Holocene hydroclimate reconstruction based on pollen, XRF, and grain-size analyses and its implications for past societies of the Korean Peninsula

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Abstract
The dynamics of the East Asian Summer Monsoon (EASM) and their link to past societies during the Holocene are topics of growing interest. In this study, we present results of pollen, geochemistry, and grain-size analyses from the STP18-03 core sampled from Miryang in the Korean Peninsula, which spans ca. 8.3–2.3 ka BP. In-phase relationships of these proxies revealed an imprint of the Holocene Climate Optimum (HCO) during the early to mid-Holocene and subsequent drying toward the late-Holocene in accordance with decreasing solar insolation. At centennial timescales, our study indicates drier climate during ca. 7.5–7.1, 6.4–6.0, and 4.8–3.6 ka BP. Notably, our finding for ca. 6.4–6.0 ka BP contributes further evidence of a drying event in the Korean Peninsula during this period. We suggest that the Pacific Ocean played a role in the underlying mechanism of hydroclimate change in the region. A strong Kuroshio Current (KC) and long-term El Niño–Southern Oscillation (ENSO)-like variability in the Western Tropical Pacific (WTP) were closely linked to the influence of the EASM over the Korean Peninsula. In particular, dry phases during ca. 4.8–3.6 and 2.8–2.3 ka BP, which were synchronous with a more active ENSO, closely corresponded to lower population levels indicated by a summed probability distribution (SPD) of archaeological records previously assembled in the Korean Peninsula. This finding implies that past human societies of Korea were highly vulnerable to climate deterioration caused by precipitation deficits.

Keywords
EASM, grain size, Holocene, Korea, past societies, pollen, XRF

Received 15 December 2020; revised manuscript accepted 3 May 2021

Introduction
A topic of rising interest in academia is a relationship between climate change and human societies. This issue has been addressed by many studies worldwide over the past few decades in regions including Europe (Büntgen et al., 2011; Tallavaara et al., 2015), Greenland (D’Andrea et al., 2011), America (Munoz et al., 2010), Mesopotamia (Carolin et al., 2019; Weiss et al., 1993), India (Dixit et al., 2018), Egypt (Manning et al., 2017), and China (An et al., 2005; Wang et al., 2014; Xie et al., 2013) and regarding issues of population changes and migration (D’Andrea et al., 2011; Tallavaara et al., 2015), rise and fall of civilizations (Weiss et al., 1993), societal unrest (Manning et al., 2017), etc. Most of these studies have supported a significant role of climate in the vicissitudes of past societies, whereas others have recently reported human resilience to environmental stresses (Blockley et al., 2018; Flohr et al., 2016). Thus, the relationship between climate and human societies cannot be always simplistically determined and requires case-specific investigations. Even at a sub-regional scale of East Asia, Northwest, North, and South China, the Qinghai Tibetan Plateau, Japan, and Korea all show apparently different human activity patterns during the Holocene (Crema et al., 2016; Oh et al., 2017; Wang et al., 2014). Therefore, to accurately analyze a socio-environmental relationship in a particular area, a direct comparison of continuous, local, and high-resolution paleoclimate proxy data with human activity indicators is crucial. However, in the Korean Peninsula, this work has remained at a nascent stage (Constantine et al., 2019; Park et al., 2019). The topic of climate–society relationships in Korea has mostly depended on historical periods where individual records of natural disasters exist (e.g. Yoon and Hwang, 2009, 2010), leading to assumptive storytelling about their potential impact on contemporary societal unrest (Cho, 2009) as well as prohibiting a
continuous and direct comparison of amplitudes in societal and climatic fluctuations. Although Ahn and Hwang (2015) tried to cover this issue in a prehistoric period, their study was confined to central-western area of the peninsula and not successful in linking climate and societal response in a quantitative manner, due to a lack of reliable high-resolution climate data.

A key to investigating a relationship between climate change and prehistoric societies in the Korean Peninsula is using continuous and high-resolution data both in archaeology and paleoclimatology suitable for a direct comparison. Recently, a proxy for prehistoric population changes in the Korean Peninsula was presented (Oh et al., 2017), opening up a possibility for application to climate-society investigations (Park et al., 2019). As for paleoclimatic records, the East Asian Summer Monsoon (EASM) has received a substantial attention as a major driver of the East Asian Holocene climate and mostly been reconstructed from mainland China as its key region (An, 2000; Chen et al., 2015b; Dykoski et al., 2005; Ji et al., 2005; Stebich et al., 2015; Wang et al., 2005). However, the temporally and spatially complex nature of the EASM (An, 2000; Chen et al., 2015a; Zhou et al., 2016) requires additional studies from many different locations within the region to better reconstruct its site-specific impact during the Holocene. For example, different imprints of the EASM within the region are being increasingly addressed between the interior and coastal parts of China (Chen et al., 2015a; Zhou et al., 2016; Zhu et al., 2017) as well as between China and Korea (Park et al., 2017). However, a scarcity of high-resolution EASM reconstructions in the Korean Peninsula covering a continuously long period throughout the Holocene hinders a direct comparison of the EASM with archaeological proxies. Moreover, there are issues to be unraveled regarding the EASM history of the Korean Peninsula, such as occurrence of cooling events around 7.5–7.1 and 6.4–6.0 ka BP, which have been reported in mainland China (Chen et al., 2015b; Feng et al., 2020; Wang et al., 2005) but not yet been clearly identified in the Korean Peninsula. Therefore, there is a significant need to produce a high-resolution EASM reconstruction by combining multiple types of proxies with sufficient time spans, ranging from the early- to the late-Holocene.

In this study, we performed a multi-proxy-based analysis of Holocene hydroclimate change in the Miryang area of the Korean Peninsula from ca. 8.3 to 2.3 ka BP. We combined pollen, X-ray fluorescence (XRF), and grain-size data to reconstruct the EASM history over the Korean Peninsula. The high temporal resolution of our XRF analysis allowed us to analyze the paleoclimate data set at decadal to annual scales. We identified millennial-scale oscillations in EASM intensity over the Korean Peninsula and examined the role of the Pacific Ocean as an underlying mechanism, particularly in terms of variance in the El Niño–Southern Oscillation (ENSO) and heat and moisture supply along the Kuroshio Current (KC). We also identified a synchronous change of EASM with archaeological proxies with sufficient time spans, ranging from the early- to the late-Holocene.

### Regional setting

Our coring site (35°26′18.84″ N, 128°46′41.26″ E; 4.64 m a.s.l.) is located at a former backswamp area of a floodplain of the Miryang River in the southeastern part of the Korean Peninsula (Figure 1a–c), which is now reclaimed as farmland. The Miryang River is a tributary of the Nakdong River, which eventually flows out to the Korea Strait between the Korean Peninsula and the Japanese Archipelago. The backdorp of the coring site is a group of small mountains several hundred meters in elevation, including Jongnamsan (663 m) and Palbongsan (391 m) (Figure 1c). The Korean Peninsula is situated amid the East Asian monsoonal system, and experiences seasonal differences in air pressure due to its location between the Eurasia continent and the Pacific Ocean (An, 2000) (Figure 1a). In summer, warm and humid southeasternly wind flows into the Korean Peninsula under the influence of the warm KC, whereas in winter, cold, and dry northerly wind from the Siberian High dominates. According to a 30-year (1981–2010 AD) climate record from a meteorological observation station 6 km north of the coring site, the mean annual temperature is 13.3°C, with a monthly-averaged January temperature of 0.0°C and 25.8°C in August. The mean annual precipitation is 1229.4 mm; precipitation is lowest in December (16.4 mm) and highest in July (269.5 mm) (Figure 1d). A total precipitation of 829.5 mm during June–September, JJAS accounts for 67.5% of the total annual precipitation, indicating significant influence of the EASM on the hydroclimate of the region (Korea Meteorological Administration, 2020).

