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**Tephrostratigraphy and paleo-environmental implications of Late Quaternary sediments and interstitial water in the western Ulleung Basin, East/Japan Sea**

**Abstract** Two piston cores, collected from the western Ulleung Basin of the East/Japan Sea, were used to investigate the Late Quaternary tephrostratigraphy, lithology, and mineralogy of the sediments, as well as the elemental composition of both sediments and interstitial water, and their paleo-environmental implications. The cores show two lapilli tephra layers and one rhyolitic ash layer at the boundaries between sedimentary lithofacies units I, II, and III. These layers can be correlated with the well-known Ulleung-Oki (ca. 9.3 ka; boundary of units I/II), Ulleung-Yamato (ca. 25–33 ka), and Aira-Tanzawa (ca. 25.1 ka; boundary of units II/III) layers, respectively. These data suggest that the cores cover the period from the middle stage of marine isotope stage (MIS) 3 to the Holocene. In each core, a so-called dark laminated mud (DLM) layer in unit II commonly has high contents of Si and Al, suggesting that the DLM layer generally contains significant amounts of fine-grained silicates and/or aluminosilicates. In addition, these DLM layers are enriched in Fe relative to Mn, iron being predominantly bound in framboidal pyrite. The size distribution of pyrite in the DLM layers suggests that these have formed under anoxic ( euxinic) conditions at times of reduced water circulation in the basin during sea-level lowstand ( euxinic environment). The C/N ratios [5–12] suggest that sedimentary organic matter in the cores is predominantly of marine origin. Total organic carbon (TOC) contents increased shortly before Termination I. This could have been caused by an increased flux of marine organic matter in association with sea-level rise. The C and S values of units II and III (Late Pleistocene sediments; C: <2.0%; S: 0.5–2.1%) suggest a more euxinic environment than that of unit I (Holocene sediments; C: 1.0–3.5%; S: <1.0%). Concentrations of SO$_4^{2-}$ in the interstitial water decrease with increasing burial depth, whereas CH$_4$ concentrations show the reverse trend. Therefore, it seems that sulfate reduction, probably related to microbial activity, predominates in the upper core sections (<5 m), shifting to methanogenesis in the lower core sections.

**Introduction**

The East/Japan Sea is a semi-enclosed marginal sea between the Asian continent and the Japanese island arc, connected to the open ocean through 12–140 m deep straits (Chough et al. 2000). East Sea sediments are characterized by decimeter- to meter-thick alternations of light olive gray and dark olive gray muds, composed of varying proportions of siliceous and calcareous sediments (Chough et al. 2000; Tada 2002). Of these, the uppermost sediments consist mainly of alternating turbiditic and hemipelagic muds deposited during the Late Quaternary period (Bahk et al. 2000; Park et al. 2002).

The Ulleung Basin, located in the southwestern part of the East Sea, has been strongly affected by paleoceanographic changes since the Early Quaternary (Lee et al. 1996). Above all, slope stability and bottom-water conditions experienced strong variations. Large-scale slope failures and other associated depositional processes, such as downslope slide/slump and debris flow events, often occurred along the margins of the basin and in the troughs (Lee et al. 1996; Chough et al. 2000).
The uppermost (about 10 m) sediments in the East Sea are interbedded with several tephra layers that provide a useful tool for tephrostratigraphic correlation with dated volcanic eruptions (Arai et al. 1981; Oba et al. 1991; Machida and Arai 1992; Chun et al. 1997; Miyairi et al. 2004). In the southern East Sea, well-known tephra layers are the Ulleung-Oki (U-Oki; ca. 9.3 ka), Aira-Tanzawa (AT; ca. 25.1 ka) and Ulleung-Yamato (U-Ym; ca. 25–33 ka) layers (Machida and Arai 1992; Chun et al. 1997; Park et al. 2003a; Miyairi et al. 2004). According to Machida and Arai (1992), the Ulleung tephra layers are found not only in the southern part of the East Sea but also in central Japan. Therefore, the fallout distribution of the Ulleung tephra layers suggests a predominant NW wind regime. Furthermore, Park et al. (2003b) found that the occurrence of the tephra layers and their stratigraphic correlation extended the distribution of pumice-type glass shards, originating from Ulleung Island, about 50–100 km further west of the previously known fallout zone.

