1 Holocene vegetation dynamics in response to climate change and hydrological processes

2 in the Bohai region

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13 ABSTRACT

14 Coastal vegetation both mitigates the damage inflicted by marine disasters on coastal areas and plays an important role 15 in the global carbon cycle (i.e. blue carbon). Nevertheless, detailed records of changes in coastal vegetation composition 16 and diversity in the Holocene, coupled with climate change and river evolution, remain unclear. To explore vegetation 17 dynamics and their influencing factors on the coastal area of the Bohai Sea (BS) during the Holocene, we present high-18 resolution pollen and sediment grain size data obtained from a sediment core of the BS. The results reveal that two rapid 19 and abrupt changes in salt marsh vegetation are linked with the river-system changes. Within each event, a recurring 20 pattern-starting with a decline in Cyperaceae, followed by an increase in Artemisia and Chenopodiaceae-suggests a 21 successional process that is determined by the close relationship between Yellow River (YR) channel shifts and the 22 wetland community dynamics. The phreatophyte Cyperaceae at the base of each sequence indicate lower saline 23 conditions. Unchannelized river flow characterized the onset of the YR channel shift, caused a huge river-derived 24 sediment accumulation in the floodplain, and destroyed the sedges in the coastal depression. Along with the formation 25 of a new channel, lateral migration of the lower channel stopped, and a new intertidal mudflat was formed. Pioneer

species (Chenopodiaceae, *Artemisia*) were the first to colonize the bare zones of the lower and middle marsh areas. In addition, the pollen results revealed that the vegetation of the BS land area was dominated by broadleaved forests during the early Holocene (8500–6500 a BP) and by conifer and broadleaved forests in the middle Holocene (6500–3500 a BP), which was followed by an expansion of broadleaved trees (after 3500 a BP). The pollen-based temperature index indicated that a warmer early and late Holocene and colder middle Holocene were consistent with previously reported temperature records for East Asia. The main driving factors of temperature variation in this region are insolation, the El Ni ño-Southern Oscillation and greenhouse gases forcing.

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34 *Keywords*: Coastal salt marsh; Pollen; Delta superlobe; Temperature; El Niño-Southern Oscillation

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36 **1. Introduction**

37 Coastal areas, where cities, populations and industries are clustered, are playing an increasingly critical role in trade 38 globalization (Hemavathi et al., 2019). Because they are located between marine ecosystems and terrestrial ecosystems, 39 coastal areas are prone to many natural hazards such as flooding, storms, and tsunamis (Hou and Hou, 2020). Coastal 40 vegetation, which acts as a natural barrier, is widely distributed in coastal areas and could effectively mitigate the damage 41 caused by marine disasters to the economy and environment of coastal areas (Zhang et al., 2018). Moreover, despite their 42 relatively small global extent (between 0.5 and 1×10^6 km²), coastal vegetation ecosystems, tidal marshes, mangroves, 43 and seagrasses play an important role in the global carbon cycle (Serrano et al., 2019; Spivak et al., 2019). Per unit area, 44 their organic carbon sequestration rates exceed those of terrestrial forests by 1-2 orders of magnitude and contribute 45 ~50% of carbon sequestered in marine sediments (Serrano et al., 2019). Hence, it is important to understand the long-46 term spatial-temporal dynamics of coastal vegetation, which are favorable for the global carbon cycle research and 47 coastal restoration.

Climatic fluctuation, post glacial sea-level rise and changes in river discharge provoked dramatic habitat changes
along coastal areas during the Late Pleistocene and Holocene (Neumann et al., 2010; Cohen et al., 2012; Pessenda et al.,
2012; Fran ça et al., 2015). Presently, the relationship of sea-level change and coastal vegetation (especially mangrove)

evolution has been widely studied by many researchers (e.g. Engelhart et al., 2007; Gonza lez and Dupont, 2009; Fran aet al., 2012; Woodroffe et al., 2015; Hendy et al., 2016). Contrarily, studies on the long-term dynamics of coastal vegetation, coupled with climate change and river evolution, are sparse. During the Holocene, the global rivers delivered large amounts of material to the ocean, the total suspended sediment delivered by all rivers to the ocean was approximately 13.5×10^9 tons annually (Milliman and Meade, 1983). The material transported by the rivers had huge impacts on the coastal ecosystem. Hence, a deeper understanding of correlations between coastal vegetation and river variables is required to better assess coastal vegetation responses to global warming in the future.

58 In the coastal areas of the Bohai Sea (BS), vegetation is dominated by warm temperate deciduous broadleaved 59 forests and shrub grasslands (Wang et al., 1993). The Yellow River (YR), as one of the largest river in the world in terms 60 of sediment discharge (Milliman and Meade, 1983), transports large amounts of sediment into the BS every year; hence, 61 it has developed a delta complex in the west coastal region of the BS since 7000 a BP (He et al., 2019). Deposition of 62 the Yellow River delta (YRD) complex resulted in the formation of a vast area of floodplain and estuarine wetland (Xue 63 et al., 1995; Cui et al., 2009; Liu et al., 2009b). Based on the study of coastal vegetation of the BS, it is helpful to 64 understand the spatial and temporal drivers of ecological variability, and thus of the vegetation-climate and river 65 relationship, especially wetland dynamics. However, there have been few studies investigating the vegetation dynamics 66 and their response to climate and river variables in the Bohai region.

Pollen records have been useful in terms of reconstructing vegetation dynamics and environmental changes associated with climatic changes in the geological record (Bao et al., 2007; Cohen et al., 2008; Giraldo-Giraldo et al., 2018). In this study, we carried out a detailed investigation of core sediments from Laizhou Bay, BS. We analyzed pollen and grain size proxies under high resolution and refined the chronology of the core by ¹³⁷Cs and accelerator mass spectrometry (AMS) ¹⁴C dates. With this in mind, the specific objectives of the current research are formulated as follows: (1) to reconstruct the vegetation evolution history in the Bohai region and (2) to tentatively discuss the effects of climate and environment on coastal vegetation (especially wetlands) during the Holocene.

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75 **2. Study area**

76 2.1. Geographical settings

The BS, a shallow inland sea in China, is connected with the Yellow Sea through the narrow Bohai Strait (Figure 1). The main rivers flowing into the BS are the YR, Haihe River, Luanhe River, and Liaohe River. Among these, the YR is the largest and is the main source of sediments in this region. Over the past 2000 years, the YR has annually provided approximately 1.1×10^9 tons of sediment discharged into the BS (Milliman et al., 1987). This immense amount of sediment has resulted in the rapid seaward progradation of YRD, and a rapid change in the location of the main distributaries in the lower delta plain.

The tidal current plays a critical role in the transportation and distribution of sediments in the BS. The tidal currents of the modern BS are dominated by semi-diurnal tides. The velocity of tidal currents varies from 20 to 80 cm/s. Three strong tidal current areas are observed in the northern Bohai Strait, the central part of Bohai Bay, and the eastern part of Liaodong Bay (Huang et al., 1999). In Laizhou Bay close to the core location, the speed of tidal currents is weak (Gu and Xiu, 1996).

The wind waves off the YRD are dominated by the East Asian monsoon and show significant seasonal variations. The prevailing northerly winds are much stronger in winter than the dominant southerly winds in summer. Strong winter winds cause strong wind waves, and thus strong bottom shear stresses which readily erode seabed sediment into water (Yang et al., 2011; Wang et al., 2014; Zhou et al., 2017).

92 The circulation of the BS is weak, and the mean flow velocity is small. In winter, the predominant extension of the 93 Yellow Sea Warm Current (YSWC) intrudes and crosses the Bohai Strait, moving westward along the central part of the 94 BS and splits into two branches. One branch moves toward the northeast to form a clockwise gyre (Liaoxi Coastal 95 Current [LXCC]), and the other veers southward and then turns eastward along the southern coast to form a 96 counterclockwise gyre (Libei Coastal Current [LBCC]). In summer, the YSWC disappears in the BS, and eddies 97 generated in the BS are stronger than in winter. During this time, the central eddy is missing, the eddy in Laizhou Bay is 98 more pronounced, and the coastal current along the southern and western coastlines of the BS is established (Figure 2; 99 Liu et al., 2015; Yang et al., 2016).

