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What is This?
Holocene shifts in riverine fine-grained sediment supply to the East China Sea Distal Mud in response to climate change

Bangqi Hu,1,2 Zuosheng Yang,1 Shuqing Qiao,1,3 Meixun Zhao,4 Dejiang Fan,1 Houjie Wang,1 Naishuang Bi1 and Jun Li2

Abstract
Holocene changes in fine-grained sediment supplies to the East China Sea outer shelf were uncovered, through the mineralogical and geochemical analysis of Core B3 in the East China Sea Distal Mud (ECSDM). Based on the lithology, accelerator mass spectrometry (AMS) 14C dating, and sea-level change, Core B3 can be divided into two major units: transgressive stage (Unit 1: 12.5–6.8 kyr) and highstand stage (Unit 2: 6.8–0 kyr). Significant discrepancy of dolomite/calcite ratio in the fine fractions (<16 µm) of Changjiang (dolomite/calcite = 3:1) and Huanghe (dolomite/calcite = 1:2.2) sediments was used as a new uniqueness provenance tracer to distinguish these two riverine sources. Both of the dolomite/calcite ratio and rare earth elements fractionation parameters in the fine-grained sediment indicated distinct provenance shifts of Core B3 during the Holocene. Unit 1 of Core 3 (12.5–6.8 kyr) mainly consists of the reworked and resuspension sediments of the East China Sea shelf during the Holocene transgression, while Unit 2 sediments (6.8–0 kyr) are most likely sourced from the Changjiang and Huanghe. Moreover, mixing curves of dolomite/calcite ratio reveal that the ECSDM continuously received the Changjiang sediment since 6.1 kyr with notable fluctuations, whereas the Huanghe sediment supply began in 6.8 kyr but abruptly stopped during 4.2–0.8 kyr and then appeared again since 0.8 kyr. Temporal changes of the Changjiang and Huanghe fine-grained sediment contribution to the ECSDM are closely related to the formation of modern oceanic circulation system since 6.8 kyr (shelf sea-level change), the ‘4.2 kyr’ climate event, and the followed transition to cold and dry climate condition in the northeastern China (global climate change), as well as the artificial shift of lower Huanghe course in AD 1128 in the war against invasion of the northern nomadic nation (human activities).

Keywords
carbonate minerals, climate change, East China Sea, Holocene, human activities, provenance, rare earth elements

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Introduction
The East China Seas (ECSs) is one of the largest epicontinental seas in the world, consisting of the Bohai Sea (BS), Yellow Sea (YS), and ECS (Figure 1). It is surrounded by Kyushu and the Ryukyu Islands, Taiwan Island, mainland China, and the Korean Peninsula, and characterized by a well-ventilated, shallow (<130 m on average), and wide (500 km) shelf. Until recently, the ECSs annually received ~1600 Mt sediments from the Changjiang (Yangtze River) and Huanghe (Yellow River), occupying ~10% of the global riverine sediment flux to the oceans (Milliman and Farnsworth, 2011). Under the context of huge amounts of terrestrial sediment supply and the modern oceanic circulation, several muddy depositional areas have developed in the broad and shallow shelf of ECSs during the Holocene (Figure 1) (Alexander et al., 1991; Li et al., 2005; Milliman et al., 1989; Qin and Li, 1983; Xu et al., 2012; Yang and Liu, 2007). These muddy depositional areas have been considered as sinks of the surrounding riverine sediments, mostly from the Changjiang and Huanghe, since the deposition rate of the vast sandy area on the ECS shelf is practically zero (DeMaster et al., 1985; Hu et al., 2011a; Li et al., 2012b; Lim et al., 2007; Nitrourer et al., 1984). Owing to the unique geographic location, the ECSs are influenced by both the high-latitude Northern Hemisphere (East Asian monsoon) and the tropic (Kuroshio Current) climatic changes. Muddy deposits in the ECSs can thus provide insight into the detailed land–ocean-climate interactions on geologic time scales (e.g. Ge et al., 2014; Hu et al., 2012b, 2014; Li et al., 2009a; Qiao et al., 2011; Wang et al., 2011a; Xiang et al., 2006; Xing et al., 2013; Yuan et al., 2013).

The East China Sea Distal Mud (ECSDM) (or Southwestern Cheju Island Mud in some references), located on the northeastern margin of ECS outer shelf, has been widely documented over the past three decades (Alexander et al., 1991; DeMaster et al., 1985; Lee and Chough, 1989; Milliman et al.,

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1985, 1989; Yang and Milliman, 1983). Formation of the ECSDM is closely related to the seasonal dynamics of regional ocean circulation formed since the Holocene highstand (c. 7 kyr) (Hyun et al., 2006; Kim and Lim, 2014; Xing et al., 2013; Yoo et al., 2002). With modern (~100 years) accumulation rates ranging from 2 to 5 cm/yr, the ECSDM is viewed as a modern material sink or fine-sediment depocenter in the ECS outer shelf (Alexander et al., 1991; DeMaster et al., 1985; Li et al., 2012b), and thus offered potential archives of high-resolution natural and anthropogenic records (Li et al., 2009a; Qiao et al., 2011; Xiang et al., 2006; Xing et al., 2013; Yang et al., 2009a). Recent researches further highlight that the ECSDM is the sink of organic pollutants (Guo et al., 2006), and also plays a key role in the fate, transport, and burial of organic carbon (Deng et al., 2006). Addressing temporal and spatial dispersal pattern of the riverine sediments and associated biochemical elements (McKee et al., 2004; Walsh and Nittouer, 2009), as well as deciphering paleoenvironmental changes archived in the shelf muddy deposits (Bianchi and Allison, 2009; Mendes et al., 2012), require first in-depth understanding of the sediment provenance and its possible temporal variability. The ECSDM sediments are believed to be mainly sourced from the Chinese large rivers (i.e. Changjiang and Huanghe) (Liu et al., 2003; Milliman et al., 1985; Yang and Milliman, 1983; Youn and Kim, 2011) and partially from Korean small rivers (Jung et al., 2012; Lim et al., 2006; Xu et al., 2009; Yang et al., 2003). In contrast, there are considerable uncertainties for distinguishing the Changjiang and Huanghe sediments in the potential mixed-sources areas (e.g. the ECSDM) using the traditional bulk geochemical and mineralogical methods because of the complex hydrodynamic processes and the possible influence of grain size (Jung et al., 2012; Yang et al., 2003). For example, geochemical and mineralogical compositions are found to vary with grain size in the Changjiang and Huanghe riverine sediments (Qiao and Yang, 2007; Yang et al., 2009b; Zhou et al., 2014b). Selection of relevant grain-size fraction in the source areas is thus a necessary precondition of the correct comparison between the ECSs muddy sediments and the potential sources. In addition, magnetic properties recently have been proposed to discriminate the Changjiang and Huanghe sediments (Li et al., 2012a; Wang et al., 2009; Zhang et al., 2008b). When using these magnetic-based approaches in the ECSs, however, several processes (hydrodynamic sorting, diagenesis, redox reaction, and bacterial magnetosomes) may have limited the provenance credibility (Li et al., 2012a; Liu et al., 2010; Wang et al., 2010). So far, there is lack of a fast, low-cost uniqueness provenance tracer, to the authors’ knowledge, for distinguishing the Changjiang and Huanghe sediments in the marine environment, especially in the mixed-sources areas.