In the Ulleung Basin, so-called dark laminated mud (DLM) layers commonly occur in the upper part of the hemipelagic deposits. It is considered that these DLM layers consist mainly of Mn-carbonates in which the average Mn/Ca ratio is about 2.3, suggesting that the crystallites are mixed Mn-carbonates or Ca-rhodochrosites (Bahk et al. 2001). Also, Bahk et al. (2001) suggested that these laterally continuous, non-porous laminated intervals of Mn-carbonate indicate formation at the sediment–water interface, rather than by subsurface diagenetic precipitation (see also Sternbeck and Sohlenius 1997). We have also found dark laminated mud layers in our cores from this region, which can be correlated with those documented by Bahk et al. (2000, 2001), but our preliminary results (KIGAM 2002) indicate that, in this case, the DLM layers are not composed of mixed Mn-carbonates but mainly of pyrite-bearing muddy sediments.

Within this context, we have (1) investigated the occurrence and geochemical characteristics of Late Quaternary DLM layers, and (2) measured major ion concentrations of interstitial water in cores from the western Ulleung Basin. In addition, we have (3) established a stratigraphic framework for these cores, using tephrostratigraphic correlation with dated volcanic eruptions, and finally (4) evaluated the possible paleoenvironmental implications of these tephrostratigraphic/geochemical data.

Geological setting

In the East Sea, there are three main basins—the Japan, Yamato, and Ulleung basins (Fig. 1a). The Ulleung Basin is located in the southwestern part of the East Sea, surrounded by the continental slopes of the Korean Peninsula and the southwestern Honshu Island of Japan, and by the volcanic islands Ulleung, Dok, and Oki (Fig. 1b). The northern and western margins of the basin are steep (> 10°), whereas the southern and eastern margins have more gentle slopes (1–2°) and a broad continental shelf (about 50–150 km wide). The basin floor shows water depths of 1,500–2,300 m, and gradually deepens toward the Ulleung interplain gap (about 2,400–3,200 m depth; Fig. 1b).

The Ulleung Basin was formed by extension of the continental crust, accompanied by a progressive southward or southeastward drift of the Japanese island arc during the Late Oligocene to Early Miocene (Chough et al. 2000). Since the Late Miocene, the southern margin of the basin has been under compressive deformation (Jolivet et al. 1995; Chough et al. 1997).

The sedimentary sequence of the basin consists of four seismic units (Chough et al. 2000): (1) the lowermost unit consists of volcanic flows and sills intercalating with sedimentary deposits; (2) the second lowest unit (Middle Miocene sequence) is uniform in thickness, and its seismic facies is indicative of marine shale of various origins on the basin plains; (3) the middle unit (Upper Miocene to Lower Pliocene sequences) comprises thick sections of marine shale interbedded with thin sandstone and siltstone beds containing slump deposits; and (4) the uppermost unit (Upper Pliocene and Quaternary sequences) consists of turbidites and hemipelagic muds and/or mass-flow deposits. Core sediments (ca. 8–10 m long) from the basin generally show a change in lithology.

Fig. 1 a Major physiographic features of the East Sea. b Location of the two piston cores in the Ulleung Basin. UIG Ulleung interplain gap. Bathymetry in meters
from Late Pleistocene turbidites to Holocene hemipelagic muds (Lee et al. 1996; Bahk et al. 2000).

**Materials and methods**

**Materials**

Two piston cores, core 01GHP-5 (37°54′12.6″N, 130°47′17.0″E; water depth 2,170 m) and core 01GHP-7 (37°03′02.2″N, 130°14′55.5″E; water depth 2,154 m), were collected in the northwestern part of the Ulleung Basin, during the 2002 cruise aboard the R/V Tamhae II of the Korea Institute of Geoscience and Mineral Resources (Fig. 1). The cores were cut in the laboratory, and half of each core was preserved as archive core, the other half being processed as a working core, the lithologies being established by macroscopic visual inspection. Color descriptions were carried out using the Munsell color chart (Munsell 1923; Goddard et al. 1951).

