2a Palaeozoic basement and associated cover

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Pre-Cenozoic rocks of the Japanese islands are largely composed of latest Palaeozoic to Cretaceous accretionary complexes and Cretaceous granitic intrusives. Exposures of older rocks are restricted to a limited number of narrow terranes, notably the Hida, Oeyama and Hida Gaien belts (Inner Zone of SW Japan), the Kurosegawa Belt (Outer Zone of SW Japan) and the South Kitakami Belt (NE Japan). In these belts, early Palaeozoic basement rocks are typically overlain by a cover of middle Palaeozoic to Mesozoic shelf facies strata. This chapter describes these basement inliers and their cover, grouping them under four subheadings: Hida, Oeyama, Hida Gaien and South Kitakami/Kurosegawa belts. Although opinions are varied among authors whether the Unazuki Schist should be placed in the Hida Belt (TT) or in the Hida Gaien Belt (KT & NM) sections, this chapter will describe the Unazuki Schist in the Hida Belt section.

Hida Belt (TT)

The overall structure of the Japanese archipelago, particularly well displayed in SW Japan, comprises a stack of NW-rooting, subhorizontal nappes, with older sheets normally occupying upper structural positions. The Hida Belt, situated along the back-arc (northern) side of SW Japan from the Hida Mountains to Oki Island, is essentially a remnant fragment of moderately deep-level continental crust that was once a part of the Asian continental margin prior to the opening of the Japanese Sea in Miocene times. The continental orogenic history recorded in the Hida Belt is therefore markedly different from other geotectonic units that instead record oceanwards growth and landwards erosion during Pacific (Cordilleran) -type orogenesis. Komatsu (1990) postulated that the Hida Belt has been thrust southwards as a large-scale nappe onto the Hida Gaien (Hida Marginal) Belt.

Polymetamorphosed Permo-Triassic granite-gneiss complexes with migmatite, impure marble, amphibolite and minor high-aluminous pelitic schist occur as a basement terrane in the Hida Mountains in central western Japan (Fig. 2a.1), and have been referred to informally as the 'Hida Gneiss' since the 1930s (Kobayashi 1938). Most rocks in this Hida Gneiss Complex are characterized by amphibolite facies mineralogy, and a few small inliers of similar rocks in Oki Island and the northern Chugoku Mountains of SW Japan have been regarded as western extensions of the same metamorphic unit. Miyashiro (1961) postulated the concept of 'paired metamorphic belts', and considered that the Hida low-pressure/high-temperature metamorphic belt was paired with the 'Sangun' high-pressure/low-temperature metamorphic belt (Fig. 2a.2). However, geochronological data for these metamorphic rocks have revealed that the timing of metamorphism of the Hida Belt was not coeval with that in the 'Sangun' Belt (e.g. Isozaki 1997; Isozaki *et al.* 2010; Wakita 2013). The eastern margin of the Hida Gneiss Complex is separated by a mylonite zone from Barrovian-type, medium-pressure, pelitic schists ('Unazuki Schist') which crop out as a north–south-aligned elongated, narrow subunit 2–3 km wide and 17 km long (e.g. Kano 1990; Takagi & Hara 1994). Another important tectonic boundary within the Hida Belt is the 'Funatsu Shear Zone', which comprises dextrally sheared, mostly metagranitoid mylonitic rocks (Komatsu *et al.* 1993).

The presence in the Hida Gneiss Complex of high-aluminous metapelites, metamorphosed acidic volcanic rocks and abundant impure siliceous marble associated with orthogneiss suggest a passive-margin lithology for the protoliths, probably as continental shelf sediments and basement rock on a rifted continental margin (e.g. Sohma & Kunugiza 1993; Isozaki 1996, 1997; Wakita 2013). Finally, the granite-gneiss complexes of the Hida Belt are unconformably overlain by a cover sequence of Lower Jurassic–Lower Cretaceous shallow marine and non-marine sedimentary rocks with rare dinosaur fossils, and by a thick Cenozoic volcaniclastic succession.

Hida Gneiss Complex and related granitic rocks

In the Hida Mountains, gneissose rocks show north–south- to NNW–SSE-striking and west-dipping foliations with mineral lineations plunging gently southwards (e.g. Kano 1980; Arakawa 1982; Sohma & Akiyama 1984). The metamorphic lithologies exposed are mainly of calcareous gneiss, quartzofeldspathic gneiss, marble, amphibolite (hornblende gneiss), granitic gneiss and minor pelitic gneiss (e.g. Naito 1993). These rocks are associated with granitic plutons and migmatite showing multiple stages of anatexis, deformation and intrusion (Figs 2a.3 & 2a.4).

The Hida Gneiss Complex has been subdivided into 'inner' lower-temperature and 'outer' higher-temperature regions, based on the mineralogy of pelitic gneiss (Suzuki *et al.* 1989). Rare staurolite, cordierite and andalusite have been found in sillimanite-bearing pelitic gneiss of the inner region (e.g. Asami & Adachi 1976). Mafic gneiss and amphibolite in this inner region contain hornblende + biotite + plagioclase + K-feldspar + clinopyroxene and scapolite occurs in calcareous gneiss (Jin & Ishiwatari 1997). Garnet-biotite Fe–Mg exchange geothermometry on pelitic gneiss suggests re-equilibration at a temperature of *c.* 550–650°C and a pressure of 0.4–0.5 GPa (e.g. Suzuki *et al.* 1989; Jin & Ishiwatari 1997). Overall, the metamorphic rocks of the inner region record amphibolite facies conditions and contrast with the outer region where granulite facies orthopyroxene- and/or spinel-bearing mineral assemblages have been described from mafic gneiss and sillimanite-bearing pelitic gneiss (e.g. Sohma *et al.* 1986). Outer zone granulite facies felsic gneiss contains rare corundum + K-feldspar mineral assemblage (Suzuki & Kojima 1970), and Fe-rich mafic gneiss (up to 26.3 wt% total FeO) contains garnet +

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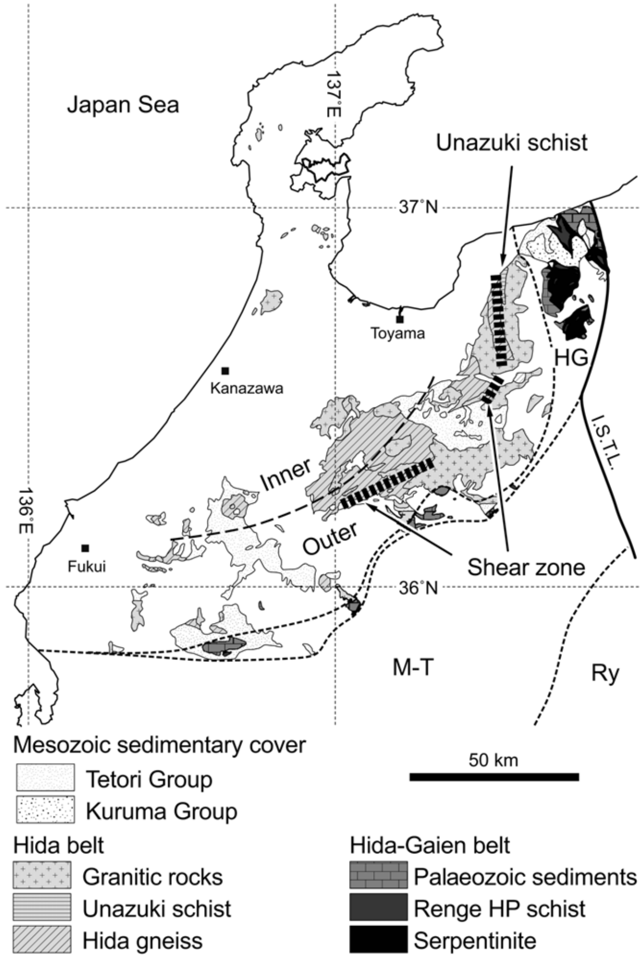

Fig. 2a.1. Simplified geological map of the Hida Mountains showing exposures of the gneiss-granite complexes of the Hida Belt and Unazuki Schist (modified after Sohma & Kunugiza 1993). Exposures of Palaeozoic rocks of the Hida Gaien Belt and Mesozoic sedimentary rocks are also shown. Dashed line is a boundary between inner and outer metamorphic regions proposed by Suzuki *et al.* (1989). Dotted lines are boundaries of geotectonic units (HG, Hida Gaien Belt; M-T, Mino–Tanba Belt; Ry, Ryoke Belt).

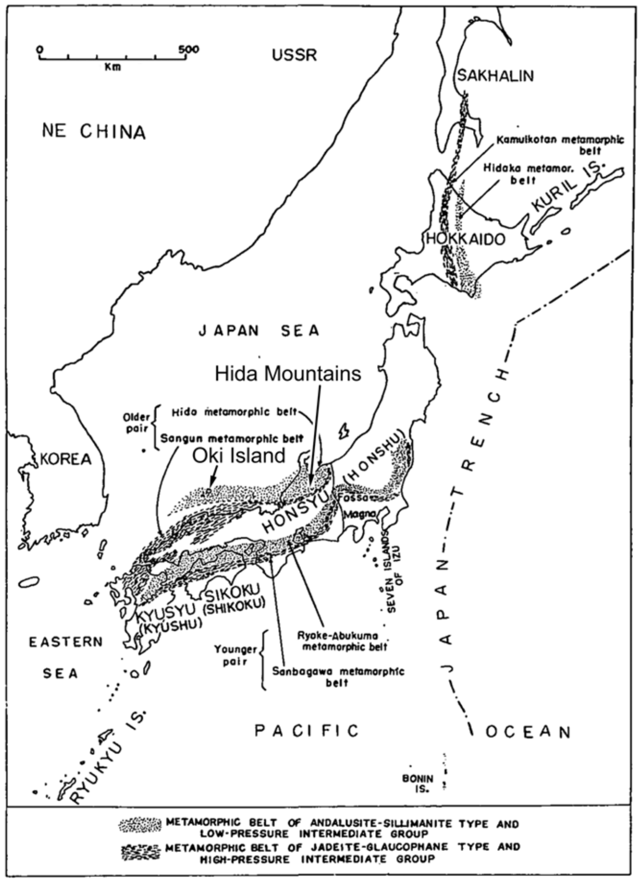

Fig. 2a.2. A conceptual map of 'paired metamorphic belts' proposed by Miyashiro (1961) [Copyright© by the Oxford University Press] showing the Hida Belt.

clinopyroxene (augite) (Suzuki 1973). In this outer region, garnet in pelitic gneiss is characterized by a higher pyrope component (up to 28 mol%) than that of the inner region, and the garnet-biotite geothermometry data indicate apparent temperatures of *c.* 700°C at a pressure of *c.* 0.4–0.7 GPa (Suzuki *et al.* 1989). Calcareous gneiss contains clinopyroxene and scapolite, and marbles contain dolomite and rare olivine (Naito 1993). Some of the clinopyroxene-rich gneiss has been interpreted as a reaction product with marble and amphibolite (Kunugiza & Goto 2006). Calcite-graphite carbon isotope geothermometry for marble of the outer zone gives a temperature of >750°C (Wada 1988). In the granulite facies marble, a millimetre-scale oxygen isotope anomaly (depleted $\delta^{18}O$ value) along calcite-calcite grain boundaries and a systematic correlation of oxygen isotope with Mn and Sr suggest a cooling rate of >1°C myr^{-1} after amphibolite facies conditions (Wada 1988; Graham *et al.* 1998). In some places Zn–Pb ore deposits are associated with skarns and migmatitic gneiss (e.g. Kano 1991; Kunugiza 1999).