In this study, we present the combined mineralogical and geochemical records of fine-grained riverine sediments supply to the ECSDM during the Holocene. Our objectives are threefold: (1) to provide a new uniqueness proxy for discriminating the Changjiang and Huanghe sediments; (2) to identify the sediment provenance; and (3) to identify the possible temporal variability.
provenance shifts of the ECSDM during the Holocene, especially changes in the Changjiang and Huanghe sediment influx; and (3) to illustrate the response of sedimentary environment in the ECSDM to multiple changes in the sea level, the shelf water circulation, and the sediment supply from the Changjiang and Huanghe because of the global climate events and anthropogenic activities in the river basins. The results will help to deepen understanding of the sedimentary evolution history of ECS and its response to the global changes during the Holocene.

Regional setting

Oceanography

The modern circulation pattern in the ECSs is largely governed by seasonal monsoon wind fields, the Changjiang run-off, and the Kuroshio Current (KC) (Figure 1) (Lee and Chao, 2003; Yuan et al., 2008). The warm and salty KC flows northward along the ECS shelfbreak. In winter, the Chinese coastal cold waters (i.e. SDCC, YSCC, ECSCC, see Figure 1) flow southward forced by strong northerly winds, whereas the Korean coastal currents always flow southward and the Changjiang Diluted Water (CDW) flowing mainly northeastward (Yuan et al., 2008) (Figure 1). The Yellow Sea Warm Current (YSWC) is the main component of the YS circulation in winter, and transports warm and saline water of the open ocean into the YS. Historically, the YSWC is thought to be a branch of the Tsushima Warm Current (TC) moving northward along the YS trough. Later studies indicated that this current is highly time dependent and is just a compensating current of the wind driven coastal currents in the YS in winter (Yuan and Hsueh, 2010). The Taiwan Warm Current (TWC) flows northeastward over the ECS middle shelf, which provides warm and saline water to the Fujian and Zhejiang offshore areas and the Changjiang mouth.

The transport and fate of fine sediments and associated materials in the ECS are thought to be strongly influenced by seasonal hydrodynamics, resulting from the East Asian monsoon system (Milliman et al., 1986, 1989; Yang et al., 1992). In winter, sediments off the Old Huanghe delta are frequently resuspended by storms and transported in a southeastward direction by the strengthened YSCC. These resuspended sediments supply the main sediment sources in the ECSDM, which are then trapped by the cyclonic circulation and its upwelling effects (Hu, 1984; Mao et al., 1983). In contrast, sediment deposition flux in the ECSDM is very weak in the summer because of the limited supply of suspended sediments and weak coastal current (Guo et al., 2002; Milliman et al., 1985, 1986).

Surrounding rivers

The Huanghe, with a drainage area of 0.79 × 10^6 km² and a length of 5464 km, used to be the second largest river in the world in terms of sediment load (Milliman and Syvitski, 1992) (Figure 1). It originates at the Bayan Har Mountains of Qinghai-Tibet Plateau and flows through the Loess Plateau and the North China Plain before debouching to the ECSs. The climate of the Huanghe basin is typically subtropical, wet and warm in summer and moist and cool in winter. The average annual precipitation and evaporation in the Huanghe Basin are 1000–1400 mm and 700–800 mm, respectively. The Changjiang annually delivered 900 km³ of water discharge and 480 Mt of sediment load to the ECS (Milliman and Syvitski, 1992). Most of the Changjiang sediment discharge to the sea mainly originates from the upper reaches, whereas water discharge from the upper reaches accounts for only 50% of that at Datong (Hu et al., 2009, 2011c; Yang et al., 2006). In the tributary basins of the upper reaches, the source rock types are dominated by Paleozoic carbonate rock, Mesozoic detrital rocks (mudstone and red sandstone), igneous rocks (granite and basalt), and so on. (Yang et al., 2004). Additionally, carbonates are widely spread throughout the Changjiang basin and are particularly abundant in the southern region of the middle reaches and in the upper and middle reaches of the Hanjiang (Chen et al., 2002).

There are also many small rivers from Korea Peninsula and Taiwan Island debouched into the ECSs. The Korean river basins are located in the area of 34°N–38°N and 126°E–128°E, and their bedrock consists predominantly of Precambrian gneiss and Jurassic and Cretaceous granitic, with local limestone, schist, volcanic rock, and phyllite (Yang et al., 2003). The Korean rivers are much smaller than the Chinese rivers, both in size and in water and sediment fluxes, and annually delivered a total of c. 5–25 Mt sediments to the YS (Lim et al., 2007; Yang et al., 2003). Taiwan Island is located at the collision boundary between the Philippine Sea Plate and the Eurasian Continental Plate, consisting primarily of Tertiary marine sedimentary outcrops, such as sandstone, siltstone, and shale, with scattered limestone outcrops (Li et al., 2013). However, small mountainous rivers draining the Taiwan Island collectively discharge c. 184–380 Mt/yr of sediment to the surrounding ocean, resulting from its high relief and steep gradients, frequent tectonic activity, highly erodible rocks, and the recurrent typhoons-derived heavy rainfall (Dasdson et al., 2003; Kao and Milliman, 2008).

Materials and methods

River samples

Twelve samples collected from the Huanghe downstream and 11 samples from the Changjiang estuary were separated into seven particle-size fractions (<2 µm, 2–4 µm, 4–8 µm, 8–16 µm, 16–32 µm, 32–63 µm, 63–125 µm) in the laboratory. All the sediment samples were pretreated with 10% H₂O₂ for 24 h to remove organic matter and then dispersed in sodium hexametaphosphate solution for 24 h. The 125–63 µm and 63–32 µm fractions of riverine samples were separated by wet sieving, and then the minus sieve of the samples were sub-fractionated into five fractions (<2 µm, 2–4 µm, 4–8 µm, 8–16 µm and 16–32 µm) using a settling method based on the Stokes’ law.