**Analytical procedures**

From each tephra layer, glass shards of the coarse fraction (>63 μm) were examined by scanning electron microscopy (SEM; JEOL JSM-5410). Major element compositions of the volcanic glasses were analyzed on single glass shards using SEM in combination with an OXFORD ISIS-300 energy-dispersive X-ray spectrometer (SEM-EDS) at 1-nA current, 20-kV voltage, and with a beam diameter of 10 μm. Relative accuracy was 1–3%.

The fine structure and mineral composition of the DLM layers were investigated by means of optical micrographs on polished thin sections of impregnated sediments (for methods, see Jim 1985). Impregnation was by dehydration of wet sediments with acetone, and embedding the dehydrated sediments in Spurr low-viscosity resin. Elemental compositions were determined on the polished thin sections using a Microanalysis OXFORD (EDS) apparatus at 1-nA current, 20-kV voltage, and with a beam diameter of 10 μm. Compositional variations were assessed by line measurement of the relative contents of Fe, S, Mn, Ca, Al, and Si (count/s).

The cores were routinely sampled at 20-cm intervals, and the material freeze-dried for 24 h. The dried samples were then powdered in an agate vessel for total organic carbon (TOC), total nitrogen (TN), and total sulfur (TS) elemental analyses. TN content was determined by the combustion method, using a LECO CHN-900 apparatus, and TS content was measured by means of a LECO SC-132 analyzer. Detection limits were 0.01 and 0.005% for TN and TS, respectively. TOC was determined by adding the pyrolyzed carbon (PC) and residual carbon (RC) values, using VINC1 Rock-Eval 6 (Lafargue et al. 1998). TOC, TN and TS values are expressed in units of dry weight% of bulk sediment.

For the headspace-gas analysis of CH₄, a 5-cm³ sample taken from each core immediately onboard was placed in a 20-cm³ glass vial (Pimmel and Claypool 2001). The vials were then heated in an oven at 70°C for 30 min. The gas extracted through the septum was injected with a glass syringe into an HP 5890II gas chromatograph.

Interstitial water was extracted from sediment samples, transferred to 50-cm³ Nalgen polyethylene tubes and then centrifuged for 30 min. Water was collected with a syringe and filtered through 0.45-μm membrane filters. Chloride concentration was determined by Mohr titration with silver nitrate, using the indicator potassium chromate/potassium dichromate (Gieskes et al. 1991). Sulfate was analyzed by means of a Dionex 80 IC Analyzer ion chromatograph.

**Results**

**Lithology**

Sediment cores 01GHP-5 and 01GHP-7 consist mainly of muddy sediments interbedded with silty sands, lapilli, and ash layers (Fig. 2). The muddy sediments are mostly olive gray to dark olive gray (5GY 5/1 to 5GY 3/1), whereas the coarser sediments (silty sands and sandy muds) are light olive gray (5GY 7/1) to olive gray (5GY 4/1).

The cores consist of two types of muddy sediments: hemipelagic (units I and III) and non-hemipelagic muds (unit II). The boundaries of the units correspond roughly to those of MISs 1 and 2 (Fig. 2). The hemipelagic facies is dominated by bioturbated and slightly laminated mud, whereas the non-hemipelagic facies includes mainly thinly laminated and homogenous mud. Thus, the muddy sediments in the cores can be broadly divided into four mud types, i.e., laminated mud, homogenous mud, slightly laminated mud, and bioturbated mud.

In the laminated mud, laminae are mostly less than a few millimeters thick and consist of silt–clay couplets, whereas in the slightly laminated mud the laminae are less sharp, and not distinctly differentiated into silty and clayey layers. The most common mud type is bioturbated mud, characterized by burrow structures of circular, oval or tubular shapes. The thickness of the bioturbated mud is very variable, ranging from a few centimeters to more than 1 m. Some scattered foraminal shells and volcanic grains are often found. The mud sequence is partly disturbed in core 01GHP-7 (Fig. 2).