Based on deformation and cross-cutting and deformational relationships in the field, 'granitic rocks' (*sensu lato*) associated with the Hida Gneiss Complex have been grouped into at least two types: older and younger 'granites'. The older granites display a wide range of rocks from gabbro (diorite) to granite and are commonly associated with migmatitic gneiss and skarns, with the gabbroic/dioritic rocks occurring as dykes (e.g. Kano & Watanabe 1995; Arakawa *et al.* 2000), and are characterized by high Sr/Y ratios and $\delta^{18}O$ values >9.8‰ (Ishihara 2005). In contrast, the younger granites have $\delta^{18}O$ values <9.1‰, cross-cut older granites, Hida Gneiss and Unazuki Schist, and can be further subdivided into pre- and post-mylonitization intrusions (Ishihara 2005). The pre-mylonitic younger granites are characterized by initial $^{87}Sr/^{86}Sr$ ratios <0.705, and are clearly distinguished from other granitic rocks in the Japanese archipelago (Arakawa 1990; Arakawa & Shinmura 1995; Jahn 2010).

Zircon ion microprobe U–Pb geochronology for pelitic gneiss has given a concordant detrital age of *c.* 1.84 Ga, whereas metamorphic zircons record 250 Ma (and more rarely *c.* 285 Ma) with discordant detrital zircons suggesting upper intercept ages of 3.42 and 2.56 Ga (Sano *et al.* 2000). Zircons from the older granites and the pre-mylonitic younger granites yielded ion microprobe U–Pb ages of 248–245 and 193 Ma, respectively; zircons from a felsic gneiss yielded *c.* 242 Ma and rare 330 Ma (Zhao *et al.* 2013). Euhedral zircons in migmatitic gneiss hosting Zn–Pb skarn deposits yield a U–Pb age of 234 Ma for a regional metamorphic event (Sakoda *et al.* 2006). Zircons in mylonitized granites in the Funatsu Shear Zone show an ion microprobe U–Pb age of 250–240 Ma and zircons from a post-mylonitization granitic intrusion yield 191 Ma, constraining the timing of mylonitization to Triassic–Early Jurassic times (Takahashi *et al.* 2010).

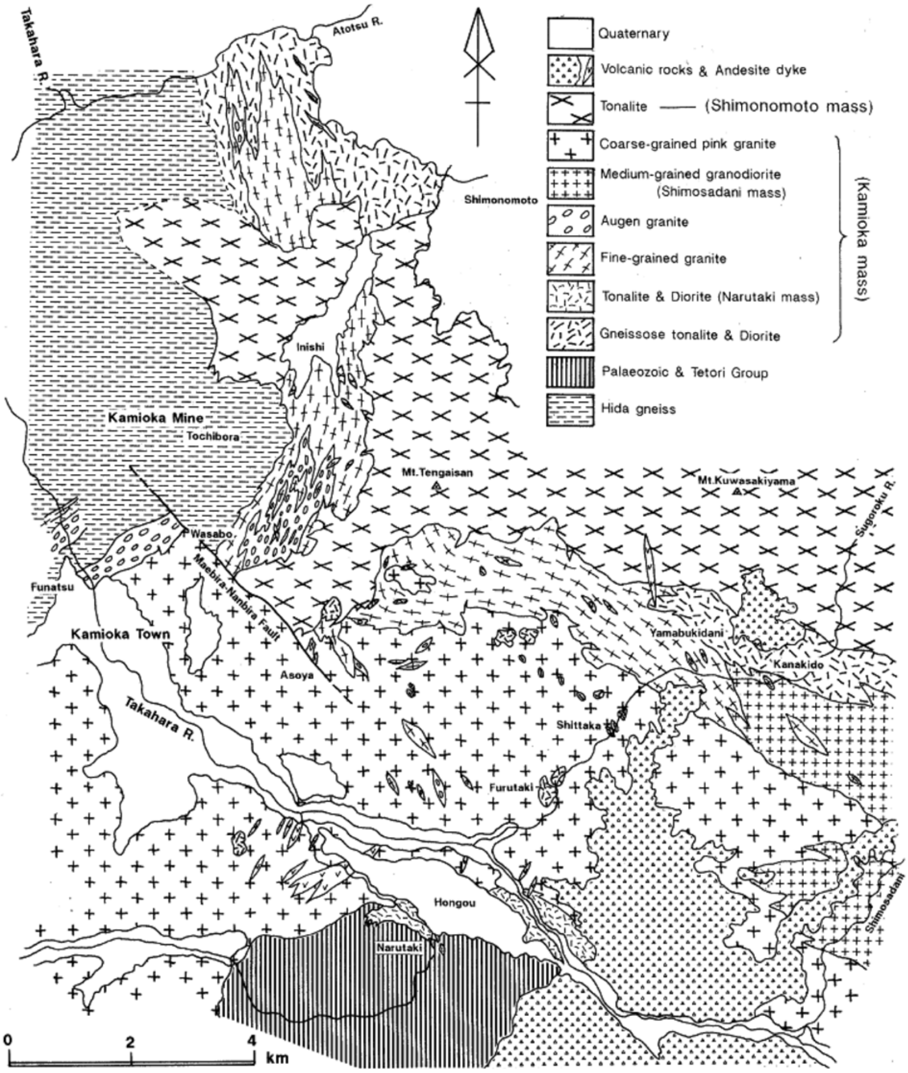

Fig. 2a.3. Geological map of the Hida gneiss-granite complex in the Kamioka area by Kano & Watanabe (1995) [Copyright© by the Geology Society of Japan]. The map shows multiple stages of deformation and igneous activities.

Electron microprobe Th–U–total Pb chemical ages of zircon in the pelitic gneiss have yielded 250–230 Ma for sillimanite-grade amphibolites facies metamorphism (Suzuki & Adachi 1994).

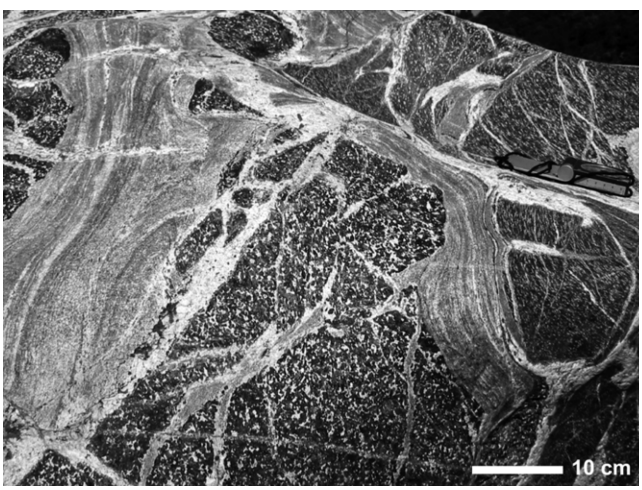

Fig. 2a.4. Amphibolite migmatite of the Hida Belt showing different stages of melting, segregation and deformation process.

Sm–Nd whole-rock–mineral isochron ages of amphibolite and pelitic gneiss are 413 ± 60 Ma and 274 ± 13 Ma, respectively (Asano *et al.* 1990). Finally, K–Ar hornblende or biotite areas in both metamorphic rocks and granitic intrusions cluster at *c.* 180 Ma (cf. Ohta & Itaya 1989).

Unazuki Schist

This noth–south-aligned elongated subunit of the Hida Belt is composed of medium-pressure-type metamorphic rocks and gabbro–diorite and granite plutons. The metamorphic rocks include mafic schist, quartzofeldspathic schist (metamorphosed rhyolite and acidic tuff, which have been described as ‘leptite’), high-aluminous pelitic schists, metamorphosed impure limestone and rare conglomerate schist (Suwa 1966; Hiroi 1978). Schistose rocks show north–south- to NNW–SSE-striking and west-dipping foliation, and the mineral lineation plunges gently to the south. The high-aluminous pelitic schists contain abundant staurolite porphyroblasts (Fig. 2a.5), with the discovery of staurolite- and kyanite-bearing pelitic schist by Ishioka (1949) being the first recognition of medium-pressure regional metamorphism in Japan. Bulk-rock compositions of the Unazuki pelitic schist are high in Al_2O_3 but low in K_2O. The compositional trend on Thompson (1957)’s AFM ternary diagram (A–Al_2O_3–K_2O, F–FeO, M–MgO, projected

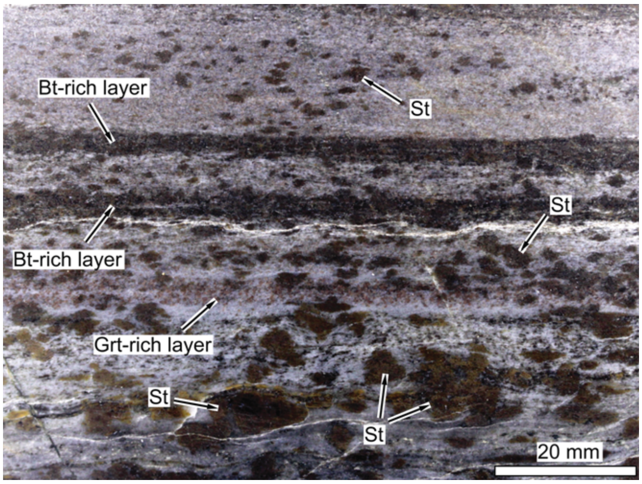

Fig. 2a.5. Polished slab of staurolite-bearing Unazuki Schist of Hiroi (1983)'s zone III (St, staurolite; Bt, biotite; Grt, garnet). The specimen shows compositional layering that might represent an original sedimentary alternation.

from ideal muscovite) is very different from that of metamorphosed trench-fill semi-pelagic sediments such as Sanbagawa pelitic schist (Fig. 2a.6), and it is noteworthy that pelitic gneiss in the Hida Belt is also typically peraluminous.

