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Development and timing of two orthogonal fold systems in the western Gyeonggi Massif, Korea

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We present the deformation history based on analysis of two orthogonal fold phases and U–Pb zircon geochronology of metasedimentary and igneous rocks from the western Gyeonggi Massif (Gukhwado and Ippado areas) in the Korean Peninsula. Multiple generations of folds are recognized in three lithological units: (1) a psammitic/pelitic unit consisting of quartz–mica schist and quartzite, (2) a calcareous unit consisting of calcareous schist and calc-silicate, and (3) a mafic unit consisting of amphibolite with and without garnet porphyroblast. These rocks are intruded by a Triassic syenite. First-generation folds (F_n) are tight to isoclinal, oriented N–S, and have both shallow and steep fold axes. The F_n folds include asymmetric and buckle folds that are refolded by vertical F_{n+1} folds. The latter are oriented ~E–W, approximately orthogonal to the earlier folds. We infer from U–Pb zircon ages of the psammitic/calcareous units and the syenite that the two superimposed folding phases associated with crustal-thickening and shearing in the western Gyeonggi Massif occurred between the Pennsylvanian and the Middle Triassic (~228 Ma).

Introduction

Folds on different scales have long attracted the attention of structural geologists. They are highly variable in shape due to differences of viscosity ratio, anisotropy, and the thickness and spacing distribution of layers of different competence (Hudleston and Stephansson, 1973; Hudleston and Lan, 1992; Carreras et al., 2005; Hudleston and Treagus, 2010; Ord and Hobbs, 2013; Butler et al., 2019). Folds in layered rocks can be classified by passive and active folds based on the relation between how stress is applied and fold mechanism (Tanner, 1989; Carreras et al., 2005; Llorens et al., 2013; Ord and Hobbs, 2013). Furthermore, the active, passive, flexural folds can be formed by transpression that accommodates simultaneous shearing and shortening, and it is common in oblique convergent plate margins (Harland, 1971; Sanderson and Marchini, 1984; Fossen and Tikoff, 1993;

Woodcock and Schubert, 1994; Lin et al., 1998).

The identification of fold interference patterns provides important insights into the shortening history and kinematics associated with contractional deformation (Thiessen, 1986; Ramsay and Huber, 1987; Forbes et al., 2004). There are three generations of folds have been reported from the Anmyeondo and Yeongheungdo regions in the western Gyeonggi Massif (GM) in the Korean Peninsula, (Choi et al., 2015, 2016). The coexistence of buckle and passive folds in the study area appears to be formed by contraction with shearing that probably occurred during the late Paleozoic in response to the collision between the Sino–Korea Craton (SKC) and the South China Craton (SCC) (Fig. 1a; Oh et al., 2005, 2015; Kim et al., 2006; Cho et al., 2017a, b, 2018; Kim et al., 2018). Choi et al. (2015, 2016) suggested that the origin of ductile features in the western GM was associated with a combination of simultaneous shearing and shortening deformation.

The Permian–Triassic collisional orogeny affected the GM, but the exact position of this massif within the collisional belt is a matter of debate. According to some authors, the GM was part of the NE–SW-striking Hongseong–Odeasan Belt (Fig. 1) (Oh et al., 2005, 2006, 2015). Other tectonic models have attributed the western GM to the Imjingang Belt (Kim et al., 2006, 2008) or to the Gyeonggi Marginal Belt (Cho et al., 2017a, b). According to the Gyeonggi Marginal Belt model, the western GM infer one of the several sub-belts around the GM belonged to a collisional belt as like the Qinling microcontinent in China. Rocks in the western GM show evidence of multiple phases of regional metamorphism, during the Silurian–Devonian (430–400 Ma) and the Permian–Triassic (280–230 Ma), reaching upper-amphibolite-facies to high-pressure and high-temperature granulite-facies conditions (Oh et al., 2005; Kim et al., 2017b; Kim et al., 2018). However, relatively little is known about high-strain zones associated with this regional metamorphism (Kim et al., 2006; Cho et al., 2017b).