**Tephrostratigraphy**

Several lapilli and ash layers were observed in the two cores, but there are two main lapilli tephra layers and one main rhyolitic ash layer.
The lapilli layers consist predominantly of white vesiculate pumices and massive-type glass shards associated with alkali-feldspar. The volcanic glass shards show high Na$_2$O+K$_2$O values (average: 13.6 wt%) and small amounts of FeO (average: 0.9 wt%). The massive-type glass shards of the lapilli layers are of trachytic or phonolitic alkali composition (Fig. 3). In core 01GHP-5, the upper and lower lapilli layers correlate with the U-Oki and Ulleung-Yamato (U-Ym) tephra layers, respectively, identified by Chun et al. (1997) and Park et al. (2003a, 2003b). In core 01GHP-7, by contrast, only the lapilli layer correlating with the U-Oki tephra layer was identified.

The ash layers are characteristically high in SiO$_2$ (76–79 wt%), but alkali contents (Na$_2$O, K$_2$O) are low (6–8 wt%). Among these, the rhyolitic ash layer (3.92 m b.s.f. in core 01GHP-5, and 4.4 m b.s.f. in core 01GHP-7 [meters below sea floor (m b.s.f.)]) predominantly contains typical bubble-wall and/or plane-type glass shards, as shown by Machida and Arai (1992) and Park et al. (2003a). The morphology and geochemistry of volcanic glass shards dispersed in the rhyolitic ash layer indicate that this layer can be correlated with the AT ash layer (Fig. 3).

**Dark laminated mud (DLM) layers**

Previous $^{14}$C-age dating (Bahk et al. 2001) indicates that the DLM layers formed during the last glacial and interglacial transition. Such DLM layers are found in the intervals 2.38–2.42 m b.s.f. in core 01GHP-5, and 2.65–2.74 m b.s.f. in core 01GHP-7. The layers are 4–6 cm thick, and consist of dark olive gray (5Y 3/2) laminated mud with very dark brown bands (7.5YR 2.5/3). The non-porous bands are laterally variable in thickness (0.4–1.6 mm).

The element line-scan data of Bahk et al. (2001) showed that the DLM layers can contain significant amounts of fine-grained silicates and/or aluminosilicates, and individual crystallites consist of mixed Mn-carbonates. Consistent with these earlier data, the results of the present X-ray microanalyses indicate that the DLM layers commonly contain high amounts of Si and Al (100–200 and 200–400 counts on average, respectively; Fig. 4). In contrast to Bahk et al. (2001), however, throughout the DLM layers of our cores Fe and S

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**Fig. 3 a** Classification of the glassy tephra layers based on a total alkali–silica plot (after Le Bas et al. 1986). U-Oki Ulleung-Oki (closed circles), U-Ym Ulleung-Yamato (shaded circles), AT Aira-Tanzawa (closed diamonds). **b** Distribution of the U-Oki, U-Ym and AT ashes in the diagram Na$_2$O vs. K$_2$O. The data show that the Ulleung tephra layers are more sodic and potassic than the AT ash layers. The open symbols are from previous work by Park et al. (2003a, 2003b).

**Fig. 2** Description of the cores, and inferred stratigraphy correlation based on tephra layers and $^{14}$C absolute ages. M mud, S sand, G gravel, m b.s.f. meters below sea floor.
(occurring in the form of pyrite) are more enriched (mostly < 50 and < 150 counts, respectively) than Mn, which occurs as a minor constituent only (mostly < 20 counts; Fig. 4).

In SEM images, most of the pyrite is present as framboids, including both spherical and irregularly shaped aggregates of submicron-sized pyrite microcrystals (Fig. 5). The diameter of the framboidal aggregates varies in the range 4–13 μm (mean: 6.4 ± 1.4 μm for core 01GHP-5, and 5.9 ± 1.3 μm for core 01GHP-7), whereas that of individual crystals is 0.3–1 μm.