Core samples

A 230-cm-long giant piston core (B3) was recovered from a water depth of 62 m in the ECSDM at 31°30’N, 125°45’E in 2003 (Figure 1). In the laboratory, the core was split, described, and subsampled.
in 2-cm intervals for grain-size analysis. A total of 56 samples were selected to extract the <16 µm fraction by the same procedure as the riverine samples. Mixed species of planktonic foraminifera from seven horizons were picked for accelerator mass spectrometry (AMS) ¹⁴C dating at the Woods Hole Oceanographic Institution, US. All radiocarbon dates were calibrated to calendar years before present (cal. BP, 0 cal. BP = AD 1950) using the online program Calib 7.0, with an updated calibration curve Marine13 (Reimer et al., 2013) and a constant average global reservoir age of 400 years (Table 1).

Sediment grain-size analysis

Grain-size analyses were conducted using a Mastersizer-2000 laser particle-size analyzer at the Key Laboratory of Marine Sedimentology and Environmental Geology, First Institute of Oceanography, State Oceanic Administration, China, with a measurement range of 0.02–2000 µm and a size resolution of 0.01µm. The measuring error was within 3%. Before the grain-size analyses, the samples were pretreated with 10% H₂O₂ and 0.5 mol/L HCl for 24 h to remove organic matter and biogenic carbonate, respectively.

Mineral analysis

The riverine and Core B3 subsamples were analyzed by x-ray diffraction (XRD) from 5 to 65° 2θ with Ni-filtered Cu-Kα radiation (40 kV, 25 mA) using a step size of 0.02° 2θ and a counting time of 2 s per step on a Rigaku D/max-rA type diffractometer at Key Laboratory of Marine Hydrocarbon Resources and Environmental Geology, Qingdao Institute of Marine Geology. Semi-quantitative estimate of mineral percentages were calculated based on the peak areas of the x-ray diffractograms, using the MDI Jade 6.5 software. About 20% of these samples were replicated to determine the analysis error, which is generally less than 10%.

Elemental analysis

A total of 20 subsamples of Core B3 were oven-dried at 60°C overnight, and then powdered and homogenized with an automated agate mortar and pestle. The powdered sediments (200 mg) were digested with a mixture of HF–HClO₄ for 24 h in a tightly closed Teflon beaker on a hot plate at less than 180°C. The dry samples were then reacted with 1 mL HNO₃ to remove the residual HF, and then digested with a mixture of 3 mL 50% HNO₃ and 1 mL Rh (500 ppb) for 24 h in a tightly closed Teflon vessel in an oven at 150°C. The solutions were analyzed for rare earth elements (REEs) by inductively coupled plasma-mass spectrometer (ICP-MS, Thermo X series) at the Key Laboratory of Marine Hydrocarbon Resources and Environmental Geology, Qingdao Institute of Marine Geology. Blanks and China Stream Sediment Reference Materials (GSD9) were included in the analyses for data QA/QC, and the accuracy between the determined and certified values of GSD9 were better than 5%.

Results

Core lithology and age model

The lithology of Core B3 varied upward from sand (172–230 cm), silty sand (172–124 cm), sandy silt (124–106 cm), to clayey silt (106–0 cm). Core B3 can be generally divided into two major units (Figure 2): Unit 1 (12.5–6.8 kyr) and Unit 2 (6.8–0 kyr). The bottom lithology, Unit 1, is characterized by sand and silty sand with shell debris, with an averaged sand contents of 72% and mean grain size (Mz) of 4.4Φ. Unit 2 can be further divided into two sections: the lower section, Unit 2-1 (6.8–2.1 kyr), displays an upward-fining trend along with sand content and Mz increased from 70% to 10% and from 5Φ to 7.5Φ, respectively. While the upper section, Unit 2-2 (2.1–0 kyr), consists of homogeneous silt (mean of 64%) and clay (mean of 35%), with stable Mz of 8Φ.

The radiocarbon ages of Core B3 at depth of 24–26 cm, 70–72 cm, 86–90 cm, 114–118 cm, 138–142 cm, 164–168 cm, and 200–202 cm were 783, 2155, 6941, 6889, 6941, 6889, and 9889 cal. yr BP, respectively (Table 1). The disturbed ages can be ascribed to the tidal mixing process that affected the study area during the postglacial transgression period (Hyun et al., 2006; Kim and Lim, 2014; Yoo et al., 2002). An age model was thus produced by linear interpolation between these calibrated ages, excluding sample B3-5. Average sedimentation rates (SR) were calculated by dividing the depth range (cm) by the age range (kyr BP) between a series of age points. The fine-grained (<16 µm) sediment flux of Core B3 was calculated using the average sedimentation rates and 23 dry bulk density (DBD) data (sampled in 10 cm intervals) as follows (Figure 2c):

\[ \text{Flux} = \frac{<16\mu m(\%)}{100 \times DBD \times SR} \]

The fine-grained (<16 µm) sediment flux of Core B3 varied largely from 1.7 to 18.9 mg/cm²/yr, with mean of 10.1 mg/cm²/yr. Considering the AMS ¹⁴C dating and lithological characters of Core B3, together with micropaleontological and seismic stratigraphic results of previous studies (Hyun et al., 2006; Kim and Lim, 2014; Yoo et al., 2002), two major units of Core B3 are respectively corresponded to transgressive stage (Unit 1: early Holocene transgression sand facies) and highstand stage (Unit 2: mid-to-late Holocene shelf mud facies) in the ECS shelf, which is closely related to the global sea-level changes (Figure 2d). The upward-fining trend of the Unit 2-1 (6.8–2.1 kyr) may be related to the stepwise weakening of YSCC, consistent with a reduced intensity of the East Asian Winter Monsoon since c. 7 kyr (Hu et al., 2012b), while Unit 2-2 is thought to be formed under cold eddy controlled hydrological conditions since 2.1 kyr when the direct influence of the Kuroshio was reduced (Xing et al., 2013).