Here, we provide new structural constraints on the Permian–Triassic collisional orogeny in the western GM by recording temporal and spatial characteristics of ductile deformation in metasedimentary rocks in the areas of Gukhwado and Ippado that are located on the western margin of the Korean Peninsula (Fig. 1). The rocks in the study area show compositional layering of varied width (2–100 cm) and intense multiple folds. Our aims are to understand the spatial distribution of

Figure 1. (a) Simplified tectonic map of the northeast Asia displays the Chinese collision Belt (Qinling–Dabie–Sulu belts) and the locations of the Precambrian Gyeonggi and Yeongnam Massifs and the Phanerozoic Imjingang and Okcheon belts in the Korean Peninsula. (b) Geological map of the study area in the western Gyeonggi massif that is part of the Taean–Hongseong Complex (THC; modified from Kim et al., 2018).

the mesoscopic folds and to constrain the timing of ductile deformation using U–Pb zircon geochronology. On the basis of the obtained results, we discuss the deformational history of the western GM during the Permian–Triassic.

Geological Setting

The GM in the Korean Peninsula is a Precambrian basement terrane that is located between two Phanerozoic mobile belts, the Imjingang and Okcheon metamorphic belts (Fig. 1a; Chough, 2013). The GM consists mainly of quartzo-feldspathic banded and augen gneiss, migmatitic gneiss with minor quartzite, marble, calcsilicates, and amphibolite (Chough, 2013). In the western GM (Fig. 1b), major rock units include (1) Paleoproterozoic–Neoproterozoic paragneisses and orthogneisses with arc-related tonalite–trondhjemite–granodiorite suites dated at ~840 Ma (Kim et al., 2013; Park et al., 2017; Kim et al., 2018); (2) Paleozoic paragneisses (e.g., the Ordovician–Silurian Gwangcheon gneiss; Kim et al., 2017a), as well as metasedimentary rocks (Taean Formation; Cho et al., 2010); and (3) Mesozoic sedimentary and igneous rocks (Jeon et al., 2007; Seo et al., 2010; Kim et

al., 2011). Silurian–Devonian and Permian–Triassic high-pressure mafic to felsic granulites and retrogressed eclogite formed at the magmatic arc and continental collisional belt, respectively, occur also in the western GM. Pressure and temperature (P–T) conditions during the Silurian–Devonian metamorphic event were 9.0–12.0 kbar and 770–870 °C (Kim et al., 2017a). During the Permian–Triassic collisional orogeny, P–T conditions were 12.0–17.0 kbar and 800–900 °C, followed by retrogressed conditions at 9.0–10.0 kbar and 680–730 °C (Kim et al., 2018).

Paleozoic metasedimentary rocks in the western GM consist mainly of quartz–mica schist. Prior to the discovery of middle Paleozoic (~400 Ma) detrital zircon grains in the Taean Formation (Cho et al., 2010), the rocks had been considered to belong to Precambrian basement (Lee et al., 1989). The youngest detrital zircon ages have been used to infer that the metasedimentary rocks in the western GM were deposited during the Carboniferous (Kim et al., 2018). On the basis of the age population of detrital zircon grains (Cho et al., 2010; Kim, 2014), it has been suggested that the western GM was part of the SCC. This proposition has major implications for the understanding of the Permian–Triassic orogeny (i.e., the Songrim orogeny) in the western part of GM. The Taean–Hongseong Complex (THC; Fig. 1; Cho et al., 2017a) is

a branch of the Gyeonggi Marginal Belt that was intruded by Triassic and Jurassic granitoids (Kim et al., 2018). The fact that Devonian–Carboniferous metasedimentary rocks in the study area (Fig. 1b) were probably formed by late Paleozoic regional deformation and metamorphism means that these rocks can be used to study the nature of the younger (Permian–Triassic) phase of orogenesis in the western GM.

