reliable estimation of the fraction of ^{234}Th ejected due to α -recoil (i.e. f_α).

This study indicates that the transport times average approximately at 200 ± 100 kyr for the glacial lithogenic sediments in the OT, and <100 kyr for the Holocene sediments. The modern transport time for the Changjiang estuarine sediment is 120 ± 30 kyr or 160 ± 90 kyr based on a geometrical approach and its simplification with a constant grain shape factor, which is roughly comparable to the transport times for the last glacial sediments in the OT. Their differences in transport time between modern Changjiang sediment and marine sediment

could correspond to the transport time from the ECS shelf to the OT during the LGM because the current distance from sediment source region to the sea is about 600 km shorter than that during the LGM. Alternatively, the addition of sediments reworked from the exposed shelf (i.e. sediments that were deposited on the shelf during earlier high sea level stands) could also contribute sediment to the OT, which might have increased the sediment transport time defined by ($^{234}\text{U}/^{238}\text{U}$) in this study. Yet, the transport time for the Zhuoshui estuarine sediment is apparently long, about 146 ± 60 kyr or up to 210 ± 130 kyr derived from different calculations. In this regard, the upper sediments of core DGKS9604 in the OT are a better average of the Taiwan river-derived material, rather than a punctual sampling of the river because the river systems may vary considerably during different seasons, days or even in typhoon events. In addition, the large range of sediment transport time in the Zhuoshui Estuary suggests the significant uncertainties for acquiring the precise sediment transport time for Taiwan river sediments.

8. Conclusion

In this study, we estimate the sediment transport times in the central Okinawa Trough based on the comminution age theory. The core DGKS9604 sediments with a depositional age of about 27 ka were recovered for the measurement of ($^{234}\text{U}/^{238}\text{U}$) ratios and calculation of sediment transport time. The ($^{234}\text{U}/^{238}\text{U}$) ratios for the core sediments show a clear temporal trend, which is relatively low and uniform between 27 and 14 ka, while stay high and closer to secular equilibrium after 7 ka. The variations of ($^{234}\text{U}/^{238}\text{U}$) in the OT provide qualitative constraints on the time scale and thus the pathways of sediment transport from land to sea, which correspond well with the changes of sediment provenance reconstructed by independent evidences of clay mineralogy, element geochemical and Sr–Nd isotopic data. The calculated transport time for the OT sediments is longer (approximately 200 ± 100 kyr) between 27 and 14 ka, but relatively short (<100 kyr) after 7 ka. The longer transport time before 14 ka implies that the detrital sediments were predominantly derived from East Asian continent and/or the ECS shelf, while the shorter transport time after 7 ka suggesting the sediment source predominantly from Taiwan Island.

Our study confirms that the comminution age estimates based on ($^{234}\text{U}/^{238}\text{U}$) can provide deep insight into the source-to-sink transport of lithogenic sediments from land to ocean margins. However, this study also highlights the current limitations of this method and the need for more precisely estimating the recoil loss factor (f_{α}) and ($^{234}\text{U}/^{238}\text{U}$) of parent rocks (A_0), in order to obtain more robust and quantitative results. Nonetheless, the relatively constant grain size and geochemically-constrained changes in sediment provenance of the OT facilitate the calculation and interpretation of comminution ages in this unique river-dominated marginal setting. Regardless the large uncertainties in sediment transport time calculation, this study offers the first constraint on the time scale of sediment source-to-sink transport process in East Asia continental margin, which may greatly improve our understanding on the late-Quaternary land–sea interaction, and is also an important supplementary to U-series disequilibrium community.

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Appendix A. Supplementary data

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