The metamorphic grade of the Unazuki Schist increases steadily southwards and shows systematic changes based on the appearance of characteristically zoned Fe–Mg silicate and aluminosilicate minerals in high-aluminous metapelite. Hiroi (1983) therefore identified four zones in order of increasing metamorphic grade: I – chloritoid + quartz; II – staurolite + chlorite + muscovite; III – kyanite + biotite; and IV – sillimanite + muscovite. These Barrovian-type mineral isograds lie obliquely over the lithological

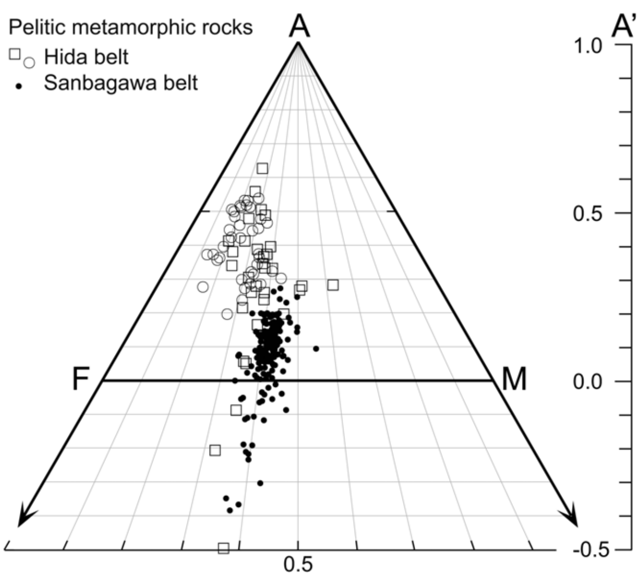

Fig. 2a.6. Thompson (1957)'s AFM diagram (A–Al_2O_3-K_2O, F–FeO, M–MgO, projected from ideal muscovite) showing bulk-rock compositions of metapelites from the Hida Belt. Unazuki Schist, open circle (Hiroi 1984); Hida Gneiss, open square (Asami & Adachi 1976; Suzuki 1977; Sohma *et al*. 1986; Naito 1993; Jin & Ishiwatari 1997) and Sanbagawa Belt (Goto *et al.* 1996). A' is Al-index in Thompson AFM projection, defined as A' = moles Al_2O_3 – 3 × moles K_2O.

structure of the schist. Bedded impure limestone of zone I contains rare fossils of Late Carboniferous bryozoa and foraminifera (Hiroi *et al.* 1978). In addition, the Unazuki Schist has been partly overprinted by contact metamorphism near younger granites, with indialite (a high-temperature polymorph of cordierite) having been found as veins in contact metamorphosed metapelite (Kitamura & Hiroi 1982) and vesuvianite + wollastonite in contact metamorphosed limestone (Okui 1985).

Zircon ion microprobe U–Pb geochronology for granitic rocks yields a concordant detrital age of 229 and 256 Ma, and inherited domains show 3.75–3.55 Ga and 1.94 Ga (Horie *et al.* 2010). The Permo-Triassic ages are consistent with Rb–Sr isochron age of the Unazuki Schist (Ishizaka & Yamaguchi 1969), whereas Rb–Sr whole-rock–muscovite and K–Ar biotite yield cooling ages of 214 and 175 Ma, respectively (Shibata *et al.* 1970).

A clast of staurolite-bearing high-aluminous pelitic schist was found in the Upper Jurassic conglomerate of the Tetori Group overlying the Hida Belt (Tsujimori 1995), and detrital chloritoid has been found in Lower Jurassic shallow-marine sandstone of the Kuruma Group, which lies on the Hida Gaien Belt (Kamikubo & Takeuchi 2010). Considering the limited number of occurrences of staurolite- and/or chloritoid-bearing metamorphic rocks in Japan, these sedimentary records suggest that the Unazuki Schist (or its equivalent) had already been exposed at the surface by Jurassic times.

Mesozoic sedimentary cover sequences

The Hida Gneiss Complex and related granitic rocks are unconformably overlain by Middle Jurassic–Lower Cretaceous Tetori Group marine and non-marine deposits. Various macrofossils, including plants, molluscs, fishes, reptiles, turtles and dinosaurs, have been found in the Tetori Group. The group is divided into Kuzuryu, Itoshiro and Akaiwa subgroups in ascending order (e.g. Maeda 1961). Illite crystallinity indicates that the Kuzuryu Subgroup exhibits a higher degree of diagenesis than the overlying Itoshiro and Akaiwa subgroups (Kim *et al.* 2007). The marine Kuzuryu Subgroup is unconformably covered by the non-marine Itoshiro Subgroup, which in turn is overlain by the non-marine Akaiwa Subgroup. Electron microprobe Th–U–total Pb geochronology shows that detrital monazites are mostly *c.* 1.74–1.25 Ga with a small peak of *c.* 250 Ma (Obayashi 1995). Recent laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) U–Pb zircon geochronology for acidic tuffs limits minimum depositional ages to 130 Ma for Kuzuryu Subgroup and 118 Ma for Itoshiro Subgroup (Kusuhashi *et al.* 2006). In conglomerate beds of the Tetori Group most abundant clasts are granitic rocks and orthoquartzites, which are likely derived from a basement rock in the East Asian continental block. However, rare radiolarian cherts derived from a Jurassic accretionary complex occur in the Akaiwa Subgroup; this suggests a change of lithology of provenance area during Early Cretaceous times (Takeuchi *et al.* 1991). Palaeomagnetic analyses show that the Tetori Basin was located at *c.* 24° N until the Late Jurassic, but during the Early Cretaceous epoch it moved northwards to 40° N (Hirooka *et al.* 2002).

Oki Gneiss Complex

In the Oki–Dogo islands, a small exposure (6 × 8 km) of gneiss-granite has been considered to mark a western extension of the Hida Gneiss Complex of the Hida Mountains. This Oki Gneiss Complex consists mainly of felsic gneiss with migmatite, granitic intrusions and minor amphibolite (Hoshino 1979). Amphibolite contains orthopyroxene-bearing mineral assemblage, and two-pyroxene geothermometry gives a granulite facies temperature of *c.* 800°C. A zircon Pb–Pb age of *c.* 1.9 Ga from the Oki Gneiss Complex has been obtained using isotope dilution thermal ionization mass spectrometry (Yamashita & Yanagi 1994), a result consistent with a Sm–Nd whole-rock isochron age 1.98 $\pm$ 0.18 Ga from amphibolite (Tanaka & Hoshino, 1987). Zircon ion microprobe U–Pb geochronology for granitic gneisses yields concordant ages of 1.87 and 1.88 Ga for the crystallization age of the granitic protolith (Cho *et al.* 2012). Moreover, recrystallized zircon rims give a concordant age of *c.* 236 Ma (Tsutsumi *et al.* 2006; Cho *et al.* 2012). Electron microprobe Th–U–total Pb chemical dating of monazite in paragneisses shows *c.* 250 Ma for regional metamorphism and rare zoned monazite preserves older ages of 1.69 $\pm$ 0.23 Ga at the core and 440 $\pm$ 30 Ma further out, with the 250 $\pm$ 20 Ma event recorded at the rims (Suzuki & Adachi 1994). Biotite K–Ar and muscovite $^{40}Ar/^{39}Ar$ plateau ages suggest a timing of cooling through closure temperature of micas at *c.* 170 Ma (Shibata & Nozawa 1966; Dallmeyer & Takasu 1998). A small outcrop of granitic gneisses found near the Daisen Volcano, interpreted as an inlier of the Hida Gneiss, has yielded a Rb–Sr whole-rock–mineral isochron (mafic gneiss) of 186 $\pm$ 6 Ma (Ishiga *et al.* 1989) and a U–Pb age of 198 $\pm$ 3 Ma (Ishihara *et al.* 2012).

Geotectonic correlation with East Asian continental margin

How does the Hida Belt correlate with petrotectonic units in NE China and the Korean Peninsula (Fig. 2a.7)? Hiroi (1981) first correlated the late Permian geotectonic unit of Unazuki Schist with the Ogcheon Belt in the Korean Peninsula, based on the occurrence of staurolite-bearing high-aluminous metapelites. Using protolith lithology and metamorphic character (Barrovian-type medium-pressure metamorphism) of the Unazuki Schist, Isozaki & Maruyama (1991) postulated that the Unazuki Schist represented an eastern extension of a Permo-Triassic collisional unit between the Shino-Korea (North China) and Yangtze (South China) blocks, and that gneiss-granite complexes in the Hida Mountains and Oki Island could be correlated with North China and South China blocks,

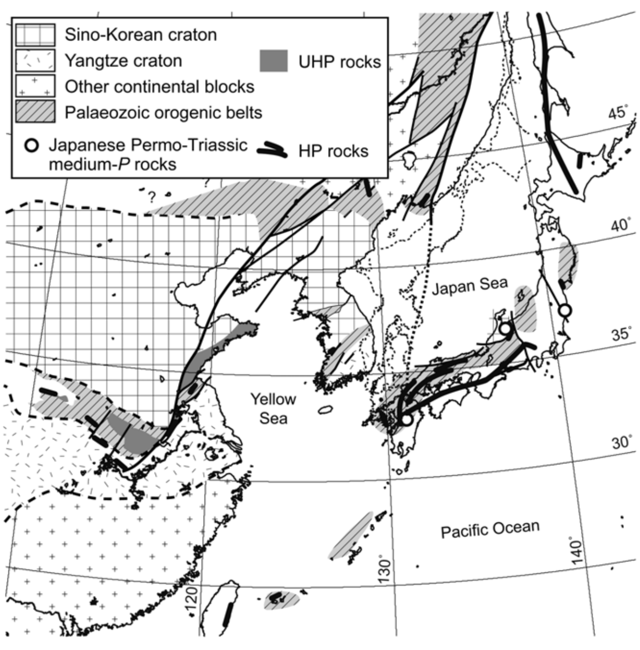

Fig. 2a.7. Generalized tectonic map of East Asia showing Permo-Triassic medium-pressure metamorphic rocks in Japan. The map is modified after Tsujimori & Liou (2005).

respectively (cf. Isozaki 1996, 1997). More recently, Isozaki *et al.* (2010) suggested that the medium-pressure-type metamorphic rocks with passive margin protoliths in the Higo, Unazuki and Hitachi units represented an eastern extension of the Permo-Triassic continent–continental collision zone in east-central China.