Quartz–mica and calcareous schists in the study area (Gukhwado and Ippado areas) have previously been interpreted to belong to the Neoproterozoic metasedimentary basement, on the basis of the age of the youngest U–Pb detrital zircon grains (Kim et al., 2018). However, these rocks are similar to the late Paleozoic metasedimentary rocks based on lithological and geochronological characteristics (Na et al., 2012; Han et al., 2016). The metasedimentary rocks in the western GM may have formed by various grades of metamorphism. The low grade metamorphic rocks (meta-arenite, calcareous sandstone, and mica schist; Taean Formation) occur in the Anmyeondo (Choi et al., 2008; Kim et al., 2017b). But the late Paleozoic metasedimentary rocks in the Ye-

ngheungdo and Daebudo area inferred from the maximum depositional ages using U–Pb detrital zircon ages consist of garnet + staurolite mica schist, calcsilicate, garnet-bearing amphibolite (Na et al., 2012; Han et al., 2016) that are quite similar to those in the study area.

Petrography

Rocks in the study area belong to three units involving a metapsammitic/metapelitic unit (quartz–mica schist and quartzite), a calcareous unit (calcareous schist and calcsilicate), and a mafic unit (amphibolite) (Fig. 2a–c). The rocks are characterized by marked compositional layering that can be recognized by alternating in various thickness of the layers, although its origin is unknown so far.

The mafic layers mainly consist of amphibolite with/without garnet porphyroblast up to 5 cm in diameter (Fig. 2b–d). The mafic unit also occur locally as boudins within the metapsammitic/metapelitic

Figure 2. Outcrop and rock-slab photos showing compositional layering in the Gukhwado and Ippado. (a) Metapsammitic/metapelitic unit showing steep S_n foliation in sample IPI6. (b) Compositional layers of calcareous and mafic units in Sample GHI8. (c) Compositional layering of metapelites and calcareous units in sample IPI9. (d) Mafic unit containing big garnet (Grt) porphyroblasts in sample IPI5. (e) and (f) Syenite crossed cut pervasive sub-vertical S_n foliation defined by compositional layering in sample GHI7. Plane view photos.

Figure 3. Photomicrographs showing mineral assemblages in three lithological units. (a) Compositional layers of mafic and metapsammitic units in sample GHI2. The mafic and metapsammitic units consist of amphibole (Amp), plagioclase (Pl), biotite (Bt) and quartz (Qtz), and Qtz, Pl and/or epidote (Ep). (b) Compositional layers of metapelite and calcareous units in sample IPI9. The metapelite consists of Bt, muscovite (Ms), Pl and Qtz, and the calcareous unit contains clinopyroxene (Cpx), calcite (Cal), Ep, Qtz with scapolite (Scp). (c) and (d) Magnified photomicrographs of box in (b) showing mineral assemblage in calcareous unit. (e) Mafic unit (amphibolite) showing Grt + hornblende (Hb) + Pl mineral assemblage with retrogressive Bt. (f) Syenite composing of Cpx, Hb, Bt with K-feldspar (Kfs) porphyry.

unit (e.g., in the Ippado area). The amphibolite consists of fine- to medium-grained amphibole, plagioclase, biotite, and quartz (Fig. 3a). These minerals are aligned parallel to the main foliation. Some of the amphibolites contain large garnet porphyroblasts with inclusions of plagioclase, amphibole, biotite, and Fe–Ti oxides (Fig. 3e).

The calcareous unit shows alternating layers with varied occurrences of calcsilicate and carbonate minerals. Abundant Ca-bearing minerals are recognized in pale- to intermediate-greenish and white layers (Fig. 2b, c). These layers alternate with white quartz- and calcite-rich layers (Fig. 2b, c). The calcareous schist contains diopside, plagioclase, amphibole, epidote, quartz, calcite, and minor scapolite (Fig. 3b). Greenish layers are characterized by a mineral assemblage of diopside, amphibole, epidote, and plagioclase. White layers consist of quartz

and calcite, with scapolite occurring in some of the layers (Fig. 3c, d).