TOC, TN, TS

In core 01GHP-5, TOC, TN and TS vary in the range 0.13–4.08, 0.01–0.40, and 0.09–2.09%, respectively, whereas in core 01GHP-7 the corresponding values are 0.24–3.32, 0.02–0.40, and 0.04–1.89%, respectively (Fig. 6). In core 01GHP-7, the contents of TOC and TN are on average slightly higher than those in core 01GHP-5. The TOC and the TN values increase sharply at the transition between units II and I in both cores. In core 01GHP-5, this increase occurs between 1.8 and 1.6 m b.s.f., whereas in core 01GHP-7 this event is recorded between 1.9 and 1.8 m b.s.f. (Fig. 6).

For units I, II and III in both cores, the C/N ratios range from ca. 5 to 12 (Fig. 7a). Overall, the C/N ratios for unit I (ca. 5–9) are smaller than those for units II and III. For unit I, the TOC and TS values vary in the range 1.0–3.4 and 0.1–0.9%, respectively, whereas the corresponding values are 0.6–1.7 and 0.5–2.1% for unit II, and 0.1–4.1 and 0.7–1.8% for unit III (Fig. 7b).

Major ions in interstitial water, and headspace gas

In the interstitial water of both cores, Cl⁻ concentrations show relatively constant values of about 540 ± 20 mM at all core depths (Fig. 8). However, SO₄²⁻ concentrations vary strongly between 26.48 and 3.12 mM in core 01GHP-5, and between 27.95 and 2.54 mM in core 01GHP-5. SO₄²⁻ decreases progressively downcore until about 6 m b.s.f. in core 01GHP-5, and about 5 m b.s.f. in core 01GHP-7. Below this depth, SO₄²⁻ concentrations are relatively constant. Least-square linear regression analyses of these downcore sulfate gradients reveal decreases of 3.64 and 3.27 mM SO₄²⁻ per meter for cores 01GHP-5 and 01GHP-7, respectively ($r^2 = 0.97$ in both cores). CH₄ concentrations are relatively low and constant in the upper 5 m of both cores (Fig. 8), but increase progressively below 5 m b.s.f.

Discussion

Framboidal pyrites and their formation under euxinic conditions

The results of the present study demonstrate that the DLM layers of cores 01GHP-5 and 01GHP-7 from the western Ulleung Basin, East/Japan Sea are enriched in framboidal pyrite relative to Mn-carbonate. Therefore, these layers can be called pyrite-bearing DLM layers.

The morphology of pyrite crystals is primarily dependent on the solution saturation state during crystal growth (Wang and Morse 1996). The cubo-octahedron to octahedron crystal shapes imply that surface nucleation was the dominant growth mechanism in the present case (Wang and Morse 1996). This suggests that the pyrite grew in a transitional state between high and low supersaturation of the elements.

According to Wilkin et al. (1996), framboid size distribution may be used to indicate whether sediments are deposited under oxic or anoxic conditions, the framboids in oxic sediments being larger than those in anoxic sediments. For the two cores in the present study, mean diameters of 5.9 ± 1.3 and 6.4 ± 1.4 μm suggest that the
sediments of units II and III were deposited in an euxinic (anoxic sulfidic) environment. These values are markedly different from those documented in oxic (e.g., Wallops Island, GA, USA) or dysoxic (e.g., Peru margin) environments, and are indicative of slightly less anoxic conditions than those inferred for the Black Sea (Wilkin et al. 1996; Calvert et al. 1996; Wilkin and Arthur 2001). Nevertheless, marine algae typically have C/N ratios between 4 and 10, whereas vascular land plants have ratios of 20 and greater (Premuzic et al. 1982; Jasper and Gagosian 1990; Stein 1991; Meyers 1994; Prahl et al. 1994). The present findings (C/N ca. 5–12) thus suggest a marine origin for the organic matter in our core material, particularly for unit I.