Ishiwatari & Tsujimori (2003) proposed the Yaeyama promontory hypothesis. They essentially supported the idea of the Unazuki Schist correlating with Permo-Triassic petrotectonic units in NE China and the Korean Peninsula, but considered that both Permo-Triassic collision-type orogeny and Pacific-type orogeny had occurred along the same plate boundary. Ernst *et al.* (2007) proposed an amalgamated suture zone 'Tongbai–Dabie–Sulu (east-central China)–Imjingang–Gyeonggi (central Korea)–Renge–Suo (SW Japan)–Sikhote-Alin Orogen' that reflects collision between the Sino–Korean and Yangtze blocks on the SW portion, and accretion of outboard oceanic arcs $\pm$ sialic fragments against the NE margin.

In the Hida Gneiss and Unazuki granites, the presence of inherited Archean and Palaeoproterozoic zircon ages and the lack of Neoproterozoic zircon suggest that this region is genetically related to the Sino–Korean Block (Sano *et al.* 2000; Horie *et al.* 2010). Based on the deformational style, zircon U–Pb geochronology and palaeogeographical location of granitic rocks, Takahashi *et al.* (2010) correlated the Funatsu Shear Zone of the Hida Mountains to the Cheongsan Shear Zone of the south-central Ogcheon Belt in the Korean Peninsula. Sr–Nd–Pb isotope geochemical study of Permo-Triassic granitic rocks in the Korean Peninsula confirmed that granitic plutons of the Gyeongsang Basin, characterized by low initial $^{87}Sr/^{86}Sr$ ratios (0.704–0.705), correlated with the pre-mylonitic younger granites of the Hida Mountains (Cheong *et al.* 2002). Jahn *et al.* (2000) linked the Hida Belt to the Central Asian Orogenic Belt, based on isotope geochemical characteristics of granitic rocks. Recently, this idea was supported by Chang & Zhao (2012)'s model in which the Permo-Triassic collisional zone was terminated by the Yellow Sea Transform Fault and did not continue into the Korean Peninsula or Hida Belt.

Because of considerable modification of the Japanese archipelago in Cenozoic times, there is still debate regarding the possible correlations between the Hida Belt and other Permo-Triassic geotectonic units along the East Asian continental margin. To further our understanding of the geotectonic configuration of this margin, a more detailed and integrated approach to geology, petrology and geochronology of metamorphic and associated granitic rocks than has so far been available is required.

Oeyama Belt (TT)

Kilometre-scale Alpine-type ultramafic bodies are sporadically exposed in the Chugoku Mountains of SW Japan, running parallel to the orogenic trend west from Oeyama to Tari–Misaka in the central Chugoku Mountains. The largest body (Sekinomiya) covers an area of *c.* 20 × 5 km (Fig. 2a.8) and the area with the most numerous outcrops is concentrated in the Central Chugoku Mountains. Arai (1980) first pointed out that these ultramafic bodies had originally constituted the lowest part of an ophiolitic suite subsequently emplaced as dismembered fragments, with the 'ophiolitic succession' being best developed in the Oeyama body (Kurokawa 1985). Overall compositional trends of Cr-spinel overlap with either forearc or abyssal peridotite. Since these ultramafic bodies have different petrological, geochemical and geochronological features from the mafic-ultramafic bodies of the Yakuno Ophiolite, Ishiwatari (1991) grouped these ultramafic bodies into the 'Oeyama Ophiolite', a unit also referred to as the 'Oeyama Belt' in geotectonic models (e.g. Isozaki & Maruyama 1991; Isozaki *et al.* 2010). The ultramafic bodies of the Oeyama Belt occupy the structurally highest position in the

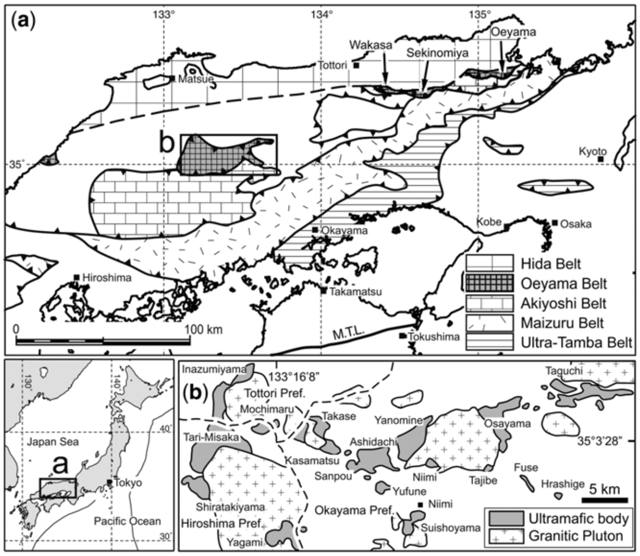

Fig. 2a.8. (**a**) Tectonic framework of the Chugoku Mountains, showing the Oeyama Belt and other Palaeozoic geotectonic units (after Tsujimori & Liou 2004). (**b**) Distributions of ultramafic bodies in the central Chugoku Mountains (after Matsumoto *et al.* 1995).

Chugoku Mountains, above the Renge, Akiyoshi and Maizuru belts (e.g. Uemura *et al.* 1979; Tsujimori & Itaya 1999), and have been broadly correlated with similar ultramafic bodies in the Hida Mountains and northern Kyushu. In the geological history of the Japanese archipelago since Early Palaeozoic time, the Oeyama Belt has acted as a basement unit for Early Palaeozoic forearc basin sediments and as a forearc mantle wedge involved with Palaeozoic subduction zone metamorphic rocks.

The ultramafic bodies of the Oeyama Belt are composed mainly of serpentinized harzburgite (more lherzolitic in eastern bodies) and dunite with minor podiform chromitite and mafic intrusions (e.g. Arai 1980). The chromitites enclosed in dunite are best developed in the Central Chugoku Mountains (Arai 1980; Hirano *et al.* 1987; Matsumoto *et al.* 1997) and more rarely in the Hida Mountains (Yamane *et al.* 1988; Tsujimori 2004). Amphibolites (metacumulate and gneissose metagabbro) occur as tectonic blocks only in the Eastern Chugoku Mountains (Kurokawa 1985; Nishimura & Shibata 1989; Tsujimori & Liou 2004), and associated jadeitite (jade) has been recorded in some places (cf. Chihara 1989).

Due to intrusions of Late Cretaceous granitic plutons, most bodies have been partly overprinted by contact metamorphism (e.g. Arai 1975; Kurokawa 1985; Matsumoto *et al.* 1995) although some ultramafic bodies, particularly in the Hida Mountains, had already undergone a regional metamorphism before contact heating (Uda 1984; Tsujimori 2004; Nozaka 2005; Khedr & Arai 2010; Machi & Ishiwatari 2010; Nozaka & Ito 2011). Sm–Nd isochron ages of *c.* 560 Ma for the gabbroic intrusions suggest a Cambrian or late Neoproterozoic igneous age (Hayasaka *et al.* 1995). Jadeitite dykes or blocks yield hydrothermal zircon U–Pb ages of *c.* 520–470 Ma (Tsujimori *et al.* 2005; Kunugiza & Goto 2010), whereas amphibolites and gneissose metagabbro give hornblende K–Ar ages of *c.* 440–400 Ma (Nishimura & Shibata 1989; Tsujimori *et al.* 2000).

Characteristics of primary peridotites

The ultramafic bodies of the Oeyama Belt in the Chugoku Mountains preserve a kilometre-scale original lithological structure of the crust–mantle transition zone or the uppermost oceanic mantle. They consist mainly of serpentinized harzburgite and dunite with minor podiform chromitite and gabbroic intrusions (e.g. Arai 1980; MITI 1993, 1994; Matsumoto *et al.* 1997). The largest chromitite pod in the Wakamatsu Mine of the Tari–Misaka body has dimensions of 40 × 25 × 210 m. Serpentinized harzburgite is characterized by occurrence of vermicular-like intergrowth of Cr-spinal and orthopyroxene, whereas Cr-spinels in dunite are euhedral to subhedral (Fig. 2a.9). The overall layered structure is not well developed in those bodies. Podiform chromitites are enclosed by dunite envelope (Fig. 2a.10), and the occurrences of relatively large chromitite pods are limited in dunite-dominant bodies. The lithological relations and mineralogical features of chromitites suggest a melt–mantle interaction and related melt mixing to precipitate Cr-spinel. The chromitite-precipitated melt was a mixture of secondary Si-rich melts formed by this interaction and the primitive magmas in the upper mantle (e.g. Arai & Yurimoto 1995). The Cr# (Cr/Cr + Al) of Cr-spinel in harzburgite, dunite and chromitite varies from 0.4 to 0.6, Cr-spinels in dunite and chromitite have slightly higher TiO_2 and the latter can contain hydrous mineral inclusions such as pargasite and Na-phlogopite (Arai 1980; Matsumoto *et al.* 1995, 1997).

In the Hida Mountains, several ultramafic bodies of the Oeyama Belt are exposed along the Hida–Gaien Belt. Although they are highly serpentinized or recrystallized, chemical compositions of relict Cr-spinel and bulk-rock compositions of serpentinite suggest mostly harzburgite and subordinate dunite protolith (e.g. Khedr & Arai 2010, 2011; Machi & Ishiwatari 2010). Based on petrological features, the ultramafic rocks have been grouped into 'serpentinized peridotite' and 'metamorphosed peridotite'. The serpentinized peridotites are subdivided into high-Al and high-Cr groups, each with Cr-spinel Cr# values of 0.3–0.4 and 0.5–0.6, respectively (Machi & Ishiwatari 2010; Khedr & Arai 2011). In contrast, relict primary Cr-spinels in metamorphosed peridotite are characterized by very high Cr# of 0.7–0.9 (Tsujimori 2004; Khedr & Arai 2011). Rare Cr-spinel in chromitite preserves primary pargasite (Tsujimori 2004).