Quartz-mica schist in the metapsammitic/metapelitic unit consists mainly of quartz, plagioclase, muscovite, and biotite, and minor epidote and chlorite (Fig. 3a, b). Some of the metapelitic rocks show a mineral assemblage consisting of garnet, biotite, epidote, and plagioclase. The minerals are aligned with the main foliation, which is slightly crenulated.

Cross-cutting relations are recognized between the three lithological units and a syenite intrusion (Fig. 2e, f). The latter is characterized by medium- to coarse-grained massive porphyritic texture with feldspar phenocrysts (Fig. 2e, f). The syenite consists of euhedral K-feldspar (perthite) phenocrysts, as well as clinopyroxene, hornblende, biotite, plagioclase, and quartz (Fig. 3f).

The lithological contacts between the three units are relatively sharp (Fig. 2b, c) and lie parallel to the orientation of the mesoscopic penetrative foliation (Fig. 2e, f). The occurrence of biotite and garnet in the psammitic/pelitic unit, and diopside, scapolite, garnet, and hornblende in the calcareous and mafic units, suggests that the rocks underwent greenschist-facies to upper-amphibolite-facies metamorphism prior to the emplacement of the syenite.

Deformation History

There are two generations of foliations (S_{n-1} and S_n) in the Gukhwado and Ippado areas. The foliations can be recognized by compositional layering in the outcrop, and their attitude is closely related with development of multiple folding.

Gukhwado Area

The first generation of mesoscopic foliation (S_{n-1}) is defined by compositional layering comprising the metapsammitic/metapelitic, calcareous, and mafic units. S_{n-1} strikes predominantly E–W and dips 30°–80° to the south (Fig. 4a). This structural fabric is overprinted by S_n that strikes NNW to WNW and dips shallowly to sub-vertically (10°–89°) to the SSW or SW (Fig. 4a). Owing to folding, the early S_{n-1} foliation is sub-parallel to the S_n foliation in the limbs of F_n folds. And the fold hinges of F_n trend NNW–SSE to NNE–SSW with shallow and steep plunge (Fig. 4b).

First-generation mesoscopic folds (F_n) are tight to isoclinal, and their orientations vary from recumbent to steeply inclined (Figs 5, 6). Recumbent and shallowly plunging (<20°) F_n folds (Fig. 5a, b) occur in the northern Gukhwado area and trend NNE–SSW or E–W (Fig.

4b). A top-to-the-east vergence is indicated by F_n folds in this area (Fig. 5a, b). In the southern Gukhwado area, F_n fold axes are steeper (>45°) and trend southward (>45°; Fig. 5c, d). The presence of asymmetric minor folds, boudins, and buckle folds around the limbs of F_n folds (Fig. 5f, g, h) indicates that folding was accompanied by non-coaxial deformation. The sense of shear associated with asymmetric minor folds is top-to-the-north or top-to-the-south (Fig. 5f, g) that only occurs in high strain zone. However mineral or stretching lineation and shear sense indicators do not well recognize in the folded layers. The buckle folds were developed in the quartz-rich metapsammitic layers of the compositional layers, and the axes of the buckle folds (Fig. 5h) plunge shallowly (<25°) to the south.

Second-generation F_{n+1} folds are open, and their orientations vary from inclined to vertical folds (Figs 4b, 5e). F_{n+1} fold axes plunge steeply (>45°) towards the northeast or east (Figs 4b, 6).