The regional vegetation of Miryang is composed of broadleaf and mixed forest. On Mt Jongnamsan, arboreal species such as *Pinus densiflora*, *Quercus serrata*, *Rhus sylvestris*, *Q. acutissima*, *Castanea crenata*, *Indigofera kirilowii*, and *Q. varia*-bilis and herb species such as *Opiliumsundundulatifolius*, *Carex lanceolata*, *Artemisia keiskeana*, *Spodiopogon sibiricus*, *Coccus trilobus*, and *Arundinella hirta* dominate at 30–298 m altitude (Korea National Institute of Ecology, 2016). In wetlands along the Miryang River, *Phragmites communis*, *Salix gracilistyla*, *Phragmites japonica*, *Persicaria longiseta*, *Persicaria thunbergii*, *Juncus effusus* var. *decipiens*, and *Zizania latifolia* are also found (Korea National Institute of Environmental Research, 2002).

### Materials and methods

#### Core materials and dating

In April 2018, the 20-m STP18-03 core was collected in 1-m sections from a former backswamp area on a floodplain of the Miryang River, using a hydraulic piston corer (Figure 1 and Supplemental Figure S1). The corer was mounted on a truck, which was anchored on solid ground during the drilling process. Depth zones of the uppermost 0–1.25 m and lowermost 14–20 m were excluded from all analyses as the former were regarded to have been disturbed by agriculture and the latter consisted mainly of gravels. We sent a total of 16 samples between 3.5 and 12.8 m to the Korea Institute of Geoscience and Mineral Resources (KIGAM), Republic of Korea for age dating (Table 1). Among these, eight samples were measured using the optically stimulated luminescence (OSL) dating method. These samples were treated with Na₂P₂O₇, HCl, H₂O₂, and H₂SiF₆ to extract quartz with a diameter of 4–11 μm. Then, OSL signals were measured using a TL-DA-20 reader (Riso DTU, Roskilde, Denmark) equipped with a blue light-emitting diode (LED; 470 ± 20 nm) stimulation source. Plant and wood fragments from the other eight samples were used for radiocarbon dating by accelerator mass spectrometry (AMS). Compiling the OSL and radiocarbon dating results, we constructed an age model using the *bacoR* package ver. 2.3 (Blaauw and Christen, 2011). The package allows a combination of different types of dates in a single age-depth modeling. Here, the radiocarbon dates were calibrated based on the IntCal13 calibration dataset (Reimer et al., 2013), while the OSL dates were not because they were already set on the calendar scale. All resulting ages applied in our analysis were expressed as calendar ages.

#### Proxy analyses

We performed palynological analysis of a total of 137 samples at intervals ranging from 1 to 70 cm (from 1 to 6 cm for 395–1010 cm and from 9 to 70 cm for the remaining sections). The average temporal resolution was 41.4 years, with minimum and maximum values of 12 and 348 years, respectively. For sample preparation, we followed the standard protocol of Faegri et al. (1989)
including HCl (10%), KOH (10%), HF (40%), and acetolysis treatments. KOH treatment was repeated twice to remove organic matter completely. For highly humic samples in the range of 501–528 cm, the KOH procedure was repeated up to three times. For each sample, one Lycopodium tablet containing 177,745 spores was added, and at least 300 pollen grains and spores were counted on each slide using a Leica microscope at 400× magnification. For palynomorph identification, we referred to a pollen atlas from Lake Suigetsu, Japan (Demske et al., 2013). Pollen and spore percentages were calculated for each taxon relative to the total sum of non-aquatic pollen grains and spores in the sample. The result was visualized using the Tilia software ver. 2.0.41 (Grimm, 1991), and stratigraphically constrained cluster analysis was conducted using CONISS (Grimm, 1987) based on non-aquatic taxa.

Grain-size analysis was performed using a Mastersizer 2000 laser diffraction particle-size analyzer (Malvern Instruments, UK) at KIGAM. Approximately 300 mg of each sample was collected at 10 cm intervals from the 205 to 1390 cm section. These subsamples were treated with H$_2$O$_2$ (35%) and HCl (1 N) to remove organic matter and carbonates. Grain sizes of <4 μm, 4–63 μm, and >63 μm were classified as clay, silt, and sand, respectively.

XRF analysis was also conducted at KIGAM using an XRF core scanner (Avaatech B.V.; Alkmaar, Netherlands), which extracts elemental concentration data in a nearly continuous manner (Croudace et al., 2006; Löwemark et al., 2011). XRF signals were measured from split core surfaces from depths of 12.5–1293.5 cm, with settings of 10/50 kV and 0.25/1.0 mA, and a sampling time of 30 s. In total, 2088 values were collected at a resolution of 0.5 cm. However, we did not perform paleoenvironmental interpretations on data at depths above 365 cm, because this section corresponded to periods later than ca. 2.3 ka BP, for which there is evidence of agricultural practices in the Miryang region (Yoon et al., 2005).

Cross-spectral analysis on the proxy data was conducted using the REDFIT-X software ver. 1.1 (Ólafsdóttir et al., 2016) with 1000 Monte Carlo simulations (nsim = 1000), an oversampling value of 4.0 for Lomb–Scargle Fourier transform (ofac = 4.0), four segments with 50% overlap (nseg = 4), and a Welch window (iwin = 1). To avoid statistical bias, we did not include two pollen values at depths of 380 and 395 cm, which reflected an abrupt shift immediately before cessation of pollen deposition. This resulted in an average temporal resolution of 38.7 years per pollen sample between 401 and 1280 cm depth. For all the other proxies, we used values within the timespan of the pollen data to match the temporal range of the cross-spectral analysis. An average temporal resolution of XRF Ti values was 6.8 years per sample.

**Figure 1.** (a) Locations of the STP18-03 core (yellow star, this study) and other proxies mentioned in this article (white squares): A7 from Okinawa Trough (Sun et al., 2005), MD98-2176 from the Arafura Sea (Stott et al., 2004), Lake Xiaolongwan (Chu et al., 2014; Xu et al., 2014, 2019), Lake Sihailongwan (Seibicht et al., 2015), Daihai Lake (Xiao et al., 2004), Gonghai Lake (Chen et al., 2015b), Qinghai Lake (Ji et al., 2005), and Dongge Cave (Dykoski et al., 2005; Wang et al., 2005). Red arrows indicate the trajectory of the Kuroshio Current. Blue arrows indicate the trajectories of the East Asian Summer Monsoon (EASM) and Indian Summer Monsoon (ISM). (b) Map of the Korean Peninsula, showing the locations of GY-1 (Park et al., 2019), Pomaeho Lake (Constantine et al., 2019), and Jeju Island (Park, 2017). The maps in (a) and (b) were produced using the GMRT tool (Ryan et al., 2009). (c) Regional satellite map of our coring site (yellow star), showing the locations of the Jongnamsan (663 m) and Palbongsan (391 m) mountains and the Miryang River. This map was adapted from the © Google Earth Pro software ver. 7.3.3.7673 (https://earth.google.com/). (d) Mean monthly temperature and precipitation in Miryang during 1981–2010 (Korea Meteorological Administration, 2020). For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.
Table 1. Optically stimulated luminescence (OSL) and radiocarbon dating results for the STP18-03 core. Calibrated radiocarbon dates were expressed as weighted mean ages calculated in the R bacon package (Blauw and Christen, 2011) ver. 2.3 using the IntCal13 dataset (Reimer et al., 2013).