101 **2.2.** Climate and vegetation

The Bohai region lies in a zone of warm temperate monsoonal climate with distinctive seasons. The annual mean air temperature is 9.5–13.1 °C. The annual average precipitation is about 600 mm, and 60–70% of the total annual precipitation occurs between June and August (Qiao et al., 2012). As the Liaodong Peninsula and Shandong Peninsula protrude into the sea, they are clearly influenced by its proximity and experience sufficient rainfall. There is less rainfall in the mountain area of the northern part (Wang et al., 1993).

107 The regional vegetation is dominated by warm temperature deciduous broadleaved forests and shrub grasslands. 108 Currently, natural vegetation only remains in the mountain areas because of widespread anthropogenic activities (e.g. 109 cultivation and farming). The predominant deciduous broadleaved species belong to *Ouercus*, such as *O. liaotungensis*, 110 O. acutissima, and O. mongolica. Co-dominant plants are Pinus, including P. densiflora that grows in the coastal humid 111 area, and P. tabuliformis that is distributed in the relatively dry North China plain. In the plain area, apart from P. 112 tabuliformis, there are some deciduous broadleaved trees, such as Ailanthus altissima, Koelreuteria paniculata, and 113 Morus alba. Other broadleaved trees, Betula ermanii, Populus tremula, Acer spp., Tilia amurensis, and Carpinus 114 turczaninowii are distributed in the hills and lowlands (Wang et al., 1993). The coastal wetlands are occupied by herbs 115 and shrubs, such as Tamarix chinensis, Salix matsudana, S. integra, Phragmites australis, Aeluropus, Limonium sinense, 116 Suaeda glauca, Typha orientalis, and Acorus calamus (Wang et al., 1993; Li et al., 2007; Xu et al., 2010).

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118 **3. Materials and methods**

119 **3.1.** Coring, sub-sampling, and chronology

Core CJ06-435 was collected in Laizhou Bay, BS in August 2007 by the R/V *Kan407* of the Shanghai Bureau. The
 core site is located at 37.50 N, 119.52 E, at a water depth of 14.6 m (Figure 1); the core had a length of 271 cm. In the
 laboratory, the core was spilt into two sections, photographed, macroscopically described, and sub-sampled.

Isotopes ¹³⁷Cs and ²¹⁰Pb were measured employing EG&G Ortec Gamma Spectrometry at the Nanjing Institute of Geography and Limnology, Chinese Academy of Sciences (NIGLAS). The sediment samples were air-dried and pulverized. ¹³⁷Cs and ²¹⁰Pb concentrations were then determined from gamma emissions at 662 and 46.5 keV. respectively. In addition, a total of 10 samples consisting of foraminifera were obtained from the core for radiocarbon dating. The radiocarbon dating was conducted at the Woods Hole Oceanographic Institution (WHOI) and Beta Analytic Inc., USA. Radiocarbon dates were corrected for the regional marine reservoir effect ($\Delta R = -139 \pm 59$ years, a regional average value determined for the BS) and calibrated using the Calib 7.04 program (Stuiver et al., 2019) with one standard deviation uncertainty (1.0 × σ) (Table 1).

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132 **3.2.** Palynological and sediment grain size sample analysis

133 A total of 127 samples were selected for pollen analyses. Each sample was oven-dried at 60 °C for 24 h. The dry-134 weight of the samples ranged from 2.5 g to 13.9 g. Samples were chemically treated according to the procedure outlined 135 by Faegri and Iversen (1992). Before treatment, a standard tablet of Lycopodium spores (mean=18,583+764 spores per 136 tablet) was added to each sample to aid in the calculation of palynological concentrations. Samples were treated with 137 15% HCl solution to remove carbonates, boiled in 10% KOH solution for 5 min to remove humic acids, then treated 138 with 40% HF to remove silicates. The residue was mounted in glycerin jelly. Fossil pollen was identified and counted 139 with a light microscope at 400 × magnification. A minimum of 200 pollen grains were counted for each sample. The palynological concentrations of per gram sediment (PCP) were calculated using the follow equation: 140

141 $PCP = \frac{18583}{Lycopodium \text{ number per slide}} * \frac{\text{Pollen or Spore Counts per slide}}{\text{Net weight of dry sample}}$

142 The percentage of each pollen type was calculated from the total sum of pollen and spores. The pollen diagram 143 was produced using the Tilia program, and the pollen assemblage zones were divided based on the results of a constrained 144 cluster analysis (CONISS) within Tilia (Grimm, 1987).

Analysis of sediment grain size was performed at 2.0 cm intervals throughout the core using a Malvern Mastersizer 2000 instrument at the laboratory of the First Institute of Oceanography. The chemical procedure of grain size experimental pretreatment was consistent with the procedures described by Chen et al. (2019a). A solution of 30% H₂O₂ and 1.0 mol/L HCl were added to decompose the organic matter and remove carbonates.

150 **4. Results**

151 **4.1. Chronological model**

Measurements of ¹³⁷Cs and ²¹⁰Pb revealed activity at the top of the profile, indicating the recovery of recently deposited sediments. ¹³⁷Cs is a bomb-derived radionuclide, first appearing in environmental samples at measurable levels around 1954 with the onset of nuclear weapon testing (Kirchner and Ehlers, 1998), and was most prevalent in 1963 (the year of maximum fallout from atmospheric weapon testing) (Palinkas and Nittrouer, 2007). Sub-surface peaks are not discernible in ¹³⁷Cs profiles of core CJ06-435 (Figure 3). However, the deepest onset of ¹³⁷Cs is an effective marker of the year 1954 (25 cm).

Often, the combined data of ¹³⁷Cs and excess ²¹⁰Pb have been used to calculate the sedimentation rates (Wu et al., 2015). Excess ²¹⁰Pb shows a downward decline owing to the decay of ²¹⁰Pb when the sediment stably accumulates for an appropriate period, and the excess ²¹⁰Pb activity could be used to calculate the sedimentation rate. However, the excess ²¹⁰Pb profiles of core CJ06-435 did not show a clear downward decline trend (Figure 3), and excess ²¹⁰Pb in the upper parts of the core is not that large when compared with the lower background values. Thus, the ²¹⁰Pb data seemed to be unsuitable for estimating the sedimentation rate of core CJ06-435. The ¹³⁷Cs–derived average sedimentation rate was 0.47 cm/yr in the upper 25 cm of core CJ06-435.

165 The results of AMS radiocarbon dating are shown in Table 1 and Figure 3. Three samples above the 20 cm depth 166 were not included in the age model because their ¹⁴C age was anomalously greater than the ¹³⁷Cs dating. The dating point 167 of 129 cm was eliminated because it appears not to be reliable. According to the result of He et al. (2019), the calculated 168 sedimentation rate (CSR) in the tidal flat and neritic area of the south BS ranged from 0.02 to 0.13 cm/year before 2000 169 a BP (calculation from cores H9601, H9602, ZK228, and ZK1, Figure 1). If the 129 cm dating is correct, the CSR would 170 be as high as 0.45 cm/yr in the section of 87–129 cm (4801–4894 a BP) for core CJ06-435. It is apparently not reasonable 171 because core CJ06-435 is offshore compared to the other cores (e.g. H9601, H9602, ZK228, and ZK1, Figure 1) reported 172 in previous researches (Xue et al., 1988; Saito et al., 2000; Li et al., 2013). It should have a lower CSR compared to 173 those cores rather than an approximate ten-fold increase in CSR. The calibrated dates of several other samples are plotted 174 against sediment depth and shown in Figure 4.

176 **4.2. Sediment grain size distributions**

177 The grain size parameters and components percentages show distinct variations. The mean grain size and the median 178 grain size both show high values at depths of 271–160 cm, 135–83 cm and 19–0 cm, lower values at a depth of 83–34 179 cm, and the lowest values at depths of 160–135 cm and 34–19 cm. There was a smaller proportion of clay in the lower 180 profile (271–160 cm) than in the upper profile (160–0 cm, except for the two sections of 160–135 cm and 34–19 cm). 181 The sequences of silt and sand contents showed a strong inverse association. There were high proportions of silt and low 182 proportions of sand at depths of 271–222 cm, 180–160 cm, 135–83 cm, 40–34 cm and 19–0 cm, lower proportions of 183 silt and higher proportions of sand at depths of 222–180 cm and 83–40 cm, and the lowest proportions of silt and the 184 highest proportions of sand occurred at depths of 160–135 cm and 34–19 cm (Figure 3).