Ippado Area

Mesoscopic penetrative foliation (S_n) in the Ippado area is defined by compositional layering. The S_n foliation strikes mainly N–S and dips steeply (50°–87°) to the east or west (Fig. 7a). An earlier crenulated foliation (S_{n-1}) is recognized locally (Fig. 8a), striking ~E–W and dipping steeply (>50°) to the south or southwest (Fig. 7a). Folding of S_{n-1} and S_n is responsible for repetitions in the compositional layering. And the fold axes of F_n and F_{n+1} are steeply plunging (Fig. 7b).

F_n folds in the Ippado area are tight to isoclinal folds, and their axis orientations vary from moderately inclined to vertical folds. Minor buckle folds (Figs. 6, 8) are similar to those found in the Gukhwado area. F_n fold axes are predominantly steeply plunging (>45°) northward (Figs. 7b, 8b, c). Various types of asymmetric fold and buckle

Figure 4. Geological map and cross sections of the Gukhwado. (a) A structural map showing orientation of S_{n-1} and S_n foliations with pole diagrams. (b) A structural map showing orientation of F_n and F_{n+1} fold axes with pole diagrams.

Figure 5. Outcrop photos showing various types of F_n and F_{n+1} folds in the Gukhwado. (a) Isoclinal recumbent F_n fold. (b) Tight and shallowly inclined F_n fold with E-vergence. (c) Steeply to moderately plunging inclined F_n fold. (d) Steeply plunged inclined F_n fold. (e) Refolded F_n fold by E-trending F_{n+1} fold. (f) Minor asymmetric F_n fold showing N-vergence. (g) Asymmetric boudins of mafic unit formed by simple shearing (top-down-to-the-SW; normal sense). (h) Buckle fold of thin quartz-rich layer within the calcareous unit. Photos (c), (d) and (e) are plane-view, and the others are cross-section.

fold are recognized around the limbs of the F_n folds (Fig. 8d, e). Their origin is attributed to both simple shear and pure shear. The shear sense associated with the sub-vertical S_n foliation is sinistral towards

the north (Fig. 8d). Buckle folds are defined by folded quartz-rich thin layers; the axes of these folds plunge $\sim 10^\circ$ towards the north (Fig. 8e).

Second-generation folds (F_{n+1}) are typically vertical or steeply inclined

Figure 6. Classification of F_n and F_{n+1} folds in the Gukhwado and Ippado based on the orientation of the hinge line and the axial surface (Fleuty, 1964).

(Fig. 6). Their axial plane strikes ENE to WSW and dips steeply to the south (Fig. 8f). The fold axes plunge steeply ($>45^\circ$) towards the NNE or SSW ($>45^\circ$; Fig. 7b).

U–Pb Zircon Ages

U–Pb isotopic analyses of zircon grains were conducted using a laser ablation–multicollector–inductively coupled plasma–mass spectrometer (LA-MC-ICPMS; Nu Plasma II/NWR193UC) and a sensitive high-resolution ion microprobe (SHRIMP IIe/MC) at the Korea Basic Science Institute, Ochang, Korea. Calibration of inter-element fractionation was performed using zircon standard 91500 (age 1065 Ma; Wiedenbeck et al., 1995). The zircon Plešovice was employed as the external standard, with a recommended weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 337.1 Ma (Sláma et al., 2008). Zircon standards SL13 and FC1 were also used for SHRIMP analysis. Reduction of raw data and final age calculations were conducted using the SQUID and Isoplot/Ex add-ins for Microsoft Excel, and IsoplotR (Vermeesch, 2018). U–Pb zircon data are reported in Supplementary Tables S1 and S2. Analyses with discordance of $<15\%$ were selected for relative probability plots. $^{207}\text{Pb}/^{206}\text{Pb}$ ages were adopted for ages >1000 Ma. For younger

ages, we used $^{206}\text{Pb}/^{238}\text{U}$ ages.