Carbonate minerals (calcite and dolomite)

The relative contents of calcite and dolomite in different size fractions of the Huanghe and Changjiang are presented in Table 2 and Figure 3. For the Huanghe samples, the calcite percentages show a steadily increasing trend as the grain size decreased. It increased from 5% in the 125–63 µm fraction to

<table>
<thead>
<tr>
<th>Sample</th>
<th>Depth (cm)</th>
<th>Conventional ¹⁴C age (years)</th>
<th>Calibrated age (cal. yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B3-1</td>
<td>24–26</td>
<td>1240 ± 40</td>
<td>783</td>
</tr>
<tr>
<td>B3-2</td>
<td>70–72</td>
<td>2490 ± 35</td>
<td>2155</td>
</tr>
<tr>
<td>B3-3</td>
<td>86–90</td>
<td>3060 ± 45</td>
<td>2827</td>
</tr>
<tr>
<td>B3-4</td>
<td>114–118</td>
<td>4440 ± 60</td>
<td>4619</td>
</tr>
<tr>
<td>B3-5</td>
<td>138–142</td>
<td>6450 ± 50</td>
<td>6941</td>
</tr>
<tr>
<td>B3-6</td>
<td>164–168</td>
<td>6410 ± 50</td>
<td>6889</td>
</tr>
<tr>
<td>B3-7</td>
<td>200–202</td>
<td>9130 ± 60</td>
<td>9889</td>
</tr>
</tbody>
</table>

Table 1. AMS ¹⁴C ages and calendar ages of Core B3 from the East China Sea outer shelf.

AMS: accelerator mass spectrometry.
approximately 25% in <2 µm and 2–4 µm fractions. Meanwhile, the dolomite percentages are much lower and varied slightly, from 0.4% to 2.5% (Figure 3a). In contrast, both of the calcite and dolomite percentages of the Changjiang samples show a slowly decreased trend with the decreasing grain size (Figure 3b). The maximum of calcite and dolomite percentages in the Changjiang sediments were 7% and 5% in 32–63 µm. In particular, calcite was not detected in the clay fractions (<2 µm) of Changjiang sediments, consistent with that of Yang and Milliman (1983) and Milliman et al. (1985). The calcite and dolomite percentages in the <16 µm particle-size fractions exhibit significant discrepancy between the Huanghe and Changjiang. The calcite is the dominant carbonate minerals in the <16 µm fractions of Huanghe, whereas the dolomite became much more enriched in the <16 µm fractions of Changjiang. The average calcite and dolomite percentage are 22% and 1% in the <16 µm fractions of Huanghe sediment, and 1% and 3% in the <16 µm fractions of Changjiang sediment, respectively. The carbonate mineral contents of Core B3 are presented in Table 3 and Figure 4. Calcite was detected in two layers of 0–28 cm (1–6%) and 108–166 cm (1–3%) of Core B3, with mean of 2% (Figure 4a), and dolomite only appeared in the upper section (0–148 cm) of Core B3, with mean of 1% (Figure 4b).

REEs

The average REE concentrations of core B3 and the reference materials (the Changjiang, Huanghe, Korean Rivers, and Taiwanese Rivers) are presented in Table 4. REEs are lithophile refractory elements, which exhibit a unique and coherent behavior during weathering, erosion, and transportation processes because of similarity in their electronic configuration (McLennan, 1989). To facilitate data comparison and to assign a pattern to the sources of REE, these concentrations are normalized to chondrite and upper continental crust (UCC) (Taylor and McLennan, 1985). The Eu and Ce anomalies (δEuN and δCeN) are two important REE parameters, which reflect the weathering of source rocks and thus link to detrital sediment sources (McLennan, 1989). In this study, δEuN and δCeN were calculated by comparing the measured concentrations of Eu and Ce with their neighboring elements:

\[
\delta \text{Eu}_N = \frac{\text{Eu}_{\text{SmGd}}}{\text{Sm}_{\text{Gd}}} 
\]

\[
\delta \text{Ce}_{N} = \frac{\text{Ce}_{\text{LaPr}}}{\text{La}_{\text{Pr}}} 
\]

Hereinafter, the subscript N indicates the chondrite normalization.

The total REE concentrations (ΣREE) in the <16 µm fractions of Core B3 ranged between 51.8 and 179.4 µg/g. Light rare earth elements (ΣLREE, from La to Eu) (46.1–160.4 µg/g) are more enriched than heavy rare earth elements (ΣHREE, from Gd to Lu) (5.9–19.0 µg/g), with the ratios of LREE/HREE varying from 6.7 to 8.4 (Figure 5). All samples generally exhibit the LREE-enriched chondrite-normalized patterns with (La/Yb)N varying between 6.9 and 9.6, which are characteristic of continental crustal material. The δEuN and δCeN values varied slightly downcore, with little or no Ce anomalies (δCeN = 1.01 ± 0.03) and moderate Eu depletions (δEuN = 0.72 ± 0.02) (Table 4). UCC-normalized REE fractionation parameters (e.g. (La/Yb)UCC, (Gd/Yb)UCC, and (La/Sm)UCC) are used to quantify the fractionation between LREEs and HREEs. All of B3 subsamples are characterized by convex pattern enriched by relatively middle REEs (specially Sm and Eu), as marked by higher values of (Gd/Yb)UCC relative to (La/Yb)UCC and (La/Sm)UCC. In addition, the ratios of LREE/HREE, (La/Yb)UCC, and (Gd/Yb)UCC exhibit regular variations in Core B3 with two remarkable changes occurring at 6.8 and 2.1 kyr, respectively (Figure 5).

**Figure 2.** (a) Vertical lithology profile, (b) mean grain size, (c) and fine-grained (<16 µm) sediment flux of Core B3 in the East China Sea shelf. (d) The global sea-level curve for the last 12 kyr is also presented (Stanford et al., 2011).
### Discussion

#### Dolomite/calcite ratio of fine-grained sediment (<16 µm) as a new uniqueness provenance tracer for the Changjiang and Huanghe

Most of the Huanghe sediments come from the easily erodible Loess Plateau and inherit most characteristics of the Loess. Therefore, the Huanghe sediments contain high content of calcite, especially in the fine fractions (~25%), Table 2 and Figure 3a). Additionally, weak chemical weathering in the Huanghe basin is conducive to calcite preservation in the Loess and Huanghe sediment. In contrast, because of the complicated and variable source rock types in the Changjiang basin, it is difficult to relate the Changjiang sediments to some specific source rock types (Yang et al., 2004). Li and Qin (1991) and Yang et al. (2003) concluded that in the Changjiang sediments, carbonate minerals are more constant compared with calcite during the weathering process, resulting from the larger lattice energy of dolomite. This difference leads to most of calcite in the Changjiang sediments (Table 2 and Figure 3b). Yang et al. (2004) indicated that significant correlation occurred between mean grain size and CaO percentage of the Huanghe sediments, whereas no correlation between them exists in the Changjiang sediments, in consistent with our results.