Metamorphosed peridotites in the Hida Mountains have been subjected to hydration and metamorphism. Due to the regional deformation, highly deformed metamorphosed peridotites often show a penetrative schistosity defined by preferred orientation of antigorite and tremolite and with a trend similar to that of high-pressure/low-temperature schists in the Renge Belt (Yamazaki

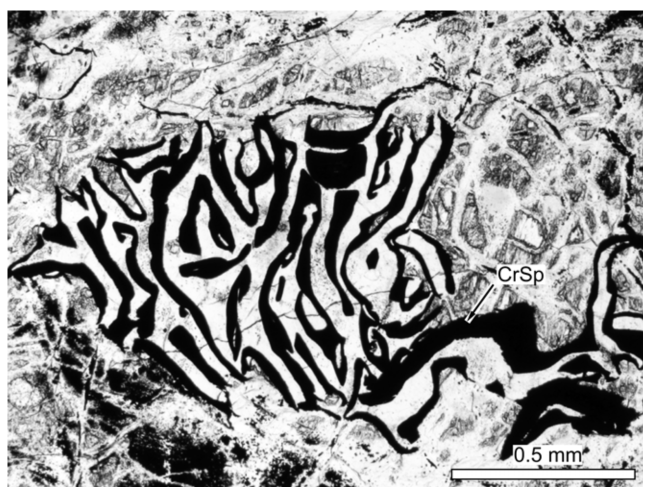

Fig. 2a.9. Photomicrograph of Cr-spinel (CrSp) in serpentinized harzburgite exhibiting a vermicular-like texture.

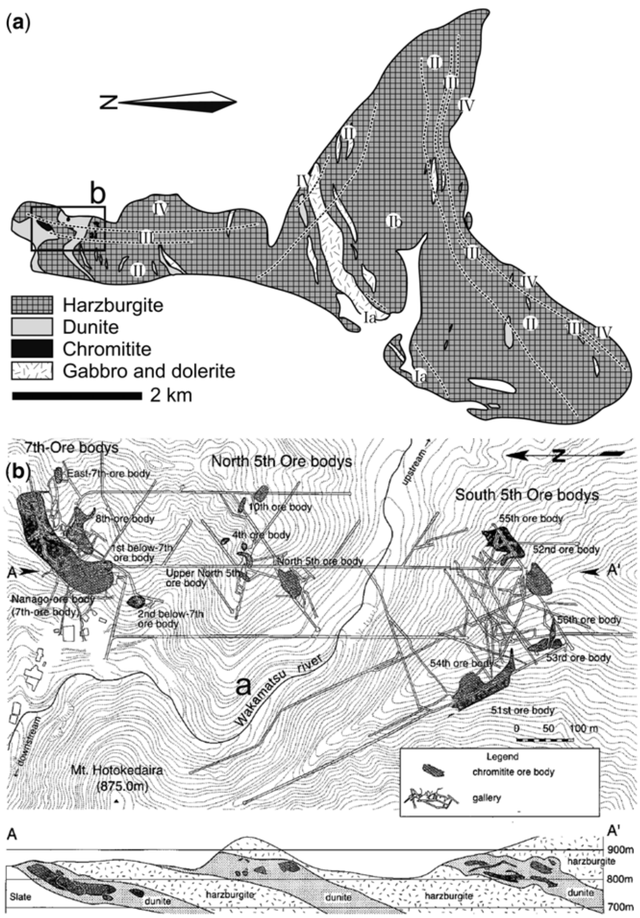

Fig. 2a.10. (**a**) Lithological map of the Tari–Misaka ultramafic body (after Matsumoto *et al.* 1997). Mineral zones of contact metamorphism by Matsumoto *et al.* (1995) are also shown. (**b**) Distribution of chromitite pods of the Wakamatsu Mine of the Tari–Misaka ultramafic body (Matsumoto *et al.* 2002 [Copyright© by the Society of Resource Geology]).

1981; Nakamizu *et al.* 1989). In the Happo peridotite body (8 × 5 km), Nakamizu *et al.* (1989) and Nozaka (2005) identified three mineral zones: the diopside zone with olivine (relic) + antigorite + diopside; the tremolite zone with olivine + tremolite + orthopyroxene; and the talc zone with olivine + talc + tremolite. The talc zone metamorphosed peridotites (tremolite–chlorite peridotites) contain a mineral assemblage of olivine + low-Al orthopyroxene + tremolite + chlorite + high-Ti-Cr# Cr-spinel, suggesting metamorphic conditions of $T = 650$–$750°C$ and $P = 1.6$–2.0 GPa (Khedr & Arai 2010). Metamorphic olivine is intergrown with antigorite and exhibits a 'cleavable' texture (Kuroda & Shimoda 1967). Tremolite contains significant Na (*c.* 0.4 wt% Na_2O), and rare richterite and edenite are associated with tremolite (Khedr & Arai 2010). In the Kotaki area, Machi & Ishiwatari (2010) have also identified deformed metaperidotite with similar petrological characteristics to the Happo peridotite.

In the Chugoku Mountains, the eastern ultramafic bodies (Oeyama, Izushi, Sekinomiya and Wakasa) often contain a characteristic mineral assemblage of olivine intergrown with antigorite showing a 'cleavable' texture (Uemura *et al.* 1979; Uda 1984; Kurokawa 1985; Nozaka & Ito 2011). The distribution of this 'cleavable' olivine is not related to aureoles of contact metamorphism by younger granitic intrusion (Nozaka & Ito 2011), but rather its ubiquitous presence in these ultramafic bodies indicates regional metamorphism of the forearc wedge-mantle in a subduction zone. However, the timing of this regional metamorphism of ultramafic rocks is still under debate.

In the topographically highest portion of the Oeyama body are exposures of metamorphosed mafic-ultramafic cumulate and amphibolite (4.5 × 1.5 km) (Fig. 2a.11) known as the 'Fuko Pass metacumulate' (Kurokawa 1975, 1985; Tsujimori 1999; Tsujimori & Ishiwatari 2002; Tsujimori & Liou 2004). This metacumulate body has been interpreted as a mafic-ultramafic cumulate member of an ophiolitic succession (Kurokawa 1985). Mafic metacumulate is subdivided three lithologic types: foliated epidote-amphibolite; leucocratic metagabbro; and melanocratic metagabbro. The foliated epidote-amphibolite and leucocratic metagabbro contain low-variance epidote-amphibolite facies mineral assemblage with kyanite, staurolite, and paragonite (Kuroda *et al.* 1976; Kurokawa 1975; Tsujimori & Liou 2004), whereas the melanocratic metagabbros preserve relict granulite facies minerals (Tsujimori & Ishiwatari 2002). Microtextural relationships and mineral chemistry define three metamorphic stages: relict granulite facies metamorphism; high-pressure epidote–amphibolite facies metamorphism; and retrogression. Relict Al-rich diopside (up to 8.5 wt% Al_2O_3) and pseudomorphs after spinel and plagioclase suggest medium-*P* granulite facies conditions ($P = 0.8$–1.3 GPa at $T > 850°C$) (Tsujimori & Ishiwatari 2002). An unusually low-variance assemblage (hornblende + clinozoisite + kyanite ± staurolite + paragonite + rutile ± albite ± corundum) constrains metamorphic P/T conditions to *c.* 1.1–1.9 GPa at 550–800°C for the high-pressure metamorphism. The breakdown of kyanite to produce a retrograde margarite-bearing assemblage at $P < 0.5$ GPa and $T = 450$–$500°C$ indicates a greenschist facies overprint during decompression (Tsujimori & Liou 2004). Foliated epidote-amphibolite yields hornblende K–Ar ages of *c.* 400–440 Ma (Tsujimori *et al.* 2000); similarly, Early Palaeozoic gneissose epidote amphibolites (but without kyanite) occur in the Wakasa ultramafic body *c.* 20 km to the west of the Oeyama area (Nishimura & Shibata 1989). The presence of Early Palaeozoic high-pressure metamorphic rocks in the Kurosegawa and Oeyama belts provides a petrotectonic constraint for the earliest subduction event in the Japanese Orogen (Tsujimori 2010). The regional metamorphism of the Oeyama Belt occurred in an Early Palaeozoic subduction zone with geothermal gradient of the order 15°C km^{-1}, a relatively high value that might explain the

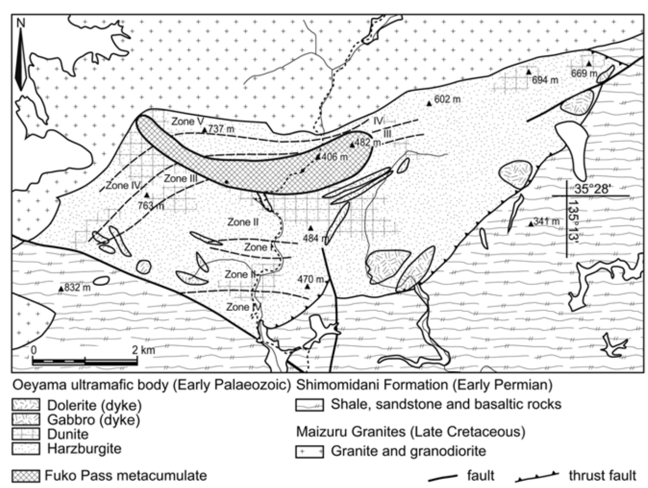

Fig. 2a.11. Geological map of the Oeyama area (after Kurokawa 1985; Tsujimori *et al.* 2000).

epidote-amphibolite facies hydrous recrystallization of the mafic cumulates and ultramafic rocks.

Jadeitite

Jadeitite is a plate tectonic gemstone that correlates forearc mantle wedge and high-pressure and low-temperature metamorphism within a supra-subduction zone at relatively shallow depths (<100 km) (Stern *et al.* 2013; Harlow *et al.* 2015). Since Kawano (1939) first identified jadeitite as boulders in the Kotaki River of the Hida Mountains, numerous jadeitite boulders have been found from at least six ultramafic bodies (Kotaki, Omi, Happo, Sekinomiya, Wakasa and Osayama) within the Oeyama Belt (e.g. Masutomi 1966; Chihara 1971; Tazaki & Ishiuchi 1976; Kobayashi *et al.* 1987) (Fig. 2a.12). The jadeitite localities are characterized by serpentinite mélange and accompanied by Late Palaeozoic high-pressure and low-temperature metamorphic rocks (see 'Renge rocks in the Hida Mountains' section in Chapter 2b). It is noteworthy that the age of jadeitite formation is significantly older than the blueschist facies metamorphism of the Renge rocks in the same mélange (Tsujimori & Harlow 2012).