U–Pb age analysis was conducted on five samples, with the aim of constraining the timing of deformation in the Gukhwado and Ippado areas. The three samples from the psammitic and calcareous schist (GHI3, GHI7, and IPI1; Table S1) contain rounded to subrounded detrital zircon grains varying from 60 to 150 μm in size (Fig. 9a–c). These zircons display oscillatory and banded internal zoning, which in some cases is associated with an inherited core. Some grains are mantled by a thin, bright cathodoluminescence (CL) rim. The other two samples (GHI7-1 and GHI7-2; Table S2) are from the syenite intrusion. They contain euhedral igneous zircon grains (50–200 μm in size) that are characterized by strong oscillatory internal zoning (Fig. 9e).

In total, 240 of 282 U–Pb spot analyses ($>85\%$) yielded concordant results. Detrital zircon ages obtained from the psammitic and calcareous schist samples show two prominent clusters, at 1100–700 and 500–280 Ma. The samples also have subordinate Paleoproterozoic (1,900–1,600 Ma) and Neoproterozoic (2,400–2,700 Ma) age peaks (Figs. 9a–c, 10a). The majority of the younger concordant ages are 460–360 Ma (Ordovician–Devonian) with a few ages of 330–280 Ma (Carboniferous–Permian) (Fig. 9a–c). Th/U values of most detrital zircon grains are in the range of 0.2–4.0, which suggests a magmatic origin (e.g., Rubatto, 2002) (Fig. 10b). However, a few younger (Carboniferous–

Figure 7. Geological map and cross sections of the Ippado. (a) A structural map showing orientation of S_{n-1} and S_n foliations with pole diagrams. (b) A structural map showing orientation of F_n and F_{n+1} folds axes with pole diagrams.

Permian) grains are characterized by lower Th/U values of <0.1 (Fig. 10c) suggesting a metamorphic origin.

Analyses of syenite samples GHI7-1 (n = 27) and GHI7-2 (n = 28) yielded weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 231.89 ± 0.46 (MSWD =

0.51) and 231.80 ± 0.44 (MSWD = 0.007), respectively (Fig. 9d, e). These ages are interpreted as crystallization ages, thus indicating a Triassic emplacement of the syenite intrusion.

Figure 8. Outcrop photos showing various types of F_n and F_{n+1} folds in the Ippado. (a) The F_n fold defined by steep plunging S_{n-1} foliation that is truncated by S_n foliation in quartzite. The moderately plunging F_{n+1} open fold recognized by S_n . (b) Steeply plunging inclined and tight-isoclinal F_n fold. (c) Steeply plunging inclined F_n fold. (d) Minor asymmetric F_n fold showing N-vergence. (e) Buckle fold with north trending axis. (f) Steeply plunging inclined F_{n+1} fold. Photos (a) and (d) are plane-view, and the others are cross-section.

Discussion

Generation of two orthogonal fold systems in the western Gyeonggi Massif

The metapsammitic/metapelitic, calcareous, and mafic units in the Gukhwado and Ippado areas were subjected to intense ductile deformation that included at least two generations of folding (F_n and F_{n+1} ; Figs. 4 and 7). Structures associated with this ductile deformation include two orthogonal sets of foliations oriented ~E–W (S_{n-1}) and

~N–S (S_n) (Fig. 11), as well as various types of fold (Fig. 6). The earlier folds (F_n) are tight to isoclinal and show vertical axis to recumbent, and their fold axes plunge shallowly to steeply to the N–S or E–W (Fig. 11). They also show buckle folds (Figs. 5 and 8). F_{n+1} fold axes are typically steeply plunging to vertical (Fig. 11).

Contractational deformation is inferred to have been initiated with the formation of an asymmetric monocline and open folding of the S_{n-1} foliation. The origin of S_{n-1} is ambiguous because it is difficult to constrain the characteristics and timing of regional metamorphism that formed compositional layering in the study area. With continued contractional deformation, the S_{n-1} foliation was transposed into the verti-

Figure 9. (a)–(c) Terra-Wasserburg concordia diagrams with CL images for the U–Pb detrital zircon data from metapsammitic (IPI1) and calcareous units (samples GHI2 and GHI7) in the Gukhwado and Ippado. (d) and (e) Terra-Wasserburg diagrams with CL image showing the SHRIMP U–Pb zircon ages combined with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of syenite in the Gukhwado. Numbers in the CL images represent spot number and ages.