| OSL sample depth (m) | Dose rate (Gy/ka) | Water content (%) | Equivalent dose (Gy) | OSL age (yr BP) |
|----------------------|------------------|------------------|---------------------|-----------------|
| 3.50–3.55            | 2.94 ± 0.20      | 28.3             | 5.09 ± 0.20         | 1700 ± 130      |
| 4.50–4.55            | 3.23 ± 0.20      | 37.4             | 14.09 ± 0.19        | 4400 ± 270      |
| 5.50–5.55            | 3.55 ± 0.20      | 26.8             | 18.75 ± 0.40        | 5300 ± 320      |
| 6.50–6.55            | 3.21 ± 0.18      | 33.3             | 22.14 ± 0.58        | 6900 ± 430      |
| 8.50–8.55            | 3.21 ± 0.20      | 39.1             | 23.88 ± 0.20        | 7400 ± 460      |
| 9.50–9.55            | 3.08 ± 0.17      | 40               | 24.32 ± 0.45        | 7900 ± 470      |
| 10.50–10.55          | 3.39 ± 0.19      | 30.6             | 29.28 ± 0.48        | 8600 ± 500      |
| 12.50–12.55          | 3.02 ± 0.19      | 40               | 25.25 ± 0.99        | 8300 ± 620      |

| AMS sample depth (m) | Material dated | Lab code | δ13C (%) | Age (14C yr BP) | Calibration age (cal yr BP) |
|----------------------|----------------|----------|----------|-----------------|---------------------------|
| 5.8                  | Wood fragments | KGM-ITg190725 | −28.8    | 5110 ± 40     | 5835                      |
| 6.94                 | Wood fragments | KGM-ITg190726 | −27.7    | 5710 ± 40     | 6538                      |
| 7.95                 | Plant fragments (stem) | KGM-ITg190727 | −32.5    | 5570 ± 50     | n/a                       |
| 8.32                 | Wood fragments | KGM-ITg190728 | −32.0    | 6310 ± 40     | 7261                      |
| 9.77                 | Wood fragments | KGM-ITg190729 | −33.9    | 7210 ± 40     | 7907                      |
| 10.32                | Wood fragments | KGM-ITg190730 | −30.8    | 6500 ± 50     | n/a                       |
| 11.81                | Wood fragments | KGM-ITg190731 | −30.8    | 7280 ± 40     | 8182                      |
| 12.8                 | Wood fragments | KGM-ITg190732 | −27.1    | 7330 ± 50     | 8343                      |

*Excluded from the age model.

Results

Lithology and chronology

Lithology of the STP18-03 core consists of very dark brown mud between 1240 and 1390 cm, muddy silt between 790 and 1240 cm, brown sand between 700 and 790 cm, muddy sand between 450 and 700 cm, mud between 400 and 450 cm, and reddish-brown clayey silt between 210 and 400 cm (Figure 2a). Chronology of the core contains a record of ca. 6420 years, from the mid- to late-Holocene, representing the period from ca. 8340 cal year BP (1280 cm depth) to ca. 1920 cal year BP (350 cm depth) (Figure 2b). Throughout the age-depth model, our dating results exhibit high stratigraphic consistency despite two different dating methods used, namely OSL and radiocarbon dating (Figure 2b). Only two radiocarbon samples at 795 and 1032 cm show deviation from other 14 dating results, which we attribute to potential disturbance by plant root penetration or carbon contamination during sample preparation processes and thus excluded from the chronology. For proxy interpretation, we only used the section between 350 and 1280 cm where dating results exist. The sedimentation rate is 0.38 cm per year for the 790–1280 cm segment and 0.09 cm per year for the 380–790 cm segment (average, 0.14 cm per year). Sedimentation is continuous throughout the core section used in this study, without noticeable trace of interruption on the core facies (Supplemental Figure S1). Stratigraphic consistency in our dating results (Figure 2b) also support that sedimentation was continuous and disturbance by hiatus did not occur, and if it did, the effect was minimal. Sediment gaps are partially observed at 901–905 and 1100–1140 cm, perhaps due to on-site technical problems while coring or disturbance by underground water.

Multi-proxy environmental data

Zone 1 (790–1390 cm, before 7040 cal year BP). This zone consists mainly of very dark brown muddy silt and sand alternating in multiple layers with ~15% of clay (Figure 2a, c–e, and Supplemental Figure S1). Sand percentages fluctuate significantly between 0.5% and 79.1% (Figure 2e). The amount of titanium (Ti) also varies, with a large amplitude generally in the opposite direction to that of sand content (Figure 2f). The tree pollen percentage is generally low at 1145–1280 cm, with a minimum value of 77.2% (Figure 2g). Artemisia (mugwort), Poaceae (wild grass), and fern species comprise 20.6% of the pollen assemblage (Figure 2h); this trend is reversed at 790–1090 cm, where the proportion of arboreal pollen remains high and stable at 87.1%–96.4%, and that of Artemisia and Poaceae pollen together with fern spores remains low, at an average of 8.7%. The largest proportion of arboreal pollen is constituted by Quercus (oak) (Figure 3); other broadleaf tree genera include Alnus (alder), Fraxinus (ash), and Ulmus (elm), which reach the highest proportion in this zone of the core. Notably, at a depth of 950 cm, a sudden increase in Ulmus pollen to 58.0% (Figure 3) coincides with abrupt shifts in sand and Ti content (Figure 2e and f). The abnormal values imply a potential sudden local disturbance event, which possibly perturbed the pre-existing vegetation and gave rise to pioneer species with high environmental tolerance (McVein, 1953; Weng et al., 2004).

Zone 2 (400–790 cm, 3130–7040 cal year BP). From 790 to 400 cm, the clay and silt contents gradually increase as depth decreases, from ~15% to ~20% and ~30% to ~70%, respectively (Figure 2c and d). Sand percentages and Ti content stabilize, changing in tandem with the pollen data (Figure 2e–h). Overall, lower sand content is concurrent with a lower proportion of tree pollen and higher Ti values, and vice versa. Between 590 and 635 cm, where the sand percentage temporarily decreases to 28.8%, the Ti content increases sharply as the proportion of tree pollen and higher Ti values, and vice versa. Between 590 and 635 cm, where the sand percentage temporarily decreases to 28.8%, the Ti content increases sharply as the proportion of tree pollen and higher Ti values, and vice versa. Between 790 to 500 cm, where the proportions of sand and arboreal pollen drop to below 10% and 80%, respectively. Upland herbs such as Artemisia and Poaceae occupy this relatively open area, whereas Pteridium ferns rise to a maximum value in this zone. This trend is in contrast with the 500–590 and 635–790 cm sections, in which the proportions of sand and tree pollen remain high whereas Ti contents are low. Notably, a high proportion of tree pollen at 500–590 cm is reflected by an increase in Quercus pollen (Figure 3).

Zone 3 (210–400 cm, after 3130 cal year BP). This part of the core is characterized by reddish-brown clayey silt (Figure 2a, c–e, and Supplemental Figure S1) with low sand content (Figure 2e)
and high Ti values (Figure 2f). Pollen deposition is interrupted after an explosive increase in upland herbs (mainly Artemisia and Poaceae), Cyperaceae (sedges), and Polypodiales undiff. ferns at the beginning of this zone (Figures 2g and h, and 3). Considering the near absence of sand content in this zone and coincident change in the sediment color (Figure 2a and e), this disruption may have been caused by a cessation of water supply from the river to the site, possibly due to climate drying, dwindling of the river, and subsequent exposure of the site to air.

**Discussion**

**Role of our proxy data as paleoclimate indicators**

At millennial timescales, our Ti, pollen, and sand content data are broadly consistent with a declining trend of summer solar insolation in the Northern Hemisphere from the mid- to late-Holocene (Berger and Loutre, 1991) (Figure 4a–e). The gradual decrease in the proportion of arboreal pollen (Figure 4c) reflects cooling and drying associated with a southward migration of the Intertropical Convergence Zone (ITCZ) induced by orbital forcing (Berger and Loutre, 1991; Haug et al., 2001). Our sand proportion data also follow this trend, as fluvial sand discharge by the Miryang River would have weakened due to less precipitation in the late-Holocene relative to earlier periods (Figure 4e). Ti contents change in the opposite direction to these two proxies, generally increasing toward the late-Holocene (Figure 4b). In many studies, Ti has been used as an indicator of terrestrial erosion, although its paleoenvironmental interpretation may vary according to regional context (Bakke et al., 2009; Sun et al., 2008). In the present study, considering its high coherency with arboreal pollen and sand proportion data (Figure 4b–e), we interpret Ti as reflecting hydroclimate change in the study area. During wet periods, more tree growth (mainly oak and pine trees, Figure 3) in nearby mountains such as Jongnamsan and Palbongsan (Figure 1c) would have suppressed soil erosion via the anchoring effect of roots, leading to lower Ti contents, while enhanced river discharge likely increased...
sand transportation to the nearby backswamp. On the other hand, as climate became drier toward the late-Holocene, tree replacement by herbs and ferns would have weakened this effect, resulting in greater Ti erosion in contrast to lower sand deposition at the site due to reduced river discharge.