Most jadeitites are principally fluid precipitates (P-type), but a few formed by metasomatic replacement (R-type) and hence preserve relict minerals and protolith textures (Tsujimori & Harlow 2012). Rutile and zircon are common accessory minerals in the jadeitites, with the rutile-bearing jadeitite having formed at higher T and P than blueschist facies jadeitite. In the Osayama jadeitite, oscillatory-zoned rutile- and jadeite-bearing zircon yielded ion-microprobe U–Pb ages scattering over the range 521–451 Ma (weighted mean age 472 ± 8.5 Ma). Inherited igneous cores of zoned zircons suggest possible protoliths of gabbro, norite or plagiogranite and therefore an oceanic crustal origin (Fu *et al.* 2010), with ages in the range 523–488 Ma (Tsujimori *et al.* 2005). Oxygen isotope compositions of zircon formed during jadeitite formation is lighter ($\delta^{18}O = 3.6 \pm 0.6$) than that of the inherited igneous core formed in equilibrium with mantle compositions ($\delta^{18}O = 5.0 \pm 0.4$) (Fu *et al.* 2010). In the Itoigawa–Omi area, zircons in jadeitites interpreted as fluid precipitates on the basis of their rhythmic zoning, inclusion suites and rare earth element (REE) patterns, yield ion-microprobe U–Pb ages of 519 ± 17 and 512 ± 7 Ma (Kunugiza & Goto 2010). The jadeitite formation is likely to be coeval with the Early Palaeozoic epidote-amphibolite facies metamorphism of the Oeyama Belt.

Contact metamorphism around granitic plutons

Mineral zones of contact metamorphism have been mapped in the ultramafic bodies of the Central Chugoku Mountains (Arai 1975; MITI 1993, 1994; Matsumoto *et al.* 1995). Arai (1975) described four zones in order of increasing metamorphic grade: I – antigorite; II – olivine + talc; III – olivine + 'anthophyllite'; and IV – olivine + orthopyroxene. Zone I was not subjected to thermal metamorphism and can be subdivided into chrysotile-lizardite and antigorite subzones (Matsumoto *et al.* 1995). In the highest grade of contact aureole, porphyroblastic spinifex-like olivine and radial aggregates or spherulitic shapes of orthopyroxene occur (Arai 1975). 'Anthophyllite' in the Tari–Misaka body has *Pnma* crystal structure and is mineralogically classified as protoanthophyllite; some protoanthophyllite contains lamellae of anthophyllite with $C2/m$ structure (Konishi *et al.* 2002, 2003).

Hida Gaien Belt (KT & MN)

This belt, variously referred to as the Hida marginal belt, Circum-Hida tectonic belt and so on, is defined here as the Hida Gaien Belt after Kojima *et al.* (2004). The rocks of the Hida Gaien Belt are restricted to narrow outcrops between the Hida and Sangun–Renge–Akiyoshi–Maizuru–Ultra-Tanba–Mino belts in central Japan (Figs 2a.13 & 2a.14). The Carboniferous rocks in the Unazuki area, referred to as the Unazuki Schist, is included in the Hida Gaien Belt in some studies based on its lithostratigraphy and palaeontology (e.g. Yamakita & Otoh 1987; Tsukada *et al.* 2004); however, it is described in the Hida Belt section here. The Nagato tectonic zone dividing the Sangun–Renge and Akiyoshi belts, SW Japan (Kabashima *et al.* 1993), is considered to be a western extension of this belt (Isozaki & Tamura 1989).

This belt was firstly defined by Kamei (1955*a*) as a complex zone dividing the continental massif of the Hida Belt and the 'Palaeozoic geosynclinal facies' rocks of the Mino Belt, and has been subsequently regarded as a suture zone between the continental mass

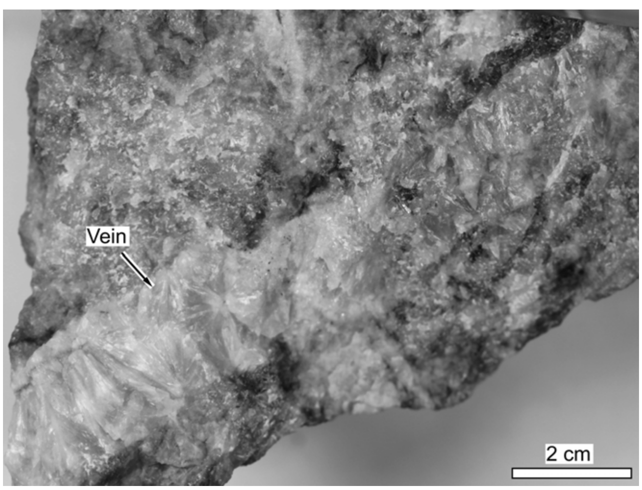

Fig. 2a.12. Early Palaeozoic jadeitite from the Osayama area, showing a texture of fluid precipitates. Radial aggregates of coarse-grained jadeite crystals were precipitated directly from a Na–Al–Si-rich aqueous fluid in a vein within pre-existing jadeitite (cf. Tsujimori & Harlow 2012).

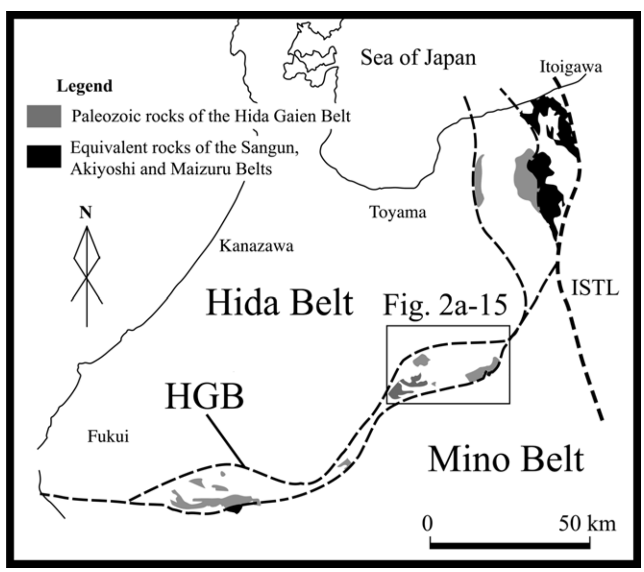

Fig. 2a.13. Index map of the Hida Gaien Belt. ISTL, Itoigawa–Shizuoka Tectonic Line; HGB, Hida Gaien Belt.

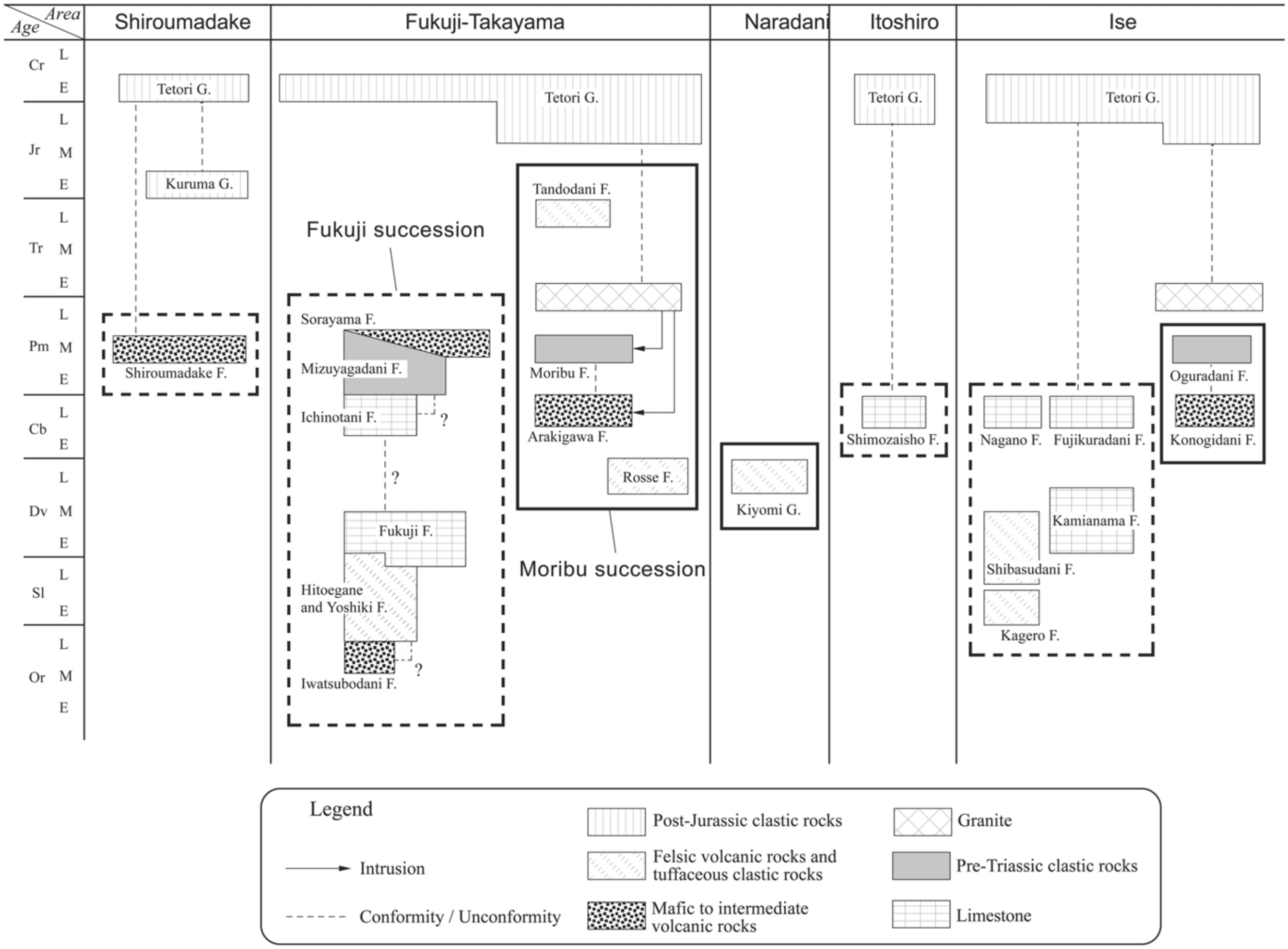

Fig. 2a.14. Stratigraphic relationship among the rocks of the Hida Gaien Belt. Cr, Cretaceous; Jr, Jurassic; Tr, Triassic; Pm, Permian; Cb, Carboniferous; Dv, Devonian; Sl, Silurian; Or, Ordovician; E, Early; M, Middle; L, Late; F., Formation; G., Group.

and middle Palaeozoic oceanic crust (Horikoshi 1972), and as a serpentinite mélange zone caused by ‘Hida Nappe’ emplacement (Chihara & Komatsu 1982; Komatsu *et al.* 1985; Sohma & Kunugiza 1993) onto the Sangun, Akiyoshi, Maizuru and Mino belts. In a more recent study, the belt has been viewed as a tectonic zone caused by Jurassic (dextral) and Cretaceous (sinistral) shearing along the continental margin (Tsukada 2003; Matsumoto 2012).