Figure 10. (a) Relative probability diagram with kernel densities (KDE; Vermeesch, 2018) of detrital zircon ages of the late Paleozoic metamorphic rocks in the Gukhwado and Ippado, western GM. (b) Concordant age versus Th/U ratio diagram for three samples (GHI4, 7 and IPI1). (c) Th/U ratio variation of the younger ages (<480 Ma).

cal orientation of S_n . The monoclines tightened into asymmetric folds, and their axial surfaces became sub-vertical. Buckle fold belonging to F_n folds, which have sub-horizontal north and south-trending axes (Figs 5h and 8e), might be generated by sub-horizontal E–W shortening if the buck folds, mainly resulted from pure shear. However, the single layer buckle folds can be also formed by simple shear (Llorens et al., 2013). As strain accumulated, lithological boundaries are sub-parallel to S_n (Figs. 5b, d, and 8b). The geometry of F_n folds is complicated because these earlier structures were refolded by inclined and sub-vertical F_{n+1} folds. In the hinge areas of F_{n+1} folds, F_n fold axes were rotated, thus producing shallowly inclined folds (Fig. 5e, 8a, and f).

A similar style of two superimposed folding phases has been reported from the southern THC, in the area of Anmyeondo (Choi et al., 2015). The Taean Formation in this area was regionally affected by an earlier event of tight to isoclinal folding and a later folding event that produced E–W-trending vertical folds (Choi et al., 2015). Choi et al. (2015) have further suggested that the development of mylonitic foliations sub-parallel to the axial plane of isoclinal folds

resulted from a combination of shortening and minor oblique dextral shearing (top-to-the-NNE). Therefore, the two orthogonal fold system in the study area was probably linked to the major phase of crustal thickening in the in the western GM during the late Paleozoic. However, further studies based on additional kinematic indicators in microscopic scale, considering different compositional layers, will enhance more reliable deformation model. This will also serve to elucidate the previously uncertain the Permian–Triassic collisional orogeny affected the western GM.

Timing of Deformation and Implications for Late Paleozoic Orogenesis

Our new zircon ages from the metapsammitic and calcareous units and from the syenite intrusion can be used to constrain the timing of ductile deformation in the western GM, which has hitherto been poorly constrained. The new U–Pb zircon ages indicate that two superimposed folding phases occurred after deposition of the sedimentary

Figure 11. Pole diagrams displaying orientation of all S_{n-1} and S_n foliations plus axes of F_n and F_{n+1} in the Gukhwado and Ippado.

protoliths and prior to emplacement of the syenite (~232 Ma).

Detrital age spectra in samples GHI1, GHI2, and IPI1 display two dominant peaks at 1000–700 Ma (Neoproterozoic) and 500–320 Ma (early–middle Paleozoic) including the late Carboniferous to early Permian ages (330–280 Ma; Fig. 10). Similar detrital zircon age clusters have been found in Paleozoic metasedimentary rocks from the THC (Fig. 12; Cho et al., 2010; Na et al., 2012; Kim et al., 2014). A comparison of age spectra using a Kolmogorov–Smirnov (K–S) test and R statistical software (v. 3.2.3; Marsaglia et al., 2003; Hornik, 2015) shows that the probabilities (p values) of the K–S test are mostly >0.05 (Fig. 12). We therefore suggest that the metasedimentary rocks in our study area and the metasedimentary rocks in Yeongheungdo (Na et al., 2012), the Taean Formation (Cho et al., 2010), Anmyeon-