This attribute of Ti data as a climate proxy in relation to vegetation change is further supported by cross-spectral analysis (Ólafsdóttir et al., 2016) (Figure 5). The analysis of Ti and arboreal pollen data implies high coherency at frequencies of 518, 148, 127, and 104 years (Figure 5a). A ~500-year frequency is widely detected in East Asian monsoonal regions and has been attributed to solar activity which is modulated by oceanic influences such as the thermohaline circulation (THC) or the ENSO (Li et al., 2021; Stebich et al., 2015; Stuiver and Braziunas, 1993; Xu et al., 2014, 2019). This link is supported by additional analyses of our Ti and pollen data with the Western Tropical Pacific (WTP) SST record from the MD98-2176 core (Stott et al., 2004) and hematite-stained grains (HSG) in the North Atlantic (Bond et al., 2001) (Figure 5b, c, e, and f). Smaller frequencies of ~150, ~130, and ~100 years are likely attributable to a solar origin (Roth and Reijmer, 2005; Scuderi, 1993); these signals have been interpreted in terms of solar modulation on the EASM strength at Qinghai Lake, central China (Ji et al., 2005) and Jeju Island (Park, 2017), south of the Korean Peninsula (Figure 1a and b). Cross-spectral analysis between total solar irradiation (TSI) (Steinhilber et al., 2009) and our data provides further evidence of a solar contribution to these cycles (Figure 5d and g), although ~130- and ~100-year frequencies between tree proportion and TSI should be treated with caution as they fail to reach statistically significant levels (Figure 5g). This was possibly due to the lower temporal resolution of pollen analysis (38.7 years per sample) relative to XRF scanning (6.8 years per sample).

**Climate change in the Korean Peninsula during the Holocene**

From ca. 8.3 to 5.4 ka BP, highly sustained proportions of arboreal pollen (Figure 4c) likely reflect the influence of the Holocene Climatic Optimum (HCO) (An et al., 2000; Wanner et al., 2008; Zhou et al., 2016), which resulted in warmer summers in the Korean Peninsula. This possibly increased annual average temperatures by 1°C–2°C in the peninsula compared with the pre-industrial period (Renssen et al., 2012). The climate was generally warm and humid, influenced by the northward advance of the EASM (Yang et al., 2015). However, one exception is an abrupt drop in the proportion of tree pollen at ca. 8.2 ka BP, when pollen from herb taxa including mugwort (*Artemisia*) and wild grass (*Poaceae*) suddenly increased (Figure 3). This may reflect the “8.2 ka event,” an abrupt global cooling phenomenon (Alley et al., 1997; Cheng et al., 2009; Veski et al., 2004), which correspond to previous reports of this event in the Korean Peninsula (Park et al., 2018, 2019).

At centennial timescales, several periods of wet and dry climate alternate throughout the mid- to late-Holocene. Our Ti, pollen, and sand data indicate relatively wet climates during ca. 8.3–7.5, 7.1–6.4, 6.0–4.8, and 3.6–2.8 ka BP, and drier conditions during ca. 7.5–7.1, 6.4–6.0, and 4.8–3.6 ka BP (Figure 4b–e). A pronounced feature of this multi-centennial-scale environmental change is a close connection with SST records from the Pacific Ocean (Figure 4f and g) (Stott et al., 2004; Sun et al., 2005). Periods of warm and wet climate are accompanied by high SST records obtained from the Okinawa Trough, which is an indicator of KC intensity (Figure 1a and 4f) (Sun et al., 2005). This agrees with the point that the KC affects the climate of the coastal East Asian region by transporting abundant heat and moisture from the WTP as part of the Pacific western boundary current system (Hu et al., 2015; Jian et al., 2000; Sasaki et al., 2012). In summer, warm seawater of the KC induces energetic evaporation and wind convergence along its route, which result in enhanced ascent motions in the troposphere and increased precipitation over the nearby regions (Sasaki et al., 2012). Located on the main route of the KC, climate of the Korean Peninsula has also been likely influenced by the same mechanism as indicated by a close connection between the WTP SSTs and EASM intensity on the peninsula (Lee et al., 2020; Lim and Fujioki, 2011; Park et al., 2016). Our present study indicates that greater heat and water vapor supply from the WTP along the KC would have enabled active atmospheric convection and stronger EASM influence over the Korean Peninsula during ca. 8.3–7.5, 7.1–6.4, 6.0–4.8, and 3.6–2.8 ka BP, whereas the opposite would have occurred during ca. 7.5–7.1, 6.4–6.0, and 4.8–3.6 ka BP.

Among these periods, a sign of drying and/or cooling around 6.4–6.0 ka BP (Figure 4b–e) at Miryang is consistent with a previous finding at Lake Pomnae in the central Korean Peninsula (Constantine et al., 2019) (Figure 1b). Outside of the peninsula,
Daihai Lake (Xiao et al., 2004) and Gonghai Lake (Chen et al., 2015b) in North China and Dongge Cave in South China (Wang et al., 2005) (Figure 1a) also record abrupt shifts toward less precipitation at ca. 6.4–6.0 and 7.5–7.1 ka BP. These findings altogether suggest a possibility that the climate events were widespread phenomena in the East Asian region. Nevertheless, this possibility should be carefully addressed, as some study sites such as Lake Xiaolongwan (Chu et al., 2014; Xu et al., 2019) and Lake Sihailongwan (Stebich et al., 2015) (Figure 1a) do not clearly exhibit a drying/cooling signal. Regarding this inconsistency, a couple of possibilities can be considered. One possible factor is an issue of temporal resolution. In the case of Dongge Cave, the high-resolution DA stalagmite (Wang et al., 2005) detects a drying signal while the D4 stalagmite (Dykoski et al., 2005), with a lower resolution, does not. It is not reasonable to assume differences in actual climate conditions because they were collected from the same cave. Similarly, in the Korean Peninsula, a previous study at Gwangyang (Figure 1b, GY-1) does not exhibit a climate shift at ca. 6.4–6.0 ka BP (Park et al., 2019) in contrast to Miryang (this study). As Gwangyang is located only ~100 km west to Miryang, it is unlikely that climate conditions were considerably different between those two study sites. Instead, temporal resolution could be a more convincing explanation as the sampling intervals within this period are large in GY-1 (~80 years) relative to our present study (~20–30 years).

Besides the resolution issue, potential bias inherent in proxy-based climate reconstructions should be also noted. In pollen records, source area and/or overestimation effects inherent in palynological methodology (Seppä and Bennett, 2003) might affect the interpretation of pure climate signals. For example, in this study, we suspect that thermal optimum during the early to mid-Holocene (Wanner et al., 2008) might have rendered a smaller amplitude of the vegetation response during ca. 7.5–7.1 ka BP, whereas the other sedimentary proxies, XRF, and sand percentage data exhibit clearer phase shifts with the Pacific Ocean (Figure 4b–g), possibly reflecting the “7.2 ka event” reported in central China (Feng et al., 2020). Similarly, in pollen records from Daihai Lake (Xiao et al., 2004) and Gonghai Lake (Chen et al., 2015b), drying signals during ca. 7.5–7.1 ka BP are less evident than ca. 6.4–6.0 ka BP. In this context, it cannot be ruled out that such climate shifts are not manifest in some records simply due to methodological problems. Furthermore, as for the cases of Lake Xiaolongwan (Chu et al., 2014; Xu et al., 2019) and Lake Sihailongwan (Stebich et al., 2015) in Northeast China (Figure 1a), regionally varying climate imprints caused by high-latitude forcing such as sea ice in the Sea of Okhotsk should also be considered (Stebich et al., 2015), given that their sampling resolutions were high enough to detect centennial-scale climate oscillations if present. Overall, in order to elaborate understanding of potential climate deterioration events at ca. 6.4–6.0 and 7.5–7.1 ka BP, further high-resolution data are required from multiple locations in East Asia. At least in this study, our finding at Miryang suggests evidence that such climate shifts were likely present in the Korean Peninsula during these two periods.