Although calcite and dolomite have similar crystal lattice structures, the dolomite is more constant compared with calcite during the weathering process, resulting from the larger lattice energy of dolomite. We proposed that the dolomite/calcite ratio in <16 µm fractions can be used as a new uniqueness tracer to distinguish the Changjiang and Huanghe sediments (Figure 4c). This provenance tracer possesses several advantages as follows. First, the <16 µm particle-size fractions of the Changjiang and Huanghe sediments belong to the evenly suspended load (Fan et al., 2002) and can be easily transported by marine hydrodynamics as indicated by the numerical modeling (Zhu and Chang, 2000). Therefore, this is the reliable grain-size fraction to trace the Changjiang and Huanghe sediments in the ECSs. Second, high content of calcite in bulk sample was regarded as a possible source indicator of the Huanghe because it is highly enriched in the Huanghe sediments but depleted in the Changjiang and Korean riverine sediments (Milliman et al., 1985; Yang, 1988; Yang and Milliman, 1983). However, the biogenic calcite in the ECSs sediments can make it confusing to identify the detrital one coming from the Huanghe by the presence of calcite (Li and Qin, 1991; Yang et al., 2003). In this regard, the carbonates in <16 µm can largely exclude the influences of marine biogenic carbonate (mostly >63 µm) (Chen et al., 2005). Although the coccolithophores (consisted by pure calcite) were basically finer than 20 µm, their influences on the total calcite content of other sediment should not be so significant in the marginal seas such as the ECSs (Li and Qin, 1991; Zhang et al., 2008a). Furthermore, the dolomite, of detrital origin, would not be influenced by the biogenic carbonates (Li et al., 2007; Ravaiolli et al., 2003). Third, there are no carbonate minerals peaks (both calcite and dolomite) on the XRD patterns in the <20 µm fractions of the Korean (Song and Choi, 2009) and Taiwanese riverine sediments (Chen et al., unpublished data). Therefore, using the dolomite/calcite ratio in <16 µm sediment fractions for sediment provenance tracer would effectively remove the disturbance of the Korean and Taiwanese riverine sediments (Figure 3b). This simple analytical methodology (XRD) to determine the calcite and dolomite contents in the fine-grained sediment

<table>
<thead>
<tr>
<th>Rivers</th>
<th>Samples</th>
<th>Calcite percentage</th>
<th>Dolomite percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>&lt;2 µm</td>
<td>2–4 µm</td>
<td>4–8 µm</td>
</tr>
<tr>
<td>Changjiang</td>
<td>C01</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>C02</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>C03</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>C04</td>
<td>0.1</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>C05</td>
<td>0.0</td>
<td>0.0</td>
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<td>Average</td>
<td>24.6</td>
<td>25.1</td>
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</table>

Table 2. The calcite and dolomite contents (%) in the different fractions of the Changjiang and Huanghe sediments.
Provenance discrimination of Core B3 fine-grained sediments (<16 µm)

Asian dust storms, annually occurring in late winter and spring, originate from three primary source areas, namely, the Taklimakan desert in the northwestern China, the Bajain Juran and Tengger deserts in northern China, and the Gobi desert in southern Mongolia (Sun et al., 2001, 2008). The ECSDM is located at the downwind area of Asian dust storms, and thus possibly influenced by dust deposition. Previous studies estimated that the total deposition flux of eolian dust into the ECS is about 2.1–2.7 mg/cm²/yr (Gao et al., 1997; Hsu et al., 2009). It is much lower than the fine-grained sedimentation flux of Core B3 during 6.8–0 kyr (mean of 12.5 mg/cm²/yr), but comparable with that of 12.5–6.8 kyr (mean of 3.3 mg/cm²/yr) (Figure 2c). This seemingly suggests that dust depositions play an important contribution to the Core B3, at least during the early Holocene. However, both of bulk and fine-grained fraction dust samples from the primary source areas are characterized by high carbonate contents (especially calcite) (Ferrat et al., 2011; Honda et al., 2004; Li et al., 2007; Wang et al., 2005), in contradiction with no-detected calcite and dolomite in Core B3 before 6.8 kyr (Figure 4). Moreover, the eolian quartz flux variations in the Cheju Island (Figure 1) during the last 6.5 kyr indicated that the high eolian quartz flux occurs between c. 4.0–2.0 kyr, with a maximum level at c. 3.1 kyr (Lim and Matsumoto, 2008; Lim et al., 2005). This situation is completely different from the temporal variations of carbonate minerals in Core B3, which have no detectable calcite during 4.2–0.84 kyr (Figure 4). As a result, dust depositions into the ECS are unlikely to dominate the ECSDM sedimentation during the Holocene; thus, we will not consider and discuss the dust contribution in the following sections.

To determine the provenance of fine-grained sediments of Core B3, we first compared their REE fractionation patterns with those of potential riverine source areas. In the ECs, the REE composition characteristics of the Chinese and Korean rivers have been well investigated, and results show that the concentrations and distribution patterns of REEs are distinctly different in Chinese and Korean riverine sediments, resulting from the different basement rocks (Li et al., 2013; Song and Choi, 2009; Xu et al., 2009). However, recent studies emphasized that besides the source rock composition, REE concentrations in riverine and marine sediments are largely influenced by many other factors, especially the heavy minerals (Jung et al., 2012; Song and Choi, 2009; Yang et al., 2002a). Great caution should be taken when using the REE distribution patterns as provenance tracer to discriminate the sandy and coarse silty marine sediments (Jung et al., 2012). As only one size-separated fraction (<16 µm) was used in this study, grain size and heavy mineral effects on the REE distribution patterns could be practically removed. Therefore, the REE fractionation ratios and UCC-normalized patterns in this study seem to be reliable proxies for assessing sediment provenance. All of fine-grained sediments in Core B3 exhibited convex trend in terms of UCC-normalized distribution pattern, which is similar with that of Chinese rivers, but distinctly different from that of Korean rivers (Song and Choi, 2009; Xu et al., 2009). Significant differences in REE fractionation patterns between Unit 1 (before 6.8 kyr) and Unit 2 (after 6.8 kyr) (Figure 5) suggest that these two depositional units may have different sediment provenances. Fine-grained sediments of Unit 1 have lower LREE/HREE, (La/Yb)UCC, and (Gd/Yb)UCC as compared with that of Unit 2. The shift of fine-grained sediment provenance since 6.8 kyr was further confirmed by a discrimination plot of (La/Sm)UCC versus (Gd/Yb)UCC (Figure 6). Unit 1 samples plot in another group and were different from any of potential source end members, and thus inferred to be most likely sourced from the ECS shelf during the transgression (Yoo et al., 2002). In contrast, Unit 2 samples overlap with the Changjiang and Huanghe end members, indicating that it is a mixed product from the Changjiang and Huanghe. This is contradicted with a more recent study by Kim et al. (2013), which suggested that considerable surface sediments of ECSDM were probably transported from the Korean rivers, and Taiwan or the Northwest Pacific.