The relationship between TOC and TS is commonly used to differentiate euxinic from non-euxinic conditions (Leventhal 1983; Berner and Raiswell 1983). For unit I, the TS values (generally <0.9%) lie well within the range for open-marine oxic environments, on average being distinctly lower than those for units II and III (Fig. 7b) which, for the most part, are more characteristic of euxinic and transitional oxic–euxinic environments. The occurrence of pyrite in units II and III would therefore be indicative for conditions under which reactive Fe was not limiting pyritization.

Paleo-environmental implications of sulfate and methane trends

The interstitial concentrations of Cl\(^-\) and SO\(_4^{2-}\) recorded in the two cores are in good agreement with those of seawater (545 mM for Cl\(^-\), and 29 mM for SO\(_4^{2-}\); Wilson 1975). This finding suggests that the interstitial water originates from the ambient seawater.

In the interstitial water, SO\(_4^{2-}\) concentration shows a strongly decreasing trend with core depth. This indicates that bacterial sulfate reduction has occurred within the sediments. Under anaerobic conditions, sulfate is used by bacteria as an electron acceptor for the oxidation of organic carbon, the generalized reaction being (Canfield and Raiswell 1991)

\[
2\text{CH}_2\text{O} + \text{SO}_4^{2-} (\text{aq}) + \text{H}_3\text{O}^+ (\text{aq}) = 2\text{CO}_2 (\text{g}) + \text{H}_2\text{S} (\text{g}) + 2\text{H}_2\text{O}
\]

where \(\text{aq}\) denotes the aqueous state and \(\text{g}\) the gaseous state.

Paleo-environmental implications of TOC, TN and TS data

In the present study, the upcore increase in TOC values (by up to 2 wt%) in the transition from units II to I (Termination I) could be due to increasing marine productivity associated with a regional sea-level rise of as much as 120 m (Shackleton 1987), and/or changes in bottom-water conditions of the Ulleung Basin, as suggested by Lee et al. (1996).

The provenance of sedimentary organic matter can be ascertained by means of, amongst others, C/N ratios and \(\delta^{13}\)C signatures. Seeing that isotopic measurements were not carried out in the present study, the dataset is not unequivocal in distinguishing degraded marine planktonic from terrestrial material (cf. Tyson 1995). Nevertheless, marine algae typically have C/N ratios between 4 and 10, whereas vascular land plants have ratios of 20 and greater (Premuzic et al. 1982; Jasper and Gagosian 1990; Stein 1991; Meyers 1994; Prahl et al. 1994). The present findings (C/N ca. 5–12) thus suggest a marine origin for the organic matter in our core material, particularly for unit I.

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In this equation, SO₄²⁻ is consumed by bacteria, and generates H₂S during reduction. The resulting dissolved H₂S can be utilized in the formation of pyrite. Sulfate reduction occurs often in combination with microbial and chemical iron (III) reduction (Berner 1984; Canfield and Raiswell 1991; Vile and
Wieder 1993). As a result, the H$_2$S produced may precipitate as sedimentary iron sulfide minerals, and subsequently convert to pyrite (Berner 1984). As mentioned above, units II and III presumably formed under euxinic conditions. Provided there is sufficient Fe, authigenic minerals, such as pyrite, could have been produced because the formation of iron sulfide is dependent on the availability of reactive iron species (Werne et al. 2003).

In both cores, CH$_4$ concentrations are relatively constant (>1 μM) in the upper 5 m. This could result both from the competitive inhibition of methanogenesis by sulfate-reducing microbial consortia, and from methane oxidation (Blair and Aller 1995). Below 5 m b.s.f., however, the concentrations of CH$_4$ show gradually increasing trends. Accordingly, this would imply that bacterial sulfate reduction has been gradually replaced by methanogenesis in the deeper parts of the sediments.

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References


Goddard EN, Trask PD, de Ford RK, Rove ON, Singewald JT, Overbeck RM (1951) Rock color chart. Geol Soc Am, Colorado


Munsell AH (1923) A color notation. Munsell Color Co, Baltimore


Pimmel A, Claypool G (2001) Introduction to shipboard organic geochemistry on the JOIDES resolution. Texas A&M University, Tech Note 30