Our proxy data also indicate a pronounced drying trend in the Korean Peninsula during ca. 4.8–3.6 ka BP (Figure 4b–e), which is consistent with global findings of significantly decreased precipitation and/or temperature around this period (Bond et al., 1997, 2001; Wang et al., 2005; Wanner et al., 2011). Cooling and drying in this period at Miryang likely reduced tree coverage (Figure 4c), which in turn resulted in increased soil erosion from the nearby mountains (Figure 4b). Located in the coastal part of East Asia, this climate impact might have been amplified by long-term ENSO-like variance, which strengthened from the mid-Holocene (Conroy et al., 2008; Donders et al., 2008; Moy et al., 2002). The ENSO exerts a dampening effect on EASM intensity by affecting low- to mid-latitude atmospheric patterns (Hu et al., 2015; Xu et al., 2019), evidence of which has been found in Northeast China (Li et al., 2021) and the Korean Peninsula (Park et al., 2016, 2017) particularly during El Niño-like phases. It has been found that lower-than-usual SSTs in the western tropical Pacific during El Niño induce a westward enhancement of a subtropical ridge, hindering a northward advance of the EASM rain belt and consequently shortening a rainy season in Northeast Asia.

Figure 5. Coherency spectra for (a) titanium – tree percentage, (b) Titanium – MD76 (Stott et al., 2004), (c) Titanium – hematite-stained grains (HSG) (Bond et al., 2001), (d) Titanium – total solar irradiation (TSI) (Steinhilber et al., 2009), (e) Tree – MD76, (f) Tree – HSG, (g) Tree – TSI. Theoretical, mean Monte Carlo (for alpha = 0.050), and 90% Monte Carlo false alarm levels are indicated by black, orange, and red lines, respectively. All cross-spectral analyses were conducted using the REDFIT-X software ver. 1.1 (Ölafsdóttir et al., 2016). For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.
Response of past societies to mid- to late-Holocene climate change

The summed probability distribution (SPD) method is increasingly used as a proxy for past population levels (Ahn and Hwang, 2015; Bevan et al., 2017; Crema et al., 2016; Oh et al., 2017; Talavaara et al., 2015; Wang et al., 2014; Xu et al., 2019), by assembling radiocarbon age calculations from archaeological findings (Gamble et al., 2005; Surovell et al., 2009). In this study, the Ti value of the STP18-03 core closely followed the SPD consisting of a total of 3127 radiocarbon dates collected from archaeological pit-house samples in the Korean Peninsula (Oh et al., 2017) (Figure 6a and b). Ti content gradually decreased and remained low until ca. 4.8 ka BP, recovered high values by ca. 3.6 ka BP, and then diminished again before increasing at ca. 2.8 ka BP (Figure 6b). This trend is coherent with, but opposite to, changes in the SPD data, which indicate larger populations during ca. 6.0–4.8 ka BP and ca. 3.6–2.8 ka BP, and lower populations during ca. 4.8–3.6 ka BP and after ca. 2.8 ka BP (Figure 6a). Three abrupt transition points around 4.8, 4.2, and 4.0 ka BP are found in both datasets, validating their link (Figure 6a and b). Together with our previous research (Constantine et al., 2019; Park et al., 2019), this robust synchronicity between Ti data and archaeological records contributes to increasing evidence that past societies of the Korean Peninsula sensitively responded to climate change.

Given its synchronicity with arboreal pollen data, we used the Ti contents as a proxy of climate change and identified two periods with a relatively warm and wet climate, when past populations increased: ca. 6.0–4.8 ka BP and ca. 3.6–2.8 ka BP (Figure 6). This link is accompanied by high SSTs in the Okinawa Trough (Figure 4f). Notably, the former period corresponds to the latter part of the HCO, characterized by a warm and wet climate during the mid-Holocene both globally (Renssen et al., 2009, 2012; Wanner et al., 2008) and in the Korean Peninsula (Park et al., 2019). During this period, favorable climate conditions would have enabled successful hunting and gathering, with sufficiently abundant food resources to sustain population growth (Wang et al., 2014). Besides hunting and gathering, evidence of foxtail, broomcorn, and legume cultivation has been found at Korean Middle-Late Chulmun (Neolithic) sites for as early as ca. 5.5 ka BP (Lee, 2011), which would have also benefited from warm and humid climate conditions (Wang et al., 2014; Xu et al., 2019). Moreover, the latter period of ca. 3.6–2.8 ka BP was a time of explosive increase in Bronze Age settlements in the Korean Peninsula, beginning as early as ca. 3.9 ka BP (Kim and Bae, 2010). Similarly, Lee (2011) suggested a major transition from Chulmun to Multum (Bronze) culture in the Korean Peninsula at ca. 3.4 ka BP, with clear evidence of intensive agriculture, including domesticated plants such as rice. The impact of temporary climate deterioration around 3.2 ka BP (Figure 6b–d), which may reflect enhanced ENSO activity (Moy et al., 2002) (Figure 6e) and/or the 3.2-ka event (Kaniewski et al., 2017), was not large enough in the Korean Peninsula to interrupt the increasing population trend during this cultural boom (Figure 6a).

During ca. 4.8–3.6 ka BP, high Ti values and a low proportion of arboreal pollen indicate a cool and dry climate in the Korean Peninsula when human activity declined (Figure 6). Evidence of significant cooling and drying events around this period (Bond et al., 1997, 2001; Wang et al., 2005; Wanner et al., 2011) and their potential impacts on the shrinkage or unrest of past societies (DeMenocal, 2001) have been widely reported from various sites worldwide including Mesopotamia (Carolyn et al., 2019; Weiss et al., 1993), India (Dixit et al., 2014, 2018), China (An et al., 2005; Li et al., 2018; Xiao et al., 2018; Xu et al., 2019), Japan (Kajita et al., 2020; Kawahata et al., 2009), and Korea (Constantine et al., 2019; Park et al., 2019). In these periods, climate drying would have increased dietary stress by hindering successful hunting, gathering, millet cultivation and even livestock domestication (Kawahata et al., 2009; Roffet-Salque et al., 2018). Abrupt and synchronous changes in Ti and pollen data together with the decline in archaeological SPD values at ca. 4.8, 4.2, and 4.0 ka BP imply a significant impact of sudden climate deterioration on Korean prehistoric societies (Figure 6a–d). These changes were likely triggered by the onset of an active ENSO at ca. 4.8 ka BP (Figure 6e) and modulated by lower SSTs in the Okinawa Trough and WTP until ca. 3.6 ka BP (Figure 4f and g).