It can be seen that distinguishing the Changjiang and Huanghe sediments using REE alone is very difficult, since they have fairly similar and overlapped REE compositions characteristics (Table 4). In contrast, the dolomite/calcite ratio of fine-grained sediment (<16 µm) can efficaciously overcome this disadvantage to distinguish the Changjiang and Huanghe sources in Core B3. As shown in Figure 4, carbonate minerals in the fine-grained sediments of Core B3 were not detected in Unit 2 (12.5–6.8 kyr). This indicates that there are almost no Changjiang and Huanghe sediments supplied for Core B3 during 12.5–6.8 kyr, in agreement with REE discrimination results (Figure 6). Using the dolomite/calcite ratios of Changjiang (3:1) and Huanghe (1:22) as their end-member values, an interesting finding emerged when the dataset of Unit 1 was plotted in Figure 4c. It indicates that the Changjiang and Huanghe sediments supplied to the ECSDM were temporarily changed over the past 6.8 kyr. The Changjiang sediments were continuously supplied to Core B3 since 6.1 kyr, whereas the influence of Huanghe sediments initially occurred at 6.8 kyr, stopped in 4.2–0.84 kyr, and then appeared again since 0.84 kyr. Furthermore, the mixing curves of Changjiang and Huanghe sediments can be quantitatively calculated, suggesting that the Huanghe is the major contributor to the ECSDM during 6.1–4.2 and 0.84–0 kyr (mostly >75%) (Figure 4c).
Table 3. Calcite and dolomite contents in the fine-grained (<16 μm) sediment of Core B3 from the East China Sea outer shelf.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Cal. kyr BP</th>
<th>Calcite (%)</th>
<th>Dolomite (%)</th>
</tr>
</thead>
<tbody>
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<td>2</td>
<td>2.00</td>
<td>2.5</td>
<td>0.9</td>
</tr>
<tr>
<td>5</td>
<td>0.07</td>
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<td>0.9</td>
</tr>
<tr>
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<td>0.5</td>
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<td>1.5</td>
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<td>0.9</td>
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<td>1.4</td>
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<td>1.2</td>
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The provenance shifts of Core B3 also reflected in variations of the fine-grained sediment flux during the Holocene (Figure 2c). The fine-grained sediment fluxes of Unit 1 (before 6.8 kyr, mean of 3.3 mg/cm²/yr) are much lower than that of Unit 2 (after 6.8 kyr) because of the inefficiency riverine sediment supplies. After 6.8 kyr, the fine-grained sediment fluxes of Core B3 increased sharply to 9.1 mg/cm²/yr in Unit 2-1 (6.8–2.1 kyr), mainly resulting from the continued contribution of Changjiang sediments since 6.1 kyr. The maximal level of fine-grained sediment fluxes in Core B3 after 2.1 kyr may be related to the intense human interventions in the Changjiang and Huanghe basins. Several studies indicated that human activities in the upper and middle Changjiang have effectively increased the Changjiang sediment load from ~240 to ~480 Mt/kyr after 2.0 kyr, along with the increase in Changjiang delta progradation rates abruptly from 38 to 80 km/kyr (Hori et al., 2001; Wang et al., 2011b). When human intervention on the Loess Plateau was almost negligible, the provenance changes of Core B3 from the East China Sea outer shelf.

Land–ocean-climate interactions of ECSDM during the Holocene

In this section, we will provide a comprehensive description of how several competitive processes (e.g. sea level, oceanic circulation, monsoon climate, drainage changes) affected the sedimentary environmental changes in the ECSDM during the Holocene (Figure 7).

The ECS shelf is largely exposed during the Last Glacial Maximum (LGM, 30–19 kyr), when the global sea level is about 120–135 m below present sea level (b.p.s.l.) (Clark and Mix, 2002). During the postglacial transgression, changes in the coastline configuration have caused shifts of the ECS tidal fields, with tidal currents on the ECS shelf during the low sea-level period being more energetic and exerting stronger influence on the sea floor than present (Chen and Zhu, 2012; Uehara and Saito, 2003). Along with the sea level rise from ~90 to 0 m b.p.s.l., high tidal bottom stress regions (>1.0 N/m²) migrated shoreward from the Cheju Island toward the west coast of Korea, and along the retreating tidal polynya during the postglacial transgression period (Oguri et al., 2000; Ujiié et al., 2001). For the study area, the tidal bottom stress was lower than 0.35 N/m² at sea level below ~15 and 0 m b.p.s.l. (e.g. 8–7 kyr) (Uehara and Saito, 2003), in accordance with the transition time of Core B3 (c. 6.8 kyr) (Figure 2). Unit 1 sediments of Core B3 are interpreted to be the product of reworking and resuspension of shoreline sediments during the postglacial transgression (Hyan et al., 2006; Yoo et al., 2002). This inference is supported by sedimentological and micropaleontological records of the adjacent cores and shows a major environmental change from an early-Holocene transgressive systems tract (TST) to a mid-to-late Holocene highstand systems tract (HST) occurring at c. 7 kyr (Kim and Lim, 2014). It is also the time point when the present-day oceanic circulation patterns were finally established (c. 6–7 kyr) (Fang et al., 2013; Li et al., 2009b; Xiang et al., 2008). None of the Changjiang and Huanghe sediment supplies to the ECSDM during 12.5–6.8 kyr can thus be explained by the combined effects of the lack of transport media (no YSCC) before 6.8 kyr and the distant Changjiang and Huanghe river mouth (see below discussion) (Figure 7a).
With the postglacial rapid sea-level rise, the paleo-Changjiang retreated landward, and then shifted into the SYS at c. 12 kyr (Li et al., 2000, 2002). However, the presence of Yangtze Shoal may have effectively prevented the southward dispersal of Changjiang-derived sediments (Xia and Liu, 2001). As the sea-level rise continued, the Yangtze Shoal was submerged and the Changjiang Estuary retreated further westward from the SYS to the Jiangsu coast (Li et al., 2000, 2002). The Changjiang delta was developed after the rapid sea level at 9–8 kyr, with the initial stage dominated by aggradation (8–6 kyr), and thus the Changjiang sediments were largely trapped within the incised valley (Hori et al., 2001, 2002; Song et al., 2013). When the sea level kept stable after 6 kyr, the Changjiang delta depocenters moved seaward and formed a series of seaward migrating river-mouth sand bars from Zhenjiang to the present river mouth (Song et al., 2013). This is the time period when the mud wedge along the ECS inner shelf began to deposit (Liu et al., 2007b) and also corresponded well to the results of this study (Figure 4b). Since c. 6.1 kyr, the Changjiang fine-grained sediments were delivered mainly by the southeastward YSCC in winter (Milliman et al., 1985, 1986; Yang et al., 1992; Yuan et al., 2008) and/or the northeastward CDW in summer (Kim et al., 2009; Moon et al., 2009) to the ECSDM (Figure 7b).

The paleo-Huanghe courses are more complex during the postglacial rapid sea-level rise. According to Xue et al. (2004), the Huanghe was deduced to empty into the Bohai Strait in 11.6–9.6 kyr, and the proximal Huanghe subaqueous delta accreted over the relict or transgressive facies along the northern shore of Shandong Peninsula (Figure 7a) (Liu et al., 2007a). After an extreme flooding of the Huanghe drainage basin at 9.2 kyr (Huang et al., 2007; Yang et al., 2000), the Huanghe has frequently shifted its lower course north-southward and formed a subaqueous deltaic deposits in the offshore area of SYS during the period of 9.6–8.5 kyr (Xue et al., 2004), as well as eight superlobes in the western plain of BS since c. 7.8–8.2 kyr (Saito et al., 2000) (Figure 7). However, our results indicate that the Huanghe sediments continuously contributed to the ECSDM during 6.8–4.2 kyr (Figure 4c), implying that the paleo-Huanghe also flowed through the

**Table 4.** Comparisons of rare earth elements (REEs) characterization in the fine-grained sediments (<16 µm) of Core B3 with those of the Changjiang, Huanghe, Korean (<20 µm, Song and Choi, 2009), and Taiwan Island (<63 µm, Li et al., 2013) riverine sediments.

<table>
<thead>
<tr>
<th>Samples</th>
<th>ΣREE (µg/g) ± SD</th>
<th>ΣLREE/ΣHREE</th>
<th>δEuN</th>
<th>δCeN</th>
<th>(La/Yb)UCC</th>
<th>(La/Lu)UCC</th>
<th>(Gd/Yb)UCC</th>
<th>(Gd/Lu)UCC</th>
<th>(La/Sm)UCC</th>
</tr>
</thead>
<tbody>
<tr>
<td>B3-total</td>
<td>96.4 ± 32.1</td>
<td>7.7 ± 0.5</td>
<td>0.72 ± 0.02</td>
<td>1.01 ± 0.03</td>
<td>0.91 ± 0.08</td>
<td>0.93 ± 0.09</td>
<td>1.26 ± 0.06</td>
<td>1.29 ± 0.07</td>
<td>0.80 ± 0.04</td>
</tr>
<tr>
<td>B3-Unit 1 (6.8–0 kyr)</td>
<td>97.7 ± 35.4</td>
<td>7.9 ± 0.3</td>
<td>0.72 ± 0.02</td>
<td>1.01 ± 0.03</td>
<td>0.95 ± 0.06</td>
<td>0.98 ± 0.06</td>
<td>1.29 ± 0.03</td>
<td>1.32 ± 0.04</td>
<td>0.81 ± 0.04</td>
</tr>
<tr>
<td>B3-Unit 2 (12.1–6.8 kyr)</td>
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<td>7.1 ± 0.3</td>
<td>0.73 ± 0.02</td>
<td>1.03 ± 0.02</td>
<td>0.81 ± 0.03</td>
<td>0.82 ± 0.03</td>
<td>1.20 ± 0.05</td>
<td>1.22 ± 0.06</td>
<td>0.77 ± 0.03</td>
</tr>
<tr>
<td>Huanghe (n = 3)</td>
<td>201.1 ± 50.8</td>
<td>9.5 ± 0.6</td>
<td>0.64 ± 0.03</td>
<td>1.02 ± 0.04</td>
<td>1.09 ± 0.10</td>
<td>1.06 ± 0.09</td>
<td>1.28 ± 0.06</td>
<td>1.25 ± 0.06</td>
<td>0.80 ± 0.04</td>
</tr>
<tr>
<td>Changjiang (n = 10)</td>
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<td>9.0 ± 0.2</td>
<td>0.67 ± 0.01</td>
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<td>1.08 ± 0.04</td>
<td>1.07 ± 0.04</td>
<td>1.26 ± 0.05</td>
<td>1.25 ± 0.05</td>
<td>0.84 ± 0.04</td>
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<tr>
<td>Korean Rivers (n = 18)</td>
<td>252.3 ± 23.1</td>
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<td>1.57 ± 0.05</td>
<td>1.53 ± 0.05</td>
<td>1.48 ± 0.04</td>
<td>1.44 ± 0.04</td>
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<tr>
<td>Taiwanese Rivers (n = 38)</td>
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<td>1.06 ± 0.16</td>
<td>1.03 ± 0.17</td>
<td>1.25 ± 0.17</td>
<td>1.21 ± 0.22</td>
<td>0.98 ± 0.08</td>
</tr>
</tbody>
</table>

REE: rare earth element; LREE: light rare earth element; HREE: heavy rare earth element.
North Jiangsu Plain, and then transported by the southeastward YSCC to the ECSDM (Figure 7b). This is supported by Yang et al. (2002b), suggesting that the Huanghe supplied considerable sediments to the North Jiangsu coastal plain during the middle Holocene. Based on the synthesis of earlier works, Shi et al. (1992, 1993) reconstructed a comprehensive history of Holocene climate changes in China. According to their results, the Megathermal Period (wet and warm) occurred from ~9.5 to ~3.3 kyr with the Mid-Holocene Climatic Optimum (warmest and wettest) occurring at ~8.0 to ~6.8 kyr in northern China. Subsequently, several comprehensive reviews regarding the Holocene climate changes in northern China have been published (e.g. An et al., 2000, 2006; Feng et al., 2006; He et al., 2004; Ran and Feng, 2013; Wang and Feng, 2013; Zhao et al., 2009), and almost all of them indicated that in the semi-arid to sub-humid northern China, a humid climate generally characterized the early and middle Holocene, and then back to a dry climate, with an abrupt shift at c. 4.4–4.5 kyr. When the Huanghe after the Loess Plateau entered into the vast North China Plain, its lower courses most likely bifurcated into many distributaries and overflowed freely both northward to the BS and southward to the YS during the early and middle Holocene, resulting from the abundant monsoon rainfall and no river dykes existed.