do, and Daebudo (Kim et al., 2014) might have been derived from source rocks with similar ages. Cho (2007) and Cho et al. (2010) suggested that the depositional age of the Taean Formation is 430–420 Ma and ca. 402–229 Ma, respectively, based on SHRIMP U–Pb ages of detrital zircon and the Triassic intrusion ages. Kim et al. (2017b) also documented that the youngest detrital zircon ages of ca. 440–424 Ma, zircon overgrowth ages of ca. 277–260 Ma and metamorphic ages of ca. 243–234 Ma based on SHRIMP U–Pb ages of detrital and inherited zircon grains and LA–MC–ICPMS U–Pb ages of zircon overgrowth and SHRIMP U–Pb ages of titanite, respectively. And they interpreted that a possible maximum depositional age of the Taean Formation in the Anmyeondo is 277–260 Ma. But it appears to be ambiguous because the ages can represent metamorphic event after

Figure 12. Cumulative age probability diagram of the detrital zircon from the metapsammitic and calcareous units in the Gukhwado and Ippado, and the Carboniferous–Devonian metasedimentary rocks in the Yeongheungdo (Na et al., 2012), Taean Formation (Cho et al., 2010), Anmyeondo and Daebudo (Kim et al., 2014). Insert table shows Kolmogorov–Smirnov (K–S) p values for all samples in the Taean–Hongseong Complex. P values larger than 0.05 are in bold text with shade.

deposition of the Taean Formation. Therefore, the lower limit for the timing of two superimposed folding phases in the study area is ca. 402–330 Ma.

The upper limit for the timing of deformation can be determined by the Triassic ages of syenite and granite intrusions in the western GM. A zone of Triassic syenites occurs in the western part of the study area (Choi et al., 2015), showing a general ~N–S trend that is sub-parallel to the Paleozoic metasedimentary rocks. A syenite intrusion from the southern part of the area shown in Fig. 1b has been dated at 220.3 ± 4.0 Ma (Kim et al., 2018). Triassic granites from the southern and northern parts of the area shown in Fig. 1b are characterized by two different textures. The first type is a massive, coarse- to fine-grained biotite granite, which has been dated at 229–227 Ma by U–Pb zircon geochronology (Kim et al., 2018). The second type is a strongly foliated biotite granite, for which the timings of emplacement and deformation were constrained at ~242 Ma and ~228 Ma, respectively (Kim et al., 2018). This information, together with our new constraint on the timing of the syenite emplacement (~232 Ma), implies ~228 Ma as an upper age limit for the timing of two generations of fold phases. Consequently, we interpret that F_n and F_{n+1} folding in the western GM took place during the Pennsylvanian – the Middle Triassic. One of attractive candidate for the deformation timing is Guadalupian (ca. 280 Ma) to Camian (ca. 230 Ma) based on titanite SHRIMP U–Pb and muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages from the Taean Formation (De Jong et al., 2014, 2015; Kim et al., 2017b) indicating regional metamorphic and deformation ages. Further research on the tectonometamorphic evolution of the Permian–Triassic orogeny will require an understanding of the spatial and temporal relationships between the multiple folding events, activity along thrusts and shear zones, and peak metamorphic conditions of the different lithological units in the western GM.

Conclusions

This study documents the two superimposed folding phases that are recognized in compositional layers composing of psammitic/pelitic, calcareous, and mafic units in the western Gyeonggi Massif, Korea. Penetrative S_{n-1} and S_n foliations record two phases of ductile deformation. The earlier deformation was associated with the development of NNW- to NNE-trending folds (F_n). The second phase of deformation produced E–W-trending steeply plunging and vertical folds (F_{n+1}), which refolded the earlier structures. The diversity of fold types likely resulted from contractional deformation with shearing. On the basis of U–Pb zircon ages from the psammitic and calcareous units and from the intruded syenite, we determine that two superimposed folding phases were associated with the Permian the Middle Triassic (~228 Ma) orogenic processes. Multiple generations of fold structures in the western GM were probably generated by crustal thickening associated with this orogeny.

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