The Hida Gaien Belt is composed of fault-bounded blocks of Palaeozoic–Mesozoic shelf facies rocks which can be divided into Moribu and Fukuji types based on their differing lithostratigraphy (Figs 2a.14 & 2a.15): (1) the Moribu succession mainly comprises Upper Devonian felsic tuffaceous rocks, Carboniferous volcanic rocks and Middle Permian clastic rocks in ascending order; and (2) the Fukuji succession comprises mainly Ordovician (?) mafic volcanic rocks, Ordovician–Devonian (?) felsic tuffaceous rocks, Lower–Middle (?) Devonian limestone, Carboniferous limestone and Lower–Middle Permian clastic and pyroclastic rocks in ascending order. Whereas both the Moribu and Fukuji successions record similar Devonian palaeoenvironments, they diverged in Carboniferous times when the Moribu rocks record abundant volcanism. In contrast, coeval Fukuji rocks reflect quiet tropical lagoonal conditions until Permian times when volcanism suddenly began, whereas in the Moribu succession volcanism was replaced by clastic sediment deposition (Fig. 2a.15).

The rocks exposed at the Fukuji–Takayama area, which is the type locality of this belt (Tsukada *et al.* 2004), are briefly described in this section and we examine a model that might explain its tectonic development.

Moribu stratigraphy

Devonian–Triassic rocks of the Moribu succession include the Upper Devonian Rosse (mainly felsic tuffaceous rocks), Carboniferous Arakigawa (mafic volcanic rocks with felsic tuffaceous rocks and limestone), Middle Permian Moribu (clastic rocks with felsic tuffaceous rocks and limestone) and Upper Triassic Tandodani (mainly felsic tuffaceous rocks) formations (Fig. 2a.14).

The Rosse Formation, which is in fault contact with the other formations, contains Upper Devonian fossils such as brachiopods, crinoids and plants, with a Lower Devonian limestone block yielding corals (e.g. Tazawa *et al.* 1997, 2000*b*). The Arakigawa Formation yields Upper Visean–Late Kasimovian (or Gzhelian?) fossils such as corals, goniatites, foraminifers, fusulinoideans, brachiopods and trilobites (Isomi & Nozawa 1957; Fujimoto *et al.* 1962; Igo 1964; Kobayashi & Hamada 1987; Tazawa *et al.* 2000*a*). The Moribu Formation, unconformably overlying the Arakigawa Formation, yields Middle Permian brachiopods and fusulinoideans with recycled Lower Permian fusulinoideans in its lowermost horizon

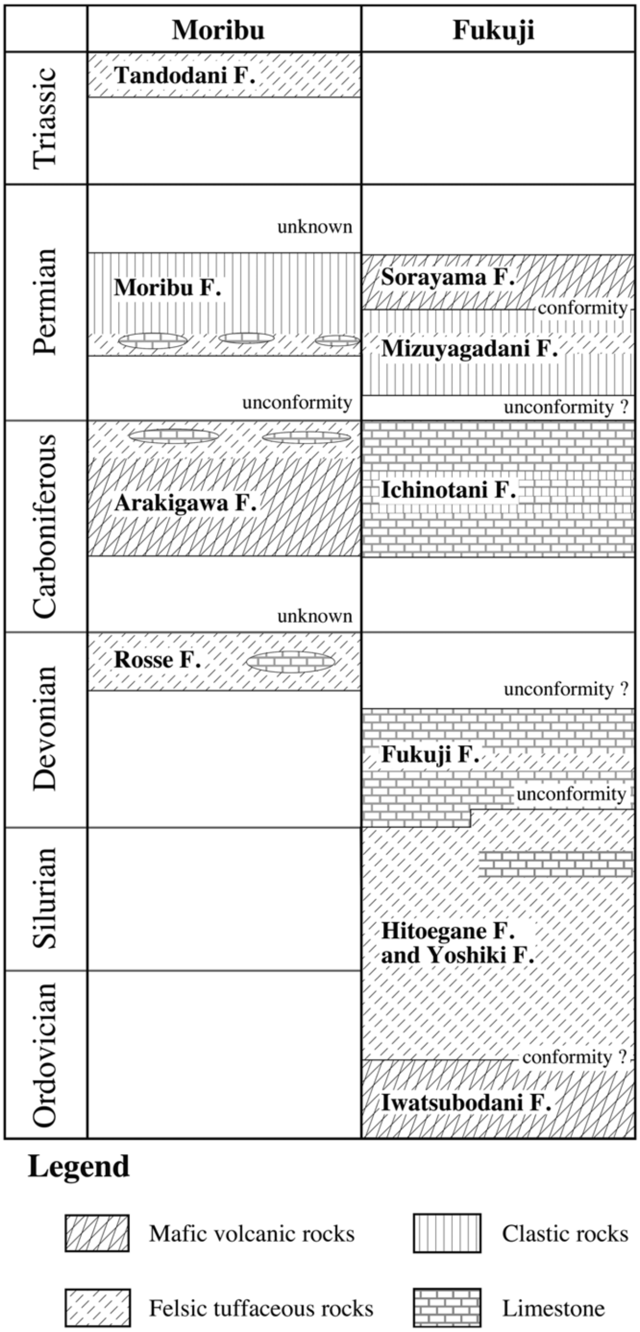

Fig. 2a.15. A columnar section of the Moribu and Fukuji successions showing their major lithologies and characteristic layers and blocks. F., Formation.

(Fujimoto *et al.* 1962; Horikoshi *et al.* 1987; Tazawa *et al.* 1993; Niwa *et al.* 2004). The Tandodani Formation yielding Carnian–Norian conodonts is in fault contact with the Arakigawa Formation (Tsukada *et al.* 1997; Tsukada & Niwa 2005; Fig. 2a.15).

The Moribu Formation is partly metamorphosed into biotite hornfels by intrusion of the 250–240 Ma granite (Funatsu Granite), which is overlain unconformably by Middle Jurassic beds of the Tetori Group (Fig. 2a.14). All pre-Cretaceous rocks are unconformably overlain by uppermost Cretaceous volcanic rocks (Fig. 2a.16).

The equivalent rocks are exposed as the Carboniferous Konogidani and Permian Oguradani formations in the Ise area and as the Devonian Kiyomi Group in the Naradani area (e.g. Tsukada *et al.* 2004; Figs 2a.13 & 2a.14).

Fukuji stratigraphy

The Fukuji succession includes a number of formations in ascending order: (1) Ordovician (?) Iwatsubodani Formation (mafic volcanic rocks); (2) Ordovician–Devonian (?) Hitoegane Formation (felsic tuffaceous rocks); (3) Upper Silurian Yoshiki Formation (felsic tuffaceous rocks); (4) Lower–Middle (?) Devonian Fukuji Formation (mainly limestone); (5) Carboniferous Ichinotani Formation (mainly limestone); (6) Lower–Middle Permian Mizuyagadani Formation (clastic rocks with tuffaceous rocks); and (7) Middle Permian Sorayama Formation (mainly mafic volcanic rocks) (e.g. Igo 1990; Tsukada & Takahashi 2000; Kurihara 2004; Manchuk *et al.* 2013*a*; Fig. 2a.15).

The Iwatsubodani Formation is completely made up of basaltic rocks, whereas the overlying Hitoegane Formation consists mostly of felsic tuffaceous rocks without any sign of basaltic volcanism except in its lower horizons (Tsukada 1997). The lithological contrast between these formations indicates a rapid change from mafic to felsic magmatism in Ordovician times. The Hitoegane Formation ranges from Middle or Late Ordovician to probably Early Devonian age based on fossil records of conodonts, trilobites, plants and radiolarians (Kobayashi & Hamada 1987; Igo 1990; Tazawa & Kaneko 1991; Tsukada & Koike 1997; Kurihara 2007).

The Yoshiki Formation, yielding uppermost Silurian radiolarians and coeval zircons with SHRIMP concordant ages of *c.* 420 Ma (Kurihara 2007; Manchuk *et al.* 2013*a*), can be regarded as an equivalent of the Upper Member of the Hitoegane Formation based on its lithological and palaeontological similarities (Manchuk *et al.* 2013*b*). The Yoshiki Formation is largely in fault contact with the surrounding geological units, although it is partly overlain by the Fukuji Formation (Igo 1990). The Fukuji Formation consists mainly of limestone with corals and stromatoporoids as the main allochemical constituents, along with minor felsic tuff layers (Kamei 1962; Tsukada 2005). Early–Middle (?) Devonian fossils, including corals, conodonts, trilobites and brachiopods, have been obtained from this formation (e.g. Kamei 1952, 1955*b*; Kamei & Igo 1955; Kobayashi & Igo 1956; Hamada 1959*a*, *b*; Koizumi & Kakegawa 1970; Research Group for the Palaeozoic of Fukuji 1973; Okazaki 1974; Ohno 1977; Niikawa 1980; Kuwano 1986, 1987). The Ichinotani Formation, which possibly overlies the Fukuji Formation (Tsukada *et al.* 1999), has yielded many Visean–Gzhelian fossils such as fusulinoideans and smaller foraminifers that indicate an origin in a tropical to subtropical environment (Igo 1956; Kato 1959; Igo & Adachi 1981; Adachi 1985). Remarkable red shale layers in this formation suggest that a pale soil formed by chemical weathering of limestone was deposited in a freshwater lake (Igo 1960). The Mizuyagadani Formation comprises mainly clastic rocks with felsic tuffaceous layers in its lower horizons, and yields Lower Permian corals and Middle Permian radiolarians (Igo 1959; Niko *et al.* 1987; Umeda & Ezaki 1997). The Sorayama Formation yielding autochthonous Middle Permian fusulinoidean fauna conformably overlies the Mizuyagadani Formation (Tsukada *et al.* 1999; Tsukada & Takahashi 2000).

The equivalent rocks of this succession are exposed as the Permian Shiroumadake Formation in the Shiroumadake area, as the Carboniferous Shimozaisho Formation in the Itoshiro area, and as the Silurian Kagero, Silurian–Devonian Shibasudani, Devonian Kamianama and Carboniferous Nagano and Fujikuradani formations in the Ise area (e.g. Tsukada *et al.* 2004; Figs 2a.13 & 2a.14).