In addition, our Ti XRF data suggest synchronicity of climate deterioration with a decline in population in the Korean Peninsula from ca. 2.8–2.3 ka BP (Figure 6a–d). During this period, SSTs in the Okinawa Trough and WTP decreased (Figure 4f and g), and
stretched ENSO activity around 2.7 ka BP (Figure 6e) may have amplified the climate impact. Although this interpretation should be carefully addressed as our core sediment of this section might reflect a very early stage of human disturbance in Miryang even before the previously suggested timing of ca. 2.3 ka BP (Yoon et al., 2005), synchronous variations among Ti contents (Figure 6b, this study), the SPD (Figure 6a) (Oh et al., 2017), and pollen data of our previous study at Gwangyang (Figure 1b and 6d) (Park et al., 2019) provide support for this socio-environmental linkage. In the Korean Peninsula, most Bronze Age pottery disappears during this period, as observed for Misa-ri, Garak-dong, and Heunam-ri-type pottery around 2.9–2.8 ka BP and Yeoksam-dong, Songguk-ri, and Geomdan-ri-type pottery during 2.8–2.4 ka BP (Ahn and Hwang, 2015; Lee, 2017). In this sense, it is likely that decreasing SPD values after ca. 2.8 ka BP (Figure 6a) primarily reflect a collapse of Bronze Age culture in Korea, and that this change was influenced by climate drying and/or cooling at that time (Figure 6b–d). As during ca. 4.8–3.6 ka BP, the population decline after ca. 2.8 ka BP was a widespread phenomenon that has been detected in Korea and at many sites worldwide including mainland China (Wang et al., 2014), Turkey (Woodbridge et al., 2019), and Britain and Ireland (Bevan et al., 2017). Therefore, climate impact on human societies during this period should be understood within a global context, possibly in connection with a cooling trend after the Bond event 2 (Bond et al., 1997; Wanner et al., 2011).

Conclusion

In this study, multi-proxy datasets of pollen, XRF, and grain-size were used to reconstruct the EASM history of the Korean Peninsula from ca. 8.3 to 2.3 ka BP. We identified a synchronous variation between hydroclimate and past human activity. The Holocene climate in Korea was sensitively modulated by the strength of the KC and ENSO-like variance. Wet conditions prevailed during ca. 8.3–7.5, 7.1–6.4, 6.0–4.8, and 3.6–2.8 ka BP, when SSTs in the western Pacific Ocean were sufficiently high to enhance EASM strength. However, during ca. 7.5–7.1, 6.4–6.0, and 4.8–3.6 ka BP, climate became drier due to KC weakening and an increase in ENSO-like variance that likely dampened the EASM. Although regional imprints of climate deterioration during ca. 6.4–6.0 ka BP remain unclear in East Asia, the findings of our study contribute to our knowledge of this climate event in the Korean Peninsula. The reconstructed hydroclimate change was also synchronous with past population levels indicated by the SPD of archaeological remains. Past societies flourished amid favorable climate conditions during ca. 6.0–4.8 ka BP and ca. 3.6–2.8 ka BP, but suffered from precipitation deficits during ca. 4.8–3.6 ka BP. This finding is consistent with multiple findings of contemporaneous collapses of past civilizations worldwide and confirms the global socio-environmental linkage in the Korean Peninsula. Nevertheless, it should be noted that the relationship between climate and past societies is not always straightforward, and further research is still needed to elaborate our understanding of its dynamics.

Acknowledgements

We thank the editor and anonymous reviewers for their useful comments and suggestions for improving the manuscript.

Author contributions

Jinheum Park: Conceptualization, Formal analysis, Investigation, Writing – Original Draft, Visualization. Jungjae Park: Conceptualization, Methodology, Writing – Review and Editing, Supervision, Project administration, Funding acquisition. Sangheon Yi: Methodology, Resources, Project administration. Jaesoo Lim: Investigation. Jin Cheul Kim: Investigation. Qiuohong Jin: Investigation. Jieun Choi: Investigation.

Funding

The author(s) disclosed receipt of the following financial support for the research, authorship, and/or publication of this article: This work was funded by the Basic Research Project (GP2017-013) of the Korea Institute of Geoscience and Mineral Resources (KIGAM), the National Research Foundation of Korea (NRF-2018R1D1A1A09083072), and the Hanmaum Peace and Research Foundation (HPRF).

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Data availability

Data used in this study are available online at the open-access repository Pangaea: https://doi.pangaea.de/10.1594/PANGAEA.922338.

Supplemental material

Supplemental material for this article is available online.

References

Ahn SM and Hwang JH (2015) Temporal fluctuation of human occupation during the 7th–3rd millennium cal BP in the Central-Western Korean Peninsula. Quaternary International 384: 28–36.
Alley RB, Mayewski PA, Sowers T et al. (1997) Holocene climatic instability: A prominent, widespread event 8200 yr ago. Geology 25(6): 483–486.
An Z (2000) The history and variability of the East Asian paleomonsoon climate. Quaternary Science Reviews 19(1–5): 171–187.
An Z, Porter SC, Kutzbach JE et al. (2000) Asynchronous Holocene optimum of the East Asian monsoon. Quaternary Science Reviews 19(8): 743–762.
An CB, Tang L, Barton L et al. (2005) Climate change and cultural response around 4000 cal yr BP in the western part of Chinese loess plateau. Quaternary Research 63(3): 347–352.
Bakke J, Lie O, Hegaard E et al. (2009) Rapid oceanic and atmospheric changes during the Younger Dryas cold period. Nature Geoscience 2(3): 202–205.
Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. Quaternary Science Reviews 10(4): 297–317.
Bevan A, Colledge S, Fuller D et al. (2017) Holocene fluctuations in human population demonstrate repeated links to food production and climate. Proceedings of the National Academy of Sciences 114(49): E10524–E10531.
Blauw M and Christen JA (2011) Flexible paleoclimatic age-depth models using an autoregressive gamma process. Bayesian Analysis 6(3): 457–474.
Blockley S, Candy I, Matthews I et al. (2018) The resilience of postglacial hunter-gatherers to abrupt climate change. Nature Ecology & Evolution 2(5): 810–818.
Bond G, Kromer B, Beer J et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. Science 294(5549): 2130–2136.
Bond G, Showers W, Cheseby M et al. (1997) A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. Science 278(5341): 1257–1266.
Büntgen U, Tegel W, Nicolussi K et al. (2011) 2500 years of European climate variability and human susceptibility. Science 331(6017): 578–582.

Carolin SA, Walker RT, Day CC et al. (2019) Precise timing of abrupt increase in dust activity in the Middle East coincident with 4.2 ka social change. Proceedings of the National Academy of Sciences 116(1): 67–72.

Chang C, Zhang Y and Li T (2000) Interannual and interdecadal variations of the East Asian summer monsoon and tropical Pacific SSTs. Part I: Roles of the subtropical ridge. Journal of Climate 13(24): 4310–4325.

Chen J, Chen F, Feng S et al. (2015a) Hydroclimatic changes in China and surroundings during the medieval climate anomaly and little ice age: Spatial patterns and possible mechanisms. Quaternary Science Reviews 107: 98–111.

Chen F, Xu Q, Chen J et al. (2015b) East Asian summer monsoon precipitation variability since the last deglaciation. Scientific Reports 5: 11186.

Cheng H, Fleitmann D, Edwards RL et al. (2009) Timing and structure of the 8.2 kyr BP event inferred from δ18O records of stalagmites from China, Oman, and Brazil. Geology 37(11): 1007–1010.

Cho N (2009) Climate and drought: a new perspective on history. History and Discourse 53: 607–619 (in Korean).

Chu G, Sun Q, Xie M et al. (2014) Holocene cyclic climatic variations and the role of the Pacific Ocean as recorded in varved sediments from northeastern China. Quaternary Science Reviews 102: 85–95.

Conroy JL, Overpeck JT, Cole JE et al. (2008) Holocene changes in eastern tropical Pacific climate inferred from a Galápagos lake sediment record. Quaternary Science Reviews 27(11–12): 1166–1180.

Constantine M, Kim M and Park J (2019) Mid-to late-Holocene cooling events in the Korean Peninsula and their possible impact on ancient societies. Quaternary Research 92(1): 98–108.

Crema ER, Habu J, Kobayashi K et al. (2016) Summed probability distribution of 14C dates suggests regional divergences in the population dynamics of the Jomon period in eastern Japan. PLoS One 11(4): e0154809.

Croudace IW, Rindby A and Rothwell RG (2006) ITRAX: Description and evaluation of a new multi-function X-ray core scanner. Geological Society, London, Special Publications 267(1): 51–63.