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186 4.3. Palynology assemblage

A total of 71 pollen taxa were identified, among which *Pinus, Quercus*, Cyperaceae and *Typha* were the most dominant taxa in the lower part (271–156 cm) of the core, and *Pinus, Quercus*, Poaceae, Compositae, *Artemisia*, Chenopodiaceae and Cyperaceae were the most dominant taxa in the upper part (156–0 cm) of the core. With respect to the fern spores, *Selaginella sinensis* and Polypodiaceae were dominant; however, their content was low throughout the core. With the aid of CONISS, the whole sequence was vertically divided into three zones, with zone 2 further divided into subzones 2a, 2b, 2c and 2d (Figures 4 and 5).

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194 **4.3.1.** Palynological zone 1 (271–156 cm)

The palynological zone 1 was characterized by abundant broadleaved trees pollen, dominated by *Quercus* (mean 18.7%), *Betula, Alnus, Pterocarya*, Ulmaceae and Moraceae (Figure 4). Percentages of conifer pollen were relatively low compared with other zones: *Pinus* ranges from 19.7% to 45.6% (mean 33.6%), and Taxodiaceae was present only occasionally. Compared to the other zones, the proportions of non-arboreal pollen types, Compositae (mean 1.2%), *Artemisia* (mean 4.2%), and Chenopodiaceae (mean 5.6%) were lowest in this zone, whereas the proportions of Cyperaceae (mean 10.3%) and *Typha* (mean 11.2%) were highest in this zone. The palynological concentrations were
high, varying between 6050 and 237 grains/g (Figure 5).

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203 **4.3.2.** Palynological zone 2 (156–30 cm)

204 Palynological zone 2 was divided into four subzones:

From depth of 156 to 128 cm (subzone 2a), the percentage of *Pinus* pollen reached its maximum (mean 46.6%), whereas the percentage of broadleaved trees *Quercus* (16.1–9.5%, mean 14%), *Betula*, *Alnus*, *Pterocarya*, Ulmaceae, and Moraceae decreased to different degrees. The proportions of non-arboreal pollen types, Compositae (mean 2.5%), *Artemisia* (mean 6.6%), and Chenopodiaceae (mean 7.4%) were higher, whereas the percentages of Cyperaceae (mean 7.2%) and *Typha* (mean 3.1%) declined sharply (Figure 4). Total pollen concentration were lower than in zone 1, especially in the interval of 156–135 cm, where the value of total pollen concentrations (62–1306 grains/g, mean 485 grains/g) was at its minimum in the core (Figure 5).

From depth of 128 to 63 cm (subzone 2b), *Pinus* (49.4–27.3%) and *Quercus* (18.1–7.9%, mean 13.5%) pollen level decreased to a low point, and the relative abundance of *Betula* slightly increased. Occasionally, there were small amounts of *Pterocarya*, Ulmaceae, and Moraceae. Non-arboreal pollen types, Compositae, *Artemisia*, and Chenopodiaceae

continuously increased to averages of 3.5%, 6.7%, and 12%, respectively (Figure 4). Pollen concentration increased up
to a high abundance (mean 1260 grains/g) in this subzone (Figure 5).

From depth of 63 to 41 cm (subzone 2c), the percentage of *Pinus* pollen started to decrease steadily, and the percentage of *Quercus* (17.4–11.8%, mean 14.8%), *Betula*, *Alnus*, *Pterocarya*, and Ulmaceae pollen slightly increased.

219 Similar to subzone 2b, this subzone had relatively high quantities of non-arboreal pollen such as Compositae, *Artemisia*,

and Chenopodiaceae (Figure 4). Pollen concentrations varied between 456 and 1381 grains/g (Figure 5).

In contrast, subzone 2d (41–30 cm) was marked by a sudden decrease in the pollen of *Quercus* (9.8%), and a steep increase in the pollen of *Pinus* (41.1%), even though the percentage of the non-arboreal were the same as those of subzone 2c (Figure 4).

4.3.3. Palynological zone 3 (30–0 cm)

This zone was characterized by the transition from dominance by the pollen of arboreal taxa to non-arboreal types. The percentage of *Pinus* and *Quercus* pollen decreased to the lowest level, averaging approximately 19.7% and 5.5%, respectively. Poaceae, Compositae, *Artemisia*, and Chenopodiaceae pollen increased, with average values of *Artemisia* and Chenopodiaceae reaching up to 24.6% and 21.1%, respectively (Figure 4). The total pollen concentration declined to 188 grains/g in the lower part of the zone (30–19 cm) and then increased slightly (to approximately 621 grains/g) at the top (Figure 5).

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233 **5. Discussion**

5.1. Key terrestrial palynomorphs proxies of environmental and climatic change

235 In sediment core CJ06–435, both *Pinus* and *Ouercus* pollen were the predominant pollen types among the arboreal 236 taxa. In order to understand the pollen provenance, the pollen records in the surface sediments of Laizhou Bay were 237 studied (Yang et al., 2016), and the concentration and percentage data of main pollen species were presented on a regional 238 map (Figure 6). Pollen results of surface sediments revealed that higher values of *Pinus* and *Ouercus* are usually found 239 in the eastern part of Laizhou Bay, and the lowest values of Pinus and Quercus occur in the nearshore area outside the 240 mouth of the YR (Figure 6a and 6b). The distributions of Pinus and Quercus pollen in the surface sediments of Laizhou 241 Bay are closely related to the distribution of the nearshore epicontinental vegetation. Except for the YRD, where there 242 is swamp and cultivated land, the epicontinental region of the Laizhou Bay is surrounded by pine and oak forests. Among 243 these, the land to the east of the Laizhou Bay (the Shandong Peninsula) belongs to the southern warm temperate zone 244 and mainly supports a pine-oak forest dominated by *Pinus densiflora* and *Q. acutissima* (Wang et al., 1993). The land 245 to the northeast of the Laizhou Bay (the Liaodong Peninsula) belongs to the southern temperate zone, and it mainly 246 developing a conifer and broadleaved mixed forest dominated by P. densiflora, O. mongolica and O. acutissima (Li et 247 al., 2007; Xu et al., 2010). The ubiquitous distribution of these plants on the adjacent terrain explains why *Pinus* and 248 *Ouercus* are the most common pollen taxa in Laizhou Bay, and why the highest concentration and percentage of *Pinus* 249 and Quercus occurred in the eastern part of Laizhou Bay and the lowest values of Pinus and Quercus occurred on the

250 nearshore area outside the mouth of the YR.

251 Previous studies have revealed that *Pinus* and *Ouercus* were the most common components of the forests in 252 northeast China (including the land areas surrounding the BS) during the Holocene. The variation of *Pinus* and *Ouercus* 253 contents were closely related to the change of temperature (Ren and Zhang, 1998; Yi et al., 2003; Li et al., 2004; Xu et 254 al., 2014; Zhang et al., 2019). Ren and Zhang (1998) investigated pollen data from northeast China and found that 255 *Ouercus* and *Ulmus* were the dominant components of the forests between 10000 and 5000 a BP, while *Pinus* were much 256 sparser, indicating warmer and drier summers in northeast China for the early to mid-Holocene. A high-resolution 1000-257 year pollen record from the Sanjiaowan Marr Lake in northeast China revealed that *Ouercus* is an effective indicator for 258 temperature reconstructions. Several notable cold periods, with lower *Ouercus* frequencies, occurred at approximately 259 1200 AD, 1410 AD, 1580 AD, 1770 AD and 1870 AD (Zhang et al., 2019). Another 5350-year pollen record from an 260 annually laminated maar lake in northeast China revealed a decrease of *Quercus* and increases of the *Pinus* component; 261 this indicates a cooling trend during the past 5350 years (Xu et al., 2014). Based on these results, we conclude that the 262 variation of *Pinus* and *Ouercus* pollen of core CJ06-435 may be also related to temperature change.

263 Herb pollen, especially Chenopodiaceae, also occupies an important position in core CJ06-435 (Figure 4). The 264 spatial distribution of herb pollen in surface sediment of Laizhou Bay suggests that a higher percentage and concentration 265 occur in the nearshore area close to the YR estuary and the southwestern part of Laizhou Bay, and a low percentage and 266 concentration are found in the eastern part of Laizhou Bay (Figure 6c and 6d). The YR is the main sediment source of 267 the BS. The annual mean sediment load of the YR was 1.08×10^9 tons before dam construction (Milliman and Meade, 268 1983), 70–90% of which was deposited and formed a huge delta complex (Zhou et al., 2016). Natural vegetation in the 269 modern YRD are dominated by wetland herbs, including Chenopodiaceae and Artemisia (Jiang et al., 2013). Furthermore, 270 under the combined action of the ocean and rivers, alluvial plains and coast plains developed widely along the southern 271 coast of Laizhou Bay. The terrestrial vegetation types in these areas change from a bare intertidal zone to seepweed 272 swamp to reed swamp to cultivated land from the shoreline landward (Xu et al., 2010). Since the transportation distance 273 for herb pollen is normally very short, the pollen percentage and concentration in samples close to the mouth of the YR

and the southwestern part of Laizhou Bay are much higher than in other samples (Figure 6c and 6d), indicating that herb
pollen of Laizhou Bay is mainly derived from the plant communities of the coastal wetlands. Therefore, the herb pollen
data in core CJ06-435 might contain some information about the evolution of wetlands.

277 It is worth noting that the composition of fossil pollen in sediment depends not only on the composition of the 278 vegetation from which the pollen originates but also on pollen dispersion, deposition and preservation, *Pinus* pollen is a 279 bisaccate grain, and has relatively high aerodynamic and hydrodynamic characteristics meaning it can be transported 280 efficiently by wind and water (Sander, 2001; Montade et al., 2011). Previous studies revealed that smaller amount of 281 *Pinus* pollen are found nearshore, and larger amounts are found in the deep ocean (Mudie, 1982; Mudie and McCarthy, 282 1994; Zheng et al., 2011; Dai et al., 2014; Luo et al., 2014; Dai and Weng, 2015). In Laizhou Bay, the concentration and 283 percentage of *Pinus* pollen followed similar patterns of distribution to broadleaved tree pollen (*Ouercus, Betula* and 284 *Carpinus*; Yang et al. 2016). However, in the eastern part of Laizhou Bay, *Pinus* pollen increased in a northeasterly 285 direction away from the coast (Figure 6a). Hence, concerning *Pinus* pollen data, caution is required because climate 286 variation alone may not be responsible for the change of *Pinus* pollen in marine sediment. Aerodynamic and 287 hydrodynamic conditions may also influence the amount of *Pinus* pollen in sediments.

288 In addition, because different pollen types are not equally well preserved (Havinga, 1967; Cheddadi and Rossignol-289 Strick, 1995), bias originating from poor preservation should be eliminated before using the net content of pollen grains 290 to reconstruct paleovegetation. In this study, the pollen concentration ranged from 62 to 6050 grains/g. Relatively low 291 pollen concentrations were found in the two sections (160–135 cm and 34–19 cm); this was largely correlated to high 292 sand contents as revealed by the lithology. Especially for the lower section (150–135 cm), the high portion of sand 293 content is consistent with a low pollen concentration and a high percentage of *Pinus* pollen. As the *Pinus* pollen is more 294 resistant to degradation, the variations of total pollen concentration as well as a higher percentage of *Pinus* pollen in this 295 section seem to be related to pollen preservation. But, as shown in Figure 4, the highest percentage of *Pinus* pollen was 296 recorded at a depth of 150–128 cm, with a low value at 148 cm, which is not completely in accordance with the high 297 sand content section in the same core (160–135 cm). Similarly, for the upper section, a high sand content was recorded 298 at a depth of 34–19 cm. However, the percentage of *Pinus* pollen is low in this section, except for a relatively high value at a depth of 23 cm. We thus suggest degradation is not a key point influencing the concentration of pollen and spore inthe study area.

301 Previous research suggests that the sedimentation mechanisms of pollen and spore in marine water is similar to that 302 of sediment with clay and fine silt grain size (Heusser, 1988). A recent investigation on the surface sediment from the 303 BS shows high pollen concentration in sediments with a high proportion of fine particles such as clay and silty clay. 304 while low pollen concentration in sediments with a high proportion of coarser sand particles (Yang et al., 2019). Yang et al. (2019) attributed the low pollen concentration in areas with a high sand content of the BS to the strong hydrodynamic 305 306 suspension and screening for sediments and pollen. We conclude that the low pollen concentrations in the two sections 307 (160–135 cm, and 34–19 cm), correlated with high sand content, could be attributed to the hydrodynamic conditions 308 rather than degradation.

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310 5.2. Sedimentary records indicative of river channel shifts

311 The most important geological events in the northern China coast after 7000 a BP were the shift of YR channel and 312 the formation of the YRD. The YR has been easily plugged and breached, and therefore its lower reaches migrated 313 because of its huge sediment load. The shifting of the lower reaches of the YR led to the formation of a new delta 314 superlobe (He et al., 2019). Based on a study of cheniers and historical documents, nine YRD superlobes have been 315 proposed by Xue and Cheng (1989) and Xue (1993) on the western shore of the BS. Among these, superlobe 1 (7000– 316 5000 a BP, He et al., 2019), superlobe 7 (11 AD-1048 AD, Xue, 1993), and superlobe 10 (1855 AD-present) are 317 positioned near the core area in this study. The information about some of these superlobes formation are recorded in 318 core CJ06-435.

As shown in Figures 4 and 7e, herb percentage abruptly increased at 34 cm and 160 cm. Herb pollen in the sediment of Laizhou Bay is mainly derived from the coastal wetlands of the western BS. In 6000–7000 a BP and 1855 AD, the YR emptied into the BS after a natural course shift, forming two huge delta superlobes in the western part of the BS (Saito et al., 2000). Wetland plants are the most important vegetation type in the YRD (Jiang et al., 2013). The development of YRD wetland would change the amounts of herb pollen that was transported to the study site. In addition, the formation of YRD caused the coastline to move closer to the position of the CJ06-435 core. Since most herb plants are small in size, their pollen grains are unable to disperse broadly (Chen et al., 2019b). The migration of the coastline would change the availability of herb pollen to the study site, and hence lead to variations in the amount of pollen. Therefore, combined with the age data, we conclude that the abrupt change of herb pollen percentage at 160 cm and 34 cm in core CJ06-435 is related to the formation of the YRD superlobe 1 and superlobe 10.

329 Compared with the pollen percentage, the pollen concentration can be interpreted in different ways. Namely, the percentage of different types of pollen is relative, whereas the pollen concentration is absolute, and it can directly reflect 330 331 the amounts of pollen that were transported to the study area (Luo et al., 2013). It is crucial that a correct interpretation 332 of pollen data is based on a percentage diagram as well as concentration. In core CJ06-435, the concentrations of herbs— 333 especially Chenopodiaceae and Artemisia (Figure 7f)—were higher at depths of 160–94 cm (6570–5000 a BP) and 34– 334 0 cm (after 1855 AD), except for the two sections of 160–135 cm and 34–19 cm. As mentioned in section 5.1, the 335 extremely low pollen concentration in the sections of 160-135 cm and 34-19 cm was closely linked with the coarser 336 sandy sediment. Combined with the results of pollen percentage and sediment grain size, we presumed that the higher 337 herb pollen concentration in the periods of 6570-5000 a BP (160-94 cm) and after 1855 AD (34-0 cm) reflects changes 338 in hydrographic conditions. Pollen data of surface sediments revealed that higher herb pollen concentrations occur in the 339 YR, and the value of these concentrations showed a decreasing trend starting from the river mouth toward the ocean. 340 The distribution pattern of surface pollen revealed that the YR is a major carrier for most herbs taxa in the sediment of 341 Laizhou Bay (Yang et al., 2016). At the site of core CJ06-435, which is close to the mouth of the YR in Laizhou Bay, 342 higher herb pollen concentrations in the Holocene samples may indicate increased fluvial discharge.

Sediment grain size provides direct information on changes of the sediment source and the sedimentary environment (Friedman and Sanders, 1978; Wu et al., 2015). The characteristics of grain size can be expressed by the grain size distribution curve, and usually the mean or median diameter is used (Xu, 1999). In this study, the value of mean grain size (Mz) showed that two major grain size boundaries occur at depths of 34 and 19 cm, separating a middle sedimentary unit (34–19 cm) that contains coarser sediment from the lower and upper sedimentary units that contain finer sediment (Figure 7d). The sand content of the upper, middle and lower layers was 11.2%, 33.6%, and 9%, respectively (Figure 7a); the silt content of these layers was 69%, 58.6%, and 76.1%, respectively (Figure 7b), and the clay content of these layers was 19.8%, 7.8%, and 14.8%, respectively (Figure 7c). On the basis of ¹³⁷Cs chronology (Figure 3), we speculate that these significant changes of grain size parameters at depths of 34 and 19 cm might represent a record of the channel shifts of the YR in 1855 AD and 1976 AD, respectively.

353 The sediment of Laizhou Bay mainly comes from the YR and other small rivers located in the southern part of 354 Laizhou Bay (Zhang et al., 2017; Gao et al., 2018). Prior to 1855 AD, when the YR entered the Yellow Sea, the sediment contribution to the BS from other small rivers was relatively larger. The fine fraction suspension sediment that was 355 356 derived from other small rivers favors the hypothesis of fine sediment accumulation in core CJ06-435 during this period. 357 When the YR reentered the BS after 1855 AD, the dispersal of YR material contributed substantially to the 358 sedimentation of the BS. It was reported that more than 80% of the YR sediment discharges into the BS during the 359 summer period (Bi et al., 2011). Owing to the barrier effect of the tidal shear front and the weak river flow, most of the 360 river-delivered sediment is deposited on the offshore delta within 15 km of the river mouth (Wang et al., 2007; Bi et al., 361 2010). Only a small part of the fine clay fraction is transported by the coastal currents over long distances and deposited 362 across or along the shore in summer (Wu et al., 2015). During the winter (October to March) season, the much stronger 363 winter monsoon generates large waves, resulting in intensive sediment resuspension in the coastal region owing to the 364 enhanced bottom shear stress (Yang et al., 2011; Bi et al., 2011). The resuspended sediment is transported southeastward 365 along the coast of Laizhou Bay by the monsoon-enhanced coastal currents passing through the location of the sediment 366 core CJ06-435. Therefore, after 1855 AD, the sediment of core CJ06-435 mainly included the fine fraction of the 367 suspended sediment dispersed from the YR mouth, the resuspended sediment from the coastal area off the YR delta in 368 the winter, and the locally resuspended sediment.

The accumulation of YR-suspended sediment during the summer season in Laizhou Bay was closely associated with the sediment dispersion pattern off the active delta lobe (Xing et al., 2016). The estuary of the YR, during most of the period 1855–1976 AD, was north of the modern YRD, and suspended sediment from the YR was transported northeastward to Bohai Bay and the central Bohai basin. The contribution of YR-suspended sediment to the sedimentation of core CJ06-435 was smaller and the resuspended sediment became a dominant material source. During 374 the winter, the large waves generated by the strong winds may result in intensive resuspension of the seabed sediment 375 and lead to part of the coarse sediment in the YR mouth and Laizhou Bay being transported to the study area, which 376 induces an evident increase in mean grain size and a decrease in the amount of fine sediment. After 1976 AD, the lower 377 channel of the YR shifted to the Oingshuigou course in Laizhou Bay; the suspended sediments derived from the YR 378 estuary were primarily driven southward and southeastward along the coast, leading an increasing transportation of most 379 of the YR-suspended sediment into Laizhou Bay (Oiao et al., 2010). As a result, the dispersal of river-laden sediment 380 contributed substantially to the sedimentation of core CJ06-435, with a fine sediment layer being formed in the upper 381 part of the core.

382 The results inferred from our grain-size data on the migrations of the YR lower channel since 1855 AD and their 383 effects on the sedimentary environments of the adjacent BS are in accordance with the results of other studies from 384 Laizhou Bay (Wu et al., 2015) and central BS (Hu et al., 2011). Based on the records of sediment core collected from 385 Laizhou Bay, Wu et al. (2015) found that when the YR mouth approached the core location, the sediment became finer; 386 otherwise, the active resuspension resulted in the accumulation of coarser sediment owing to strong hydrodynamics. The 387 grain size results from the central mud areas of the BS also point to the conclusion that the sediment supply from the YR 388 to the central BS was cut off because of the shift of the YR terminal course from the Diaokou source in outer Bohai Bay 389 to the Qingshuigou course in Laizhou Bay in 1976; hence, resuspended sediment became a primary source of sediment 390 dispersal in the central BS. As a result, there was a significant increase in the proportion of sand in surface sediment in 391 the central BS (Hu et al., 2011).

It is worth noting that the variation of grain size characteristics in the period of 6570–5000 a BP is very similar to that after 1855 AD. As shown in Figure 7d, the shift of Mz in the period of 6570–5000 a BP also began with a significantly increased of Mz at 6570 a BP (160 cm) when the YR flowed into the BS in northern Shangdong province. This similar variation of grain size in the period of 6570–5000 a BP (superlobe 1) and after 1855 AD (superlobe 10) implies that a similar YR channel shifting occurred during these two periods. However, further research is needed to reveal how the deltaic and neritic sea sedimentary environment was impacted by the river system.

399 **5.3.** Coastal salt marsh response to hydrological change

Two high-amplitude salt marsh vegetation shifts are displayed in the herb pollen record during 6570–5000 a BP (superlobe 1) and after 1855 AD (superlobe 10), indicating rapid oscillations of environmental conditions in the coastal area of BS. Within single intervals of the YR superlobe, a recurrent and directional alternation of herb pollen taxa is observed in the following order: the shift of herb pollen data began with an abrupt decrease of Cyperaceae pollen, followed by a steep increase of Chenopodiaceae and *Artemisia* pollen (Figure 8b).

405 Cyperaceae. Chenopodiaceae, and Artemisia are the three plant families/genus that contain the important 406 representatives of coastal salt marsh plants (Lu et al., 2006). In the salt marsh of the modern YRD, species composition 407 of Cyperaceae, Chenopodiaceae, and Artemisia varies with salinity and soil moisture. Plant families such as Cyperaceae 408 are mainly composed of hydrophytes and phreatophyte Eleocharis valleculosa, Cyperus rotundus, Scirpus planiculmis, 409 S. triaueter, S. vagara, S. juncoides, and Juncellus serotinus (Pan and Xu, 2011). The presence of Cyperaceae necessarily 410 indicates lower saline conditions, since hydrophytes and phreatophyte sedges typically colonize in the middle and upper 411 part of the supralitoral zone, both sides along the riverbank, the coast of the lake, and the interfluvial lowlands of the 412 paleo-river. These areas are far from the coastline, and the main type of soil is salinized soil with lower salinity (Zhang 413 et al., 2009a; Xu, 2011). Chenopodiaceae are mainly composed of halophyte Suaeda glauca, S. salsa, and Salicornia 414 europaea. Artemisia mainly consist of halophyte and xerophyte Artemisia carvifolia, A. capillaris, and A. annua (Xing 415 et al., 2003; Zhang et al., 2009b). In the modern YRD, halophytes are distributed in the intertidal zone mudflat and the 416 outside margin part of the supralittoral zone. These areas are near the coastline, characterized by a high incidence of 417 wave brakes and prolonged inundation regimes, and the main type of soil is saline (Zhang et al., 2009b). Therefore, in 418 salt marsh plant communities, the variation in the amount of Cyperaceae, Chenopodiaceae, and Artemisia is often thought 419 to reflect environmental gradients controlled by the distance from the coast, local topography, terrigenous material, and 420 freshwater input (Gonz aez and Dupont, 2009; Zhang et al., 2009a). The pollen record from the BS could provide 421 evidence of coastal salt marsh development over decades to centuries of the response to environmental alternations 422 during the period of hydrological change.

423

In the studied sequence, the YR flowed into the BS toward the northern Shandong Province after a course shift.

The lower river was initially braided upon relocation, as characterized by unchannelized river flow. At this initial stage, the river-derived sediment was largely accumulated in the floodplain and/or among the antecedent rivers owing to the lack of channelization (Wu et al., 2017), filling the coast of the lake, the interfluvial lowlands of the paleo-river, and the supralittoral zone, etc. This caused the destruction of hydrophytes and phreatophyte sedges in these areas. This process is indicated in our records by the significant decrease in the amount of Cyperaceae pollen percentages for superlobe 1 and superlobe 10 (Figure 8b).

430 Eventually, natural channel adjustments resulted in the coalescence of multiple channels into a single channel (Wu 431 et al., 2017). A large amount of river-derived sediment was deposited at the mouth of the YR, causing the progradation 432 of the YRD. Because of the strong influence of the tides, the intertidal zone in the YRD was originally bare beach. Along 433 with the seaward expansion of the newly formed beach, the influence of tides was weakening on the original bare beach 434 wetland and the salinity of the original beach wetland began to decrease (Zhang et al., 2007). Pioneer species of salt 435 marshes, e.g., Suaeda glauca, S. salsa and Salicornia europaea (Chenopodiaceae), first colonized this original bare beach 436 (Zhang et al., 2009a). The significant increase of Chenopodiaceae in our pollen record (Figure 8b) is, therefore, 437 interpreted as the development of the S. glauca population.

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439 **5.4.** Palaeovegetation reconstruction and its climate significance

440 Based on the age model, the record between 3000 a BP and 1855 AD of core CJ06-435 is somewhat confused. 441 Because the CSR was extremely low (about 0.005 cm/yr) during 3000 a BP to 1855 AD. As reported in recent studies, 442 the CSR of cores from the tidal flat and neritic sea of the south BS were in the range 0.02–0.13 cm/yr before 2000 a BP 443 (in cores H9601, H9602, ZK228, and ZK1, Figure 1; He et al., 2019), 0.04–0.06 cm/yr between 2000 a BP to 1855 AD 444 (in cores ZK228, HB-1, and GYDY, Figure 1; He et al., 2019), and 0.35–1.38 cm/yr since 1855 AD (Wu et al., 2015; 445 Oiao et al., 2017; Xu et al., 2018). Although core CJ06-435 is offshore compared to the other cores (e.g. H9601, H9602, 446 ZK228, HB-1, ZK1 and GYDY), and it should have a lower CSR compared to those cores. But, the difference of CSR 447 between CJ06-435 and those cores reach up to ten-fold. The reasonable explanation is that there might be some 448 deposition hiatus between 3000 a BP and 1855 AD of core CJ06-435. The calculated CSR in the upper layer (since 1855 AD, as calculated to 0.17–0.48 cm/yr) and the lower layer (3000–8500 a BP, as calculated to 0.016–0.057 cm/yr) of core CJ06-435 are comparable to the nearby records by He et al. (2019) and Xu et al. (2018). Therefore, we just focused on the vegetation successions and climate change between 8500 and 3000 a BP, and only gave a cautious discussion for the chronology uncertain interval in this study.

453 During the period from 8500 to 6500 a BP (palynological zone 1, 271–156 cm), the palynofloral assemblages are 454 mainly composed of the pollen of broadleaved trees, such as *Ouercus*, *Betula*, *Alnus*, and Ulmaceae, combined with the 455 pollen of hydrophytes and phreatophyte Cyperaceae and Typha; of these, the pollen of Ouercus and Typha are 456 predominant (Figures 4 and 5). In contrast, the pollen of halophytic and xerophytic herbs and conifer trees is scarce. The 457 pollen assemblages encountered herein indicated that the vegetation of the BS land area consisted mainly of oak -rich 458 temperate broadleaf deciduous forest, with some conifer trees on the uplands, and freshwater lakes and marshes 459 dominating the coastal area, under the influence of a markedly warmer and wetter climate than the present. The highest 460 values of AP pollen concentration (Figure 5), reflecting a dense vegetation cover, also represent warm conditions during 461 this period. The pollen data are comparable to that found from previous palynological studies carried out in north China 462 (e.g., Yi et al., 2003; Ren, 2007; Chen and Wang, 2012; Li et al., 2019) and northeast China (e.g., Ren and Beug, 2002; 463 Li et al., 2011; Stebich et al., 2015), from which a warm, wet climate corresponding to the Holocene Optimum was 464 inferred. Under the influence of the Holocene Optimum, the forest cover evidently increased in north and northeast China 465 (Ren, 2007). In the YR drainage area and Shandong Peninsula, the broadleaved deciduous forest thrived, accompanied 466 by the presence of monsoonal evergreen forest and the abrupt decrease in the herbaceous taxa and conifers (Yi et al., 467 2003; Chen and Wang, 2012; Li et al., 2019).

During the period from 6500–5900 a BP (palynological zone 2a, 156–128 cm), a climatic cooling period is identified by an increase of conifers, *Pinus*, combined with an abrupt reduction of broadleaved trees (*Quercus*, *Betula*, *Alnus*, *Pterocarya*, Ulmaceae, and Moraceae). Halophytic and xerophytic herbs taxa such as Compositae, *Artemisia*, and Chenopodiaceae also increase, while Cyperaceae and aquatic herbs *Typha* obviously decreased. The climate shifted from warm, wet to cool, and dry may have caused the reduction of broadleaved deciduous forests, the expansion of conifer forests, and the gradually disappearance of freshwater lakes and marshes that had spread over the coastal area of BS. 474 This result is in good agreement with previous studies. A pollen record from Shandong Peninsula revealed that *Ouercus* 475 content decreased, and herbs increased quickly following the Holocene Optimum, indicating a potential climate 476 deterioration (Chen and Wang, 2012). A pollen record from Lake Bayanchagan in southern Inner Mongolia also showed 477 that deciduous trees declined, conifers reached their maximum values, whilst steppe vegetation remained relatively high 478 during 6500–5100 a BP, indicating cold and dry climate conditions (Jiang et al., 2006). It is worth noting that, a vast 479 delta complex began to build up in the western part of the BS after 6570 a BP, which resulted in the increase of land area 480 and development of the YR delta wetland. It can be concluded that the expansion of salt marsh during this period may 481 be partly related to the formation of the YR delta complex.

482 During the period from 5900–3500 a BP (palynological zone 2b, 128–63 cm), climate cooling and drying is 483 observed by a reduction of broadleaved trees such as *Quercus*, *Pterocarya*, Ulmaceae and Moraceae, and a rising 484 frequency of halophytic and xerophytic herbs (Artemisia and Chenopodiaceae) (Figures 4 and 5). The cool, dry 485 conditions probably cause the contraction of broadleaved forest and the expansion of halophytic and xerophytic herbs, 486 which is similar to the findings of a study by Jiang et al. (2006). Based on the quantitative climatic reconstruction from 487 pollen and algal data for Lake Bayanchagan, Jiang et al. (2006) found that broadleaved trees, such as *Betula*, *Corvlus*, 488 Ostryopsis and Ulmus further declined, whereas the amount of steppe vegetation increased. The reconstruction of mean 489 annual temperature and total annual precipitation dropped to their minimum values during 5100-2600 a BP.

Between 3500 and 3000 a BP (63–56 cm), a warm climatic phase occurred, as suggested by an increase in the amount of pollen from broadleaved trees (*Quercus, Betula, Alnus, Pterocarya,* and Ulmaceae), with low frequencies of conifer pollen (*Pinus*). Moreover, halophytic and xerophytic herb pollen, including Compositae, *Artemisia*, and Chenopodiaceae, were still present at a high frequency (Figures 4 and 5). Accordingly, the assemblages reveal that a warm, dry climate probably developed during this period.

From 3000 a BP to 1855 AD (56–30 cm), as mentioned in the first paragraph of this section, the CSR was extremely low during the period from 3000 a BP to 1855 AD (56–30 cm). We suggest that there might be some deposition hiatus and only tentatively discuss this section of the pollen record. This section begins with a relatively high percentage of broadleaved trees (*Quercus, Betula, Alnus, Pterocarya,* and Ulmaceae) and low frequencies of conifer pollen (*Pinus*) 499 (56-41 cm): this is consistent with the previous stage (3500–3000 a BP, 63-56 cm). Afterward, there is a dramatic 500 decrease in the occurrence of *Ouercus*, quickly followed by a sudden increase in *Pinus* (41–30 cm, Figures 4 and 5). 501 These pollen data suggest that the climate was warm during the latter part of this period (56–41 cm, 3000 a BP–?), 502 following the climate condition of previous stage (63–56 cm, 3500–3000 a BP). In the earlier part of this period (41–30 503 cm, ?-1855 AD), the sudden increase in *Pinus* and major reductions of *Ouercus* are likely signs of human impacts on 504 the natural vegatation, including deforestation and cultivation. Park and Kim (2015) interpreted the decrease in the 505 percentage of *Ouercus* and increase of *Pinus* in the late Holocene as marking the development of secondary vegetation 506 under anthropogenic influence. Based on two boreholes palynological from the YRD, Yi et al. (2003) found a sudden 507 reduction of *Ouercus*, followed by a marked increase of *Pinus* after 1300 a BP. Their research considered that this typical 508 lag between the two taxa may indicate that, after the clearance of the local broadleaved deciduous forests, the vegetation 509 was replaced by a secondary pine forest.

After 1855 AD (palynological zone 3, 30–0 cm), a significant decline in broadleaved trees (*Quercus*) and conifers (*Pinus*) pollen, as well as an increase in the frequency of herbs (Poaceae, Compositae, *Artemisia*, and Chenopodiaceae) pollen may reflect the further strengthening of human disturbance on the vegetation and the expansion of intensive agricultural cultivation into forests of the BS coastal area. Moreover, after 1855 AD, the present YR began returning to the BS, and forming a vast area of floodplain and estuarine wetland on the southwest coast of the BS (Saito et al., 2000; Jiang et al., 2013). The variation of herb pollen may be partly related to the development of the modern YRD wetland.

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517 **5.5.** Holocene temperature variations in north China and possible driving mechanisms

518 Many previous studies of north and northeast China have used the *Quercus* pollen percentage to infer regional 519 temperature variation (Ren and Zhang, 1998; Yi et al., 2003; Li et al., 2004; Xu et al., 2014; Zhang et al., 2019). The 520 *Quercus* pollen percentage from CJ06-435 core is consistent with previous studies, which also provide a regional air 521 temperature reference. As shown in Figure 9d, the percentage of *Quercus* pollen in CJ06-435 core indicates a warm early 522 Holocene from 8500 to 6500 a BP, a cool mid-Holocene from 6500 to 3500 a BP, and then a relatively warm late 523 Holocene. These climate changes were also apparent in the change of *Quercus/Pinus* (Q/P) ratio. The average Q/P ratio was approximately 0.57 between 8500 and 6500 a BP, and changed to 0.33 between 6500 and 3500 a BP, and thengradually increased (Figure 9c).

526 The pollen-based record of temperature evolution from core CJ06-435 are broadly in-phase with published high-527 resolution sea surface temperature record from core YS01 (Figure 9e) of the Yellow Sea, suggesting at least a local 528 pattern of temperature variations during the Holocene. To investigate whether the temperature pattern was local 529 characteristics of the BS and Yellow Sea area, or whether it was rather a regional patterns of East Asia as a whole, the 530 pollen records of core CJ06-435 were compared with recently-published and relatively well-dated sequences from north 531 China, northwest China, and the Tibetan Plateau (see Figure 9a for site locations), including the sedimentary pollen-532 based temperature record from the Narenxia Peat within the Kanas Lake, Northwest China (Figure 9f; Feng et al., 2017); 533 U^K₃₇ inferred temperature record at Lake Qinghai (Figure 9g; Hou et al., 2016); and lacustrine sedimentary pollen-534 based quantitative temperature record (the mean annual temperature) from Lake Bayanchagan in Inner Mongolia in north 535 China (Figure 9h; Jiang et al., 2006). All three records indicate that the temperature was high between 8500 and 6000 a 536 BP, low between 6000 and 4000–3000 a BP, and averagely high after 4000–3000 a BP (Figure 9f-h), this is consistent 537 with the basic change pattern of our pollen-based temperature variation (Figure 9c and 9d). The comparison of the five 538 records demonstrates that the CJ06-435 core *Quercus* pollen percentage record is, at a minimum, of regional significance. 539 Insolation has been widely accepted as an important factor in Holocene climate variation. The covariation of 540 Northern Hemisphere extratropical (30 °to 90 °N) temperature and local summer insolation on an orbital scale, and the 541 long-term decrease of summer insolation make the especially pronounced cooling of the Northern Hemisphere extra-542 tropics during the Holocene (Marcott et al., 2013) appear reasonable. However, the general pattern of temperature 543 variation revealed by our study is not entirely consistent with local mean annual insolation forcing (Figure 9b). Our 544 results indicated a cold mid-Holocene from 6500 to 3500 a BP and a relatively warm late Holocene. This temperature 545 characterization of a cool mid-Holocene and a relatively warm late Holocene is also seen in many proxy records in the 546 East Asia (Thompson et al., 1997; Jiang et al., 2006; Hou et al., 2016; Wu et al., 2016; Feng et al., 2017; Jia et al., 2019). 547 The cooler mid-Holocene seen in the East Asia, could not solely be explained by the gradually decreasing summer 548 insolation during the Holocene but might be related to other forcings.

549 We compared the pollen-based temperature record of core CJ06-435 (Figure 9c and 9d) with the frequency of El 550 Ni ño events reconstructed from the Botryococcene concentration in the El Junco Lake sediment (Figure 9i; Zhang et al., 551 2014) and the ENSO variability reconstructed from δ^{18} O values of individual planktonic foraminifera retrieved from 552 deep-sea sediments (Figure 9; Koutavas and Joanides, 2012). As shown in Figure 9, lower temperature periods in the 553 mid-Holocene tend to occur during a period of low El Niño activity, a relatively high temperature period in late Holocene 554 tend to occur during a period of high El Niño activity, which indicates that there may be some link between the 555 temperature of BS and Yellow Sea area and the ENSO system. Modern research suggests that ENSO can influence the 556 evolution of temperature behavior, over interannual to multi-decadal time ranges (Hoerling et al., 2008; Triacca et al., 557 2014). In East Asia, many studies have indicated that the East Asian winter monsoon and the ENSO are tightly coupled 558 (Zhou et al., 2007; Cheung et al., 2012; An et al., 2017). Generally, the strength of winter monsoon and East Asia troughs 559 weakens in an El Niño year, and the weakening could cause the observed winter half-year warming (Xu et al., 2005). On 560 centennial/millennial time scales, using pollen data from Lake Moon in the central part of the Great Khingan Mountain 561 Range, Wu et al. (2019) recently connected increased El Ni ño frequency with the decrease of winter monsoon activity 562 in the East Asia, and the warming winter temperature in the Great Khingan Mountain Range since the mid-Holocene. 563 Feng et al. (2017) founded that warm-phased ENSO was teleconnected with weakening of the Siberian High, and that 564 the weakening was a cause of the observed winter half-year warming in southern Siberia. Likewise, the results of this 565 study, indicating more or less synchronicity of the climate change in north China and ENSO activity, provide a possible 566 linkage between the climate of north China and oceanic forcing during the mid-late Holocene.

In addition, radiative forcing by greenhouse gases (GHGs) rose 0.5 W/m² during the mid-late Holocene (Marcott et al., 2013), which would be expected to yield 1°C warming at Kinderlinskaya Cave in the southern Ural Mountains from 7000 a BP to the pre-industrial (Baker et al., 2017). Recently, both winter insolation and GHG forcing have been proposed as the major driving factors for winter warming during the Holocene in the Siberian Arctic (Meyer et al., 2015) and the southern Ural Mountains (Baker et al., 2017). Similarly, summer warming in central Asia during the mid-late Holocene, recorded by the alpine peat α -cellulose δ^{13} C record from the Altai Mountains (Rao et al., 2019), has been proposed to be mainly driven by the enhanced GHG forcing and increasing human activities. Rao et al. (2020) suggested that GHG forcing was the dominant driver of the summer and winter warming trends since ~5000 a BP. The effects of
GHG forcing are global. Hence, we suggest that enhanced GHG forcing may be an important driver for mid-late
Holocene temperature variations of East Asia.

In summary, the temperature characterizations of a cool mid-Holocene and a relatively warm late-Holocene revealed by the East Asia records could be linked with the change of insolation, ENSO activity and GHG forcing. The cooler mid-Holocene may be related to a combination of the decreasing summer insolation, weak El Niño activity, and relatively lower GHG radiative forcing during this interval. Along with strengthened ENSO activity and enhanced GHG forcing in the late Holocene, there was increased temperature.

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583 **6.** Conclusions

584 Through the palynological and grain size reconstruction of coastal area vegetation and environment in core CJ06-585 435, we were able to identify specific responses of plant communities to climatic (temperature, precipitation), 586 hydrological, and anthropogenic impacts. Our data elucidate the pattern and mechanisms driving coastal salt marsh 587 succession at decade-to-century timescales. Two intervals of expanded salt marsh vegetation correspond to the formation 588 of YR delta superlobes, indicating that soil development and salinity gradients are the main factors determining the 589 vegetation dynamics of coastal wetland. Our pollen-based temperature index revealed a warm early Holocene (8500-590 6500 a BP), a subsequent cool stage between 6500 and 3500 a BP, and a slightly warming episode after 3500 a BP. The 591 reliability of the record, especially the cooler mid-Holocene, is further supported by several other temperature records 592 from East Asia. We suggest that changes in insolation, ENSO activity and GHG forcing could have played an important 593 role in the temperature evolution at the East Asia.

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596 Code/Data availability

597 The co-authors declare that all data included in this study are available upon request by contact with the 598 corresponding author (Email: *jinxiachen@fio.org.cn*).

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601 Author contribution

| 602 | Chen Jinxia wrote the manuscript; Shi Xuefa and Liu Yanguang revised the manuscript; Qiao Shuqing provided |
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| 603 | many constructive suggestions for the manuscript; Yang Shixiong provided the pollen data of surface sediment; Li |
| 604 | Jianyong use pollen data of core CJ06-435 as base for quantitative climate reconstruction; Yan Shijuan, Lv Huahua, Li |
| 605 | Xiaoyan and Li Chaoxin provided financial support for the collection of samples and obtained samples. |
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| 608 | Competing interests |
| 609 | The authors declare that they have no conflict of interest. |
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| 612 | Acknowledgements |
| 613 | We thank the crew of the R/V Kan407 for sampling. We also thank Dr. Nan Qingyun for improving this paper. This |
| 614 | work was supported by the National Natural Science Foundation of China (Grant Nos. 41420104005 and 41576054). |
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878 Table captions

Table 1

881 AMS radiocarbon dates from core CJ06-435 and one tie points corresponding to the deepest onset of ¹³⁷Cs in environmental

samples at measurable levels; for calibration in years before present (a BP) 0 = 1950 AD.

| Core depth | Materials | Radiocarbon date | Age error | Calibrated age (1o) | Mean calibrated | Laboratory |
|------------|----------------------------|------------------|-----------|---------------------|-----------------|------------|
| (cm) | | (a) | (a) | (a BP) | age (a BP) | |
| 25 | ¹³⁷ Cs | - | - | - | -4 | NIGLAS |
| 7 | Mixed benthic foraminifera | 3020 | 30 | 2854-3039 | 2951 | Beta |
| 13 | Mixed benthic foraminifera | 2990 | 30 | 2817-2997 | 2913 | Beta |
| 17 | Mixed benthic foraminifera | 3060 | 30 | 2908-3102 | 3003 | WHOI |
| 59 | Mixed benthic foraminifera | 3340 | 30 | 3270-3485 | 3359 | Beta |
| 69 | Mixed benthic foraminifera | 3590 | 25 | 3563-3725 | 3656 | WHOI |
| 87 | Mixed benthic foraminifera | 4450 | 30 | 4695-4878 | 4801 | Beta |
| 119 | Mixed benthic foraminifera | 5200 | 30 | 5604-5770 | 5706 | WHOI |
| 129 | Mixed benthic foraminifera | 4520 | 30 | 4812-4965 | 4894 | Beta |
| 161 | Mixed benthic foraminifera | 6020 | 30 | 6501-6667 | 6592 | WHOI |
| 183 | Mixed benthic foraminifera | 6340 | 35 | 6886-7081 | 6981 | WHOI_ |

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| 897 | Figure caption |
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| 899 900 | Figure 1 : Geographic map of the Bohai Sea, with locations of core CJ06-435 (red circle) and other sites referred to in this study (purple circles). Cores references: H9601 and H9602 (Saito et al., 2000), ZK1 (Li et al., 2013), ZK228 (Xue et al., 2014), ZK28 (Xue et |
| 901 | al.,1988), HB-1 (Liu et al., 2009a), GYDY (Liu et al., 2014). |
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| 904 | Figure 2: Vegetation map around the Bohai Sea and the ocean current in the Bohai Sea during the summer (a) and |
| 905 | winter (b) (YSWC: Yellow Sea Warm Current: SBSCC: Southern Bohai Sea Coastal Current: LNCC: Liaonan Coastal Current: |
| 906 | LBCC: Lubei Coastal Current, modified from Qiao et al., 2010; the vegetation dataset was provided by the Environmental |
| 907 | and Ecological Science Data Center for West China, National Natural Science Foundation of China |
| 908 | [http://westdc.westgis.ac.cn] and is based on the Vegetation Atlas of China, 1:1000,000; Hou, 2001). |
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| 910 911 | Figure 3: Lithology, grain size, color reflectance L^* and a^* , magnetic susceptibility, and activity profiles for ¹³⁷ Cs and ²¹⁰ Ph of core CI06-435 |
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| 914 | Figure 4: Percentage diagram of the principal pollen taxa from core CJ06-435. Pollen zonation is based on CONISS |
| 915 | results. |
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| 918 | Figure 5: Concentration diagram of the principal pollen taxa from core CJ06-435. |
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| 920 | |
| 921 | Figure 6: Spatial distribution of modern pollen percentage (black solid circle, %) and concentration (red open circle, |
| 922 | grains/g) in Laizhou Bay, Bohai Sea (modified from Yang et al., 2016). |
| 923 | |
| 924 | |
| 925 | Figure 7: (a-f) Vertical profiles of grain-size parameters and halophytic and xerophytic herb (Chenopodiaceae and |
| 926 | Artemisia) pollen percentage and concentration of core CJ06-435 (Mz – mean grain size). (g) The location of Yellow |
| 927 | River superlobe 1 (Lijin superlobe) and superlobe 10 (Modern superlobe) (modified after Xue, 1993). |
| 928 | |
| 929 | |
| 930 | Figure 8: (a) Correlating proxy to paleo-superlobe variation of the YR, from top to bottom: percentage of Cyperaceae, |
| 931 | Chenopodiaceae, and Artemisia pollen; concentration of Chenopodiaceae, Artemisia, and Cyperaceae pollen; sand |
| 932 | percentage. (b) Detailed pollen and grain size profiles representing salt marsh species (Cyperaceae, Chenopodiaceae, |
| 933 | <i>Artemisia</i>) relative abundances and hydrodynamic change during the formation of Yellow River superlobe 1 and 10. 34 |

Pollen percentage of Cyperaceae, Chenopodiaceae and *Artemisia* from core CJ06-435 indicating the directional alternation of salt marshes along the Bohai Sea, (1)— Unchannelized river flow characterized the onset of Yellow River channel shift, caused a large amount of river–derived sediment accumulation in the floodplain and destroyed the sedges in the coastal depression; (2)—Along with the formation of a new channel, lateral migration of the lower channel stopped, and new intertidal mudflat was formed. Pioneer species (Chenopodiaceae, *Artemisia*) first colonize bare zones of lower and middle marsh areas.

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942 Figure 9: Comparison of relevant Holocene temperature records with solar irradiance and ENSO proxy records derived 943 from the equatorial Pacific. (a) Locations of the sites where the Holocene temperature records are derived. The 944 schematic large-scale diagram modified from Hao et al. (2017). In the diagram, the purple area is Tibetan Plateau, the 945 vellow area is Chinese Loess Plateau, and numbered dots refer to the corresponding study sites. (b) Summer (mean of 946 June) insolation irradiance for the Northern Hemisphere (40°N). (c,d) Quercus/Pinus (Q/P) rate and Quercus pollen 947 percentage records from core CJ06-435, bold-blue and bold-red lines are the five-point running average. As introduced 948 in part 5.4 "Palaeovegetation reconstruction and its climate significance", there might be some deposition hiatus 949 between 3000 cal. a BP and 1855 AD. So, Q/P rate and Quercus pollen percentage records during 3000 cal. a BP and 950 1855 AD is for reference only. (e) U^{K'}₃₇ SST record from YS01 core in the south Yellow Sea (Jia et al., 2019). (f) Pollen-951 based mean annual temperature (MAT) record from Narenxia Peat in the southern Altai (Feng et al., 2017). (g) U^K₃₇ 952 inferred temperature record at Lake Qinghai (Hou et al., 2016). (h) Pollen-based mean annual temperature record from 953 Lake Bayanchagan in Inner Mongolia, North China (Jiang et al., 2006). (i) Botryococcene concentrations in the El Junco 954 sediment, a proxy for frequency of El Niño events (Zhang et al., 2014). (j) Variance of δ^{18} O values of individual planktonic 955 foraminifera (G. ruber) in sediment core V21–30 from the Galápagos region, a proxy for ENSO variance (Koutavas and 956 Joanides, 2012).

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Figure 5

Figure 6



Figure 7