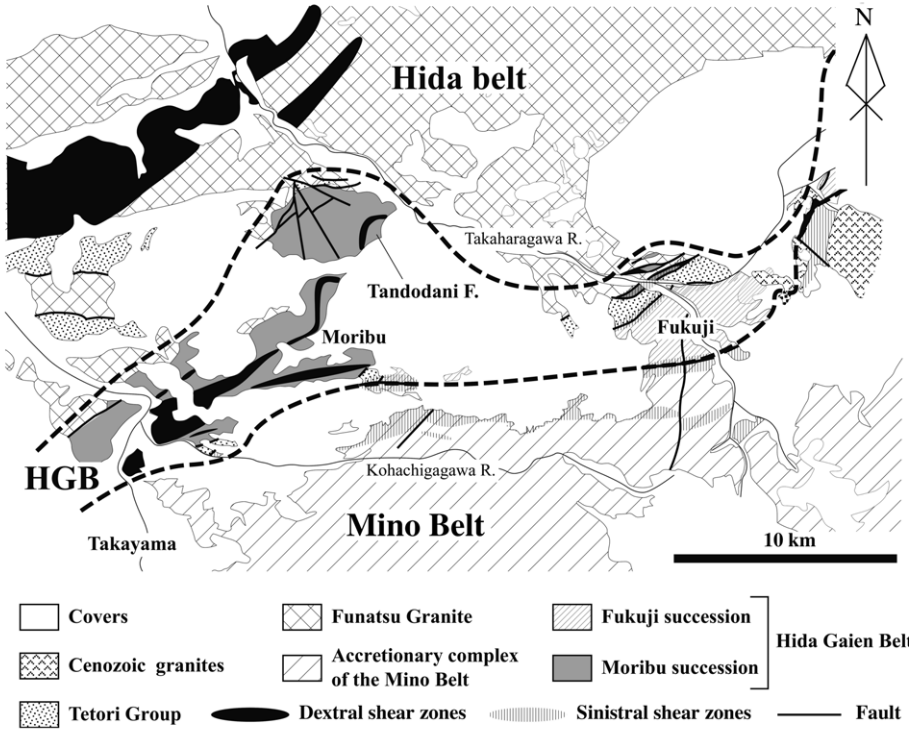

Fig. 2a.16. Simplified geological map of the Fukuji–Takayama area. F., Formation.

Tectonic history of the Hida Gaien Belt

The fault-bounded blocks of the Moribu and Fukuji successions, along with Lower Cretaceous beds belonging to the Middle Jurassic–Lower Cretaceous Tetori Group, are distributed along a narrow zone between the Hida and Mino belts in the Fukuji–Takayama area. In this area, the 250–240 Ma granite intrudes the Moribu succession (Figs 2a.14 & 2a.16) as well as the Hida Gneiss Complex, demonstrating that the Moribu rocks must have been juxtaposed against the Hida Belt by Triassic times (Fig. 2a.17). The Middle to Upper Jurassic Tetori Group overlies the rocks of the Fukuji succession and the Hida Belt, therefore the timing of the juxtaposition of the Fukuji succession against the Hida Belt needs to be before Middle–Late Jurassic times (Fig. 2a.17).

Dextral ductile shear zones, correlated with the Honam Shear Zone in Korea, cut the Carboniferous (?)–Triassic granite to form mylonitic augen gneisses (Yanai *et al.* 1985; Tsukada 2003; Fig. 2a.16). Clasts of similar augen gneisses occur in Middle Jurassic beds within the Tetori Group, which unconformably overlies rocks of the Hida Belt. The timing of deformation has been further constrained by a 191 Ma zircon age on a post-dextral shearing intrusion (Takahashi *et al.* 2010). Taken together, these facts suggest that the shearing lasted until Late Triassic times (timing of deposition of the Tandodani Formation) or later and ended by 191 Ma (Fig. 2a.17). These Mesozoic dextral movements, produced during northwards drift of the continental blocks of East Asia (Otoh *et al.* 1999), were responsible for the fragmentation and dispersal of the Moribu succession and produced the essentially fault-collage disrupted structure of the Hida Gaien Belt, that is, formation of the 'proto-Hida Gaien Belt' (Fig. 2a.17). The equivalent rocks of the Sangun–Renge Belt are in contact with the Moribu rocks across a dextral shear zone in the Ise area (Otoh *et al.* 1999). This fact may suggest that the juxtaposition of the rocks of the Sangun–Renge–Akiyoshi–Maizuru belts against those of the Moribu and Fukuji is attributable to dextral shearing.

A younger deformation phase in the area is recorded by sinistral brittle shear zones which cross the southern part of the Hida Gaien Belt and the northern part of the Mino Belt. One of these sinistral shear zones forms the boundary between the Lower Cretaceous beds of the Tetori Group and rocks of the Fukuji succession (Fig. 2a.15), and is unconformably covered by undeformed Maastrichtian volcanic rocks (Kasahara 1979; Tsukada 2003). The sheared rocks are intruded by undeformed granitoids dated as *c.* 64 Ma in age (Harayama 1990), so it can be deduced that sinistral shearing lasted until Early Cretaceous times or later but had finished by *c.* 64 Ma (Fig. 2a.17). This phase of sinistral shearing was presumably linked to that registered along the eastern margin of Asia in Cretaceous times (e.g. Ozawa 1987; Tashiro 1994; Otoh & Yanai 1996). It restructured the 'proto-Hida Gaien Belt' during sinistral oblique collision with the Mino Belt to complete the present features of the Hida Gaien Belt (Fig. 2a.16). The rocks of the Sangun–Renge–Akiyoshi–Maizuru–Ultra-Tanba belts which had been widely distributed between the proto-Hida Gaien Belt and the Mino Belt might have been crushed by the sinistral movement, to be narrowly scattered in the outer margin of the Hida Gaien Belt (Umeda *et al.* 1996; Otoh *et al.* 1999; Niwa *et al.* 2002; Tsukada *et al.* 2004).

South Kitakami and Kurosegawa belts (ME)

The South Kitakami Belt of NE Japan occupies the southern half of the Kitakami Massif and the eastern marginal part of the Abukuma Massif (Fig. 2a.18), and is composed of older basement rocks with a cover of shallow-marine Ordovician–Lower Cretaceous strata. The cover sequence shows a basically coherent lithostratigraphy, although affected by faulting and folding and containing several stratigraphic breaks represented by unconformities. The belt is bounded to the west by a left-lateral strike-slip fault (the Hatakawa Tectonic Line) which separates it from the Abukuma Belt, which

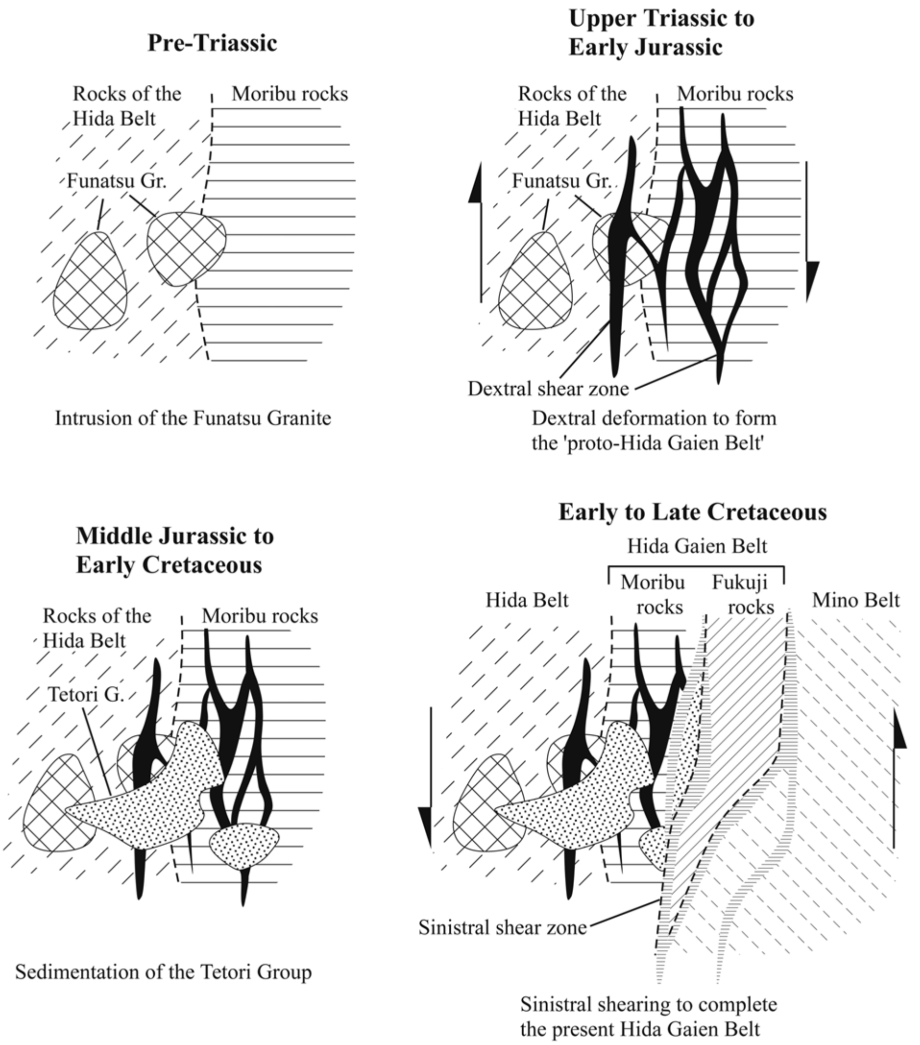

Fig. 2a.17. Schematic model for tectonic development of the Hida Gaien Belt.

comprises metamorphosed Jurassic accretionary complexes and Early Cretaceous (130–110 Ma, 100–90 Ma) granitic rocks (Fig. 2a.18). To the north it is in fault contact with Late Palaeozoic (Nedamo Belt) and Jurassic (North Kitakami Belt) accretionary complexes. The southernmost part of the Abukuma Massif, where the so-called 'Hitachi Palaeozoic' metamorphic rocks crop out (see Chapter 2b), is probably a southern extension of the South Kitakami Belt. The lower part of this Hitachi sequence is considered to be Cambrian in age (Tagiri *et al.* 2010, 2011), whereas the lithofacies and ages of the upper part are similar to the Upper Palaeozoic rocks of the South Kitakami Belt.

The Kurosegawa Belt in the Outer Zone of SW Japan is a southern extension of the South Kitakami Belt. It occupies an east–west-trending narrow zone from Kanto to Kyushu, and is in fault contact on both sides with Jurassic accretionary complexes. There are differing opinions on the tectonic division of SW Japan and tectonic setting of the Kurosegawa Belt (e.g. Matsuoka *et al.* 1998; Yao 2000; Yamakita & Otoh 2000; Isozaki *et al.* 2010). The Kurosegawa Belt includes the Kurosegawa Tectonic Zone (Ichikawa *