D’Andrea WJ, Huang Y, Fritz SC et al. (2011) Abrupt Holocene climate change as an important factor for human migration in West Greenland. Proceedings of the National Academy of Sciences 108(24): 9765–9769.

DeMenocal PB (2001) Cultural responses to climate change during the late-holocene. Science 292(5517): 667–673.

Demske D, Tarasov PE and Nakagawa T (2013) Atlas of pollen, spores and further non-pollen palynomorphs recorded in the glacial-interglacial late Quaternary sediments of Lake Suigetsu, Central Japan. Quaternary International 290: 164–238.

Dixit Y, Hodell DA, Giesche A et al. (2018) Intensified summer monsoon and the urbanization of Indus civilization in Northwest India. Scientific Reports 8(1): 1–8.

Dixit Y, Hodell DA and Petrie CA (2014) Abrupt weakening of the summer monsoon in Northwest India ~ 4100 yr ago. Geology 42(4): 339–342.

Donders TH, Wagner-Cremer F and Visscher H (2008) Integration of proxy data and model scenarios for the mid-Holocene onset of modern ENSO variability. Quaternary Science Reviews 27(5–6): 571–579.

Dykoski CA, Edwards RL, Cheng H et al. (2005) A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China. Earth and Planetary Science Letters 233(1–2): 71–86.

Faegri K, Kaland PE and Krzywinski K (1989) Textbook of Pollen Analysis. New York, NY: John Wiley & Sons Ltd.

Feng X, Yang Y, Cheng H et al. (2020) The 7.2 ka climate event: Evidence from high-resolution stable isotopes and trace element records of stalagmite in Shuiming Cave, Chongqing, China. The Holocene 30(1): 145–154.

Flohr P, Fleitmann D, Matthews R et al. (2016) Evidence of resilience to past climate change in Southwest Asia: Early farming communities and the 9.2 and 8.2 kA events. Quaternary Science Reviews 136: 23–39.

Gamble C, Davies W, Pettitt P et al. (2005) The archaeological and genetic foundations of the European population during the Late Glacial: Implications for ‘agricultural thinking’. Cambridge Archaeological Journal 15(2): 193–223.

Grimm EC (1987) CONISS: a FORTRAN 77 program for stratiographically constrained cluster analysis by the method of incremental sum of squares. Computers & Geosciences 13(1): 13–35.

Grimm E (1991) Tilia and Tiliagraph. Springfield, IL: Illinois State Museum.

Haug GH, Hughen KA, Sigman DM et al. (2001) Southward migration of the intertropical convergence zone through the Holocene. Science 293(5533): 1304–1308.

Hu D, Wu L, Cai W et al. (2015) Pacific western boundary currents and their roles in climate. Nature 522(7556): 299–308.

Ji J, Shen J, Balsam W et al. (2005) Asian monsoon oscillations in the northeastern Qinghai–Tibet Plateau since the late glacial as interpreted from visible reflectance of Qinghai Lake sediments. Earth and Planetary Science Letters 233(1–2): 61–70.

Jian Z, Wang P, Saito Y et al. (2000) Holocene variability of the Kuroshio current in the Okinawa Trough, Northwestern Pacific Ocean. Earth and Planetary Science Letters 184(1): 305–319.

Kajita H, Harada N, Yokoyama Y et al. (2020) High time-resolution alkenone paleotemperature variations in Tokyo Bay during the Meghalayan: Implications for cold climates and social unrest in Japan. Quaternary Science Reviews 230: 106160.

Kaniewski D, Van Campo E and Weiss H (2017) 3.2 Ka BP mega-drought and the late bronze age collapse. In: Weiss H (ed.) Megadrought and Collapse: From Early Agriculture to Angkor. New York, NY: Oxford University Press, pp.161–182.

Kawahata H, Yamamoto H, Ohkushi K et al. (2009) Changes of drought and the late bronze age collapse. In: Weiss H (ed.) Megadrought and Collapse: From Early Agriculture to Angkor. New York, NY: Oxford University Press, pp.161–182.

Kawahata H, Yamamoto H, Ohkushi K et al. (2009) Changes of environments and human activity at the Sannai-Maruyama ruins in Japan during the mid-Holocene hypsithermal climatic interval. Quaternary Science Reviews 28(9–10): 964–974.

Kim JC and Bae CJ (2010) Radiocarbon dates documenting the neolithic–bronze age transition in Korea. Radiocarbon 52(2): 483–492.

Korea Meteorological Administration (2020) Average climate data Available at: https://data.kma.go.kr/climate/average30Years/selectAverage30YearsKoreaList.do?pgmNo=188 (in Korean) (accessed 25 March 2020).

Korea National Institute of Ecology (2016) The 4th Natural Environment Survey 2015: Vegetation: Gyeongnam 1. Sejong: Ministry of Environment of the Republic of Korea (in Korean).

Korea National Institute of Environmental Research (2002) National Survey on Natural Environments of Inland Wetlands 2001. Gwacheon: Ministry of Environment of the Republic of Korea (in Korean).

Lee GA (2011) The transition from foraging to farming in prehistoric Korea. Current Anthropology 52(S4): S307–S329.

Lee C (2017) Correlation of origins and chronology in Korean proto-historic archaeology. Journal of the Korean Archaeological Society 102: 250–273 (in Korean).
Lee E, Yi S, Lim J et al. (2020) Multi-proxy records of Holocene hydroclimatic and environmental changes on the southern coast of South Korea. Palaeogeography, Palaeoclimatology, Palaeoecology 545: 109642.

Li CH, Li YX, Zheng YF et al. (2018) A high-resolution pollen record from East China reveals large climate variability near the Northgrippian-Meghalayan boundary (around 4200 years ago) exerted societal influence. Palaeogeography, Palaeoclimatology, Palaeoecology 512: 156–165.

Li N, Sharifi A, Chambers FM et al. (2021) Linking Holocene East Asian monsoon variability to solar forcing and ENSO activity: Multi-proxy evidence from a peatland in Northeastern China. The Holocene. Epub ahead of print 18 February 2021. DOI: 10.1177/0959683621994662.

Lim J and Fujiki T (2011) Vegetation and climate variability in East Asia driven by low-latitude oceanic forcing during the middle to late-Holocene. Quaternary Science Reviews 30(19–20): 2487–2497.

Löwemark L, Chen HF, Yang TN et al. (2011) Normalizing XRF-scanner data: A cautionary note on the interpretation of high-resolution records from organic-rich lakes. Journal of Asian Earth Sciences 40(6): 1250–1256.

Lu F, Ma C, Zhu C et al. (2019) Variability of East Asian summer monsoon precipitation during the Holocene and possible forcing mechanisms. Climate Dynamics 52(1–2): 969–989.

McVean D (1953) Alnus glutinosa (L.) Gaertn. Journal of Ecology 41(2): 447–466.

Manning JG, Ludlow F, Stine AR et al. (2017) Volcanic suppression of Nile summer flooding triggers revolt and constrains interstate conflict in ancient Egypt. Nature Communications 8(1): 1–9.

Moy CM, Seltzer GO, Rodbell DT et al. (2002) Variability of El Niño/southern oscillation activity at millennial timescales during the Holocene epoch. Nature 420(6912): 162–165.

Munoz SE, Gajewski K and Peros MC (2010) Synchronous environmental and cultural change in the prehistory of the Northeastern United States. Proceedings of the National Academy of Sciences 107(51): 22008–22013.

Oh Y, Conte M, Kang S et al. (2017) Population fluctuation and the adoption of food production in prehistoric Korea: Using radiocarbon dates as a proxy for population change. Radiocarbon 59(6): 1761–1770.

Ölafsdóttir KB, Schulz M and Mudelsee M (2016) REDFIT-X: Cross-spectral analysis of unevenly spaced paleoclimate time series. Computers & Geosciences 91: 11–18.