Previous studies indicated that an abrupt climate change event centered on 4.2 kyr has played a key role in termination or collapse of several ancient civilizations over the Old World because of severe droughts (e.g. An et al., 2005; Booth et al., 2005; Cullen et al., 2000; Drysdale et al., 2006; Liu and Feng, 2012; Magny et al., 2009; Staubwasser et al., 2003). However, extraordinary flooding events at around 4.3–4.0 kyr have been widely documented in the middle Huanghe, such as Qishuihe River (Huang et al., 2011), Jinghe River (Huang et al., 2010, 2012), and Beiluohe River (Yao et al., 2008), and also in the Hanjiang River valley of the middle Changjiang (Zhang et al., 2013b). Moreover, both the Dongge cave in southwest China (Stalagmite D4) (Dykoski et al., 2005) and Lianhua cave in central China (Stalagmite LH2) (Zhang et al., 2013a) display the maximum growth rate but with relative heavier δ¹⁸O at 4.0–4.2 kyr, implying extreme precipitation events may have frequently occurred in the context of dry climate at that time. These facts confirm that extraordinary floods resulted from storm rains occurred commonly over the monsoon-influenced region of China at around 4.2–4.0 kyr, suggesting that severe droughts and extreme floods may be parts of the climatic variability during the 4.2 kyr event, at least in the semi-arid to sub-humid region (Huang et al., 2010, 2011, 2012).

In Chinese history, the legend of ‘the Great Yu controls the waters’ was well known for solving the 4.2 kyr megaflood

Figure 5. Downcore variations of REEs fractionation parameters of the fine-grained sediments of Core B3 from the East China Sea outer shelf.

Figure 6. Discrimination plot of (La/Sm)UCC versus (Gd/Yb)UCC for the fine-grained sediments of Core B3 (<16 µm). Values of the Changjiang, Huanghe, Korean (<20 µm; Song and Choi, 2009), and Taiwan Island (<63 µm; Li et al., 2013) riverine sediments are also shown for comparison.
problems (Chen et al., 2012) and promoting the establishment of first Chinese dynasty Xia (start from 2204 to 1989 BC) (Wagner et al., 2013). The successfulness of Great Yu is mainly attributed to climate transition from wet to dry since c. 4 kyr (Wu and Ge, 2005), in accordance with the paleo-vegetation reconstruction records in the northwestern Loess Plateau (Zhao et al., 2009) and other monsoon precipitation proxies (Dykoski et al., 2005; Zhang et al., 2013a). Benefiting from this, the Great Yu led people to successfully stop the Huanghe from flooding by dredging the river channel (Wu and Ge, 2005), thereby flowing along the Yuwang Course (2070–602 BC) toward the BS (Figure 1) (Yuan, 1998). After being delivered into the BS, parts of these Huanghe sediments can be resuspended by strong winter storm (Yang et al., 2011) and carried by coastal current eastwards into the North YS along north of Shandong Peninsula (Bi et al., 2011), and then turns southwards into the South YS (Yang and Liu, 2007). However, most of them cannot be carried to the south of 35°N (Hu et al., 2011a; Yang and Liu, 2007), and hardly be transported to the ECS outer shelf and Okinawa Trough (Bian et al., 2010). In fact, our recent study of two Cores from the SYS also illuminated that the influence of Huanghe sediments on the central SYS is rather limited since 6.4 kyr (Li et al., 2014) because of the trapping effect of oceanic fronts. Thus, abrupt disappearance of the Huanghe sediments in Core B3 at 4.2 kyr may be a direct response to this large drainage transformation, resulting from both the climatic change and anthropic activities in the river basin.

The lower Huanghe mainly flowed into the BS until AD 1128, when the south banks were artificially destroyed by Song Dynasty (AD 960–1279) to check the invasion of the nomadic Jurchen horse army from northern China (Yuan, 1998). In AD 1128–1855, the lower Huanghe course shifted southward to the SYS for more than 700 years, forming a large delta lobe along the Jiangsu coast (Superlobes 10, the Old Huanghe delta, Figure 7b) (Zhang, 1984). Although the Huanghe shifted northward into the BS in AD 1855, the Old Huanghe delta has experienced a severe marine erosion because of strong tidal regime in the BS during the past ~150 years, resulting in more than 20 km shoreline retreat and providing a huge amount of sediments to the SYS (Saito et al., 1994). Recent study suggested that offshore erosion in the Old Huanghe delta can provide more than 790 Mt/yr of sediments, and about half of them is transported southeastward along the mud belt that crosses the central YS to south of Cheju Island in the ECS (Zhou et al., 2014a). In Core B3, the Huanghe sediments contribution appeared again since c. 0.8 kyr (Figure 4c), corresponding well to the dramatic drainage changes that occurred in the Huanghai basin. This suggests that over the past ~800 years, the Huanghe sediments (directly input or lately erosion) are resuspended by strong northwesterly wind and then transported by the YSCC to the ECSDM in winter (Milliman et al., 1985, 1986; Yang et al., 1992).

Conclusion

In this study, we performed mineralogical and geochemical analyses of riverine and Core sediments to illustrate the history of the fine-grained sediment supplies from the Huanghe and Changjiang to the ECS outer shelf during the Holocene. Given the large discrepancy of carbonate minerals (calcite and dolomite) that existed in the fine-grained sediments (<16 µm) of Changjiang and Huanghe, a new provenance proxy, dolomite/calcite ratio in <16 µm fractions, was proposed to distinguish the Changjiang and Huanghe sediments. Both the mineralogical and geochemical evidences suggest an abrupt shift of fine-grained sediment provenance in the ECS outer shelf since 6.8 kyr. The older
sediments (12.5–6.8 kyr) were most likely sourced from the reworked and resuspension sediments of the ECS shelf during the Holocene transgression, while the younger sediments (6.8–0 kyr) are mixing products of the Changjiang and Huanghe. The mixing curves of Changjiang and Huanghe sediments were further calculated using dolomite/calcite ratios of Changjiang (3:1) and Huanghe (1:2), and clearly indicated that the Changjiang and Huanghe sediment supplies to the study area were temporarily changed during the last 6.8 kyr, in response to the global changes such as sea-level change, climate event, and human activities in the river basins.

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