Park J (2017) Solar and tropical ocean forcing of late-Holocene climate change in coastal East Asia. Palaeogeography, Palaeoclimatology, Palaeoecology 469: 74–83.

Park J, Han J, Jin Q et al. (2017) The link between ENSO-like forcing and hydroclimate variability of coastal East Asia during the last millennium. Scientific Reports 7(1): 1–12.

Park J, Park J, Yi S et al. (2018) The 8.2 ka cooling event in coastal East Asia: High-resolution pollen evidence from southwestern Korea. Scientific Reports 8(1): 1–9.

Park J, Park J, Yi S et al. (2019) Abrupt Holocene climate shifts in coastal East Asia, including the 8.2 ka, 4.2 ka, and 2.8 ka BP events, and societal responses on the Korean peninsula. Scientific Reports 9(1): 1–16.

Park J, Shin YH and Byrne R (2016) Late-Holocene vegetation and climate change in Jeju Island, Korea and its implications for ENSO influences. Quaternary Science Reviews 153: 40–50.

Reimer PJ, Bard E, Bayliss A et al. (2013) IntCal13 and Marine13 radiocarbon age calibration curves 0–50,000 years cal BP. Radiocarbon 55(4): 1869–1887.

Renssen H, Seppä H, Crosta X et al. (2012) Global characterization of the Holocene thermal maximum. Quaternary Science Reviews 48: 7–19.

Renssen H, Seppä H, Heiri O et al. (2009) The spatial and temporal complexity of the Holocene thermal maximum. Nature Geoscience 2(6): 411–414.

Roffet-Salque M, Marciniak A, Valdes PJ et al. (2018) Evidence for the impact of the 8.2-kyBP climate event on near Eastern early farmers. Proceedings of the National Academy of Sciences 115(35): 8705–8709.

Roth S and Reijmer JJ (2005) Holocene millennial to centennial carbonate cyclicity recorded in slope sediments of the Great Bahama Bank and its climatic implications. Sedimentology 52(1): 161–181.

Ryan WB, Carbotte SM, Coplan JO et al. (2009) Global multi-resolution topography synthesis. Geochemistry, Geophysics, Geosystems 10(3): 1525–2027.

Sasaki YN, Minobe S, Asai T et al. (2012) Influence of the Kuroshio in the East China Sea on the early summer (batsu) rain. Journal of Climate 25(19): 6627–6645.

Scudder LA (1993) A 2000-year tree ring record of annual temperatures in the Sierra Nevada mountains. Science 259(5100): 1433–1436.

Seppä H and Bennett KD (2003) Quaternary pollen analysis: recent progress in palaeoecology and palaeoeclimatology. Progress in Physical Geography 27(4): 548–579.

Steibich M, Rehfeld K, Schluezt F et al. (2015) Holocene vegetation and climate dynamics of NE China based on the pollen record from Shihailongwan Maar Lake. Quaternary Science Reviews 124: 275–289.

Steinhilber F, Beer J and Fröhlich C (2009) Total solar irradiance during the Holocene. Geophysical Research Letters 36(19): L19704.

Stott L, Cannariato K, Thunell R et al. (2004) Decline of surface temperature and salinity in the western tropical Pacific Ocean in the Holocene epoch. Nature 431(7004): 56–59.

Stuiver M and Braziunas TF (1993) Sun, ocean, climate and atmospheric 14CO2: An evaluation of causal and spectral relationships. The Holocene 3(4): 289–305.

Sun Y, Oppo DW, Xiang R et al. (2005) Last deglaciation in the Okinawa Trough: Subtropical northwest Pacific link to Northern Hemisphere and tropical climate. Paleoceanography 20(4): PA4005.

Sun Y, Wu F, Clemens SC et al. (2008) Processes controlling the geochemical composition of the South China Sea sediments during the last climatic cycle. Chemical Geology 257(3–4): 240–246.

Surovell TA, Finley JB, Smith GM et al. (2009) Correcting temporal frequency distributions for taphonomic bias. Journal of Archaeological Science 36(8): 1715–1724.

Tallavaara M, Luoto M, Korhonen N et al. (2015) Human population dynamics in Europe over the last glacial maximum. Proceedings of the National Academy of Sciences 112(27): 8232–8237.

Timmermann A, An SI, Kug JS et al. (2018) El Niño–southern oscillation complexity. Nature 559(7715): 535–545.

Veski S, Seppä H and Ojala AE (2004) Cold event at 8200 yr BP events, and societal responses on the Korean peninsula. The Holocene 14(3): 411–414.

Wang C, Lu H, Zhang J et al. (2014) Prehistoric demographic fluctuations in China inferred from radiocarbon data and their linkage with climate change over the past 50,000 years. Quaternary Science Reviews 98: 45–59.
Wanner H, Beer J, Bütkofer J et al. (2008) Mid-to late-Holocene climate change: An overview. Quaternary Science Reviews 27(19–20): 1791–1828.

Wanner H, Solomina O, Grosjean M et al. (2011) Structure and origin of Holocene cold events. Quaternary Science Reviews 30(21–22): 3109–3123.

Weiss H, Court MA, Wetterstrom W et al. (1993) The genesis and collapse of third millennium north Mesopotamian civilization. Science 261(5124): 995–1004.

Weng C, Bush MB and Chepstow-Lusty AJ (2004) Holocene changes of Andean alder (Alnus acuminata) in highland Ecuador and Peru. Journal of Quaternary Science 19(7): 685–691.

Woodbridge J, Roberts CN, Palmisano A et al. (2019) Pollen-inferred regional vegetation patterns and demographic change in Southern Anatolia through the Holocene. The Holocene 29(5): 728–741.

Xiao J, Xu Q, Nakamura T et al. (2004) Holocene vegetation variation in the Daihai Lake region of North-Central China: A direct indication of the Asian monsoon climatic history. Quaternary Science Reviews 23(14–15): 1669–1679.

Xiao J, Zhang S, Fan J et al. (2018) The 4.2 ka BP event: Multi-proxy records from a closed lake in the northern margin of the East Asian summer monsoon. Climate of the Past 14(10): 1417–1425.

Xie S, Evershed RP, Huang X et al. (2013) Concordant monsoon-driven postglacial hydrological changes in peat and stalagmite records and their impacts on prehistoric cultures in central China. Geology 41(8): 827–830.

Xu D, Lu H, Chu G et al. (2014) 500-year climate cycles stacking of recent centennial warming documented in an East Asian pollen record. Scientific Reports 4: 3611.

Xu D, Lu H, Chu G et al. (2019) Synchronous 500-year oscillations of monsoon climate and human activity in Northeast Asia. Nature Communications 10(1): 1–10.

Yang S, Ding Z, Li Y et al. (2015) Warming-induced northward migration of the East Asian monsoon rain belt from the last glacial maximum to the mid-Holocene. Proceedings of the National Academy of Sciences 112(43): 13178–13183.

Yoon SO and Hwang S (2009) The natural hazards and drought periodicity in Korea during the ancient times based on Samguk'saki. Journal of the Korean Geographical Society 44(4): 497–509 (in Korean).

Yoon SO and Hwang S (2010) The natural hazards and drought periodicity during the medieval times in Korea based on the history of Goryeo (Goryeosa). Journal of the Korean Geographical Society 17(4): 85–98 (in Korean).

Yoon SO, Kim H, Hwang S et al. (2005) Environmental change and agricultural activities during the late-Holocene in Geumcheon-ri, Milyang City, Korea. Journal of the Korean Archaeological Society 56: 27–48 (in Korean).

Zhou X, Sun L, Zhan T et al. (2016) Time-transgressive onset of the Holocene optimum in the East Asian monsoon region. Earth and Planetary Science Letters 456: 39–46.

Zhu Z, Feinberg JM, Xie S et al. (2017) Holocene ENSO-related cyclic storms recorded by magnetic minerals in speleothems of central China. Proceedings of the National Academy of Sciences 114(5): 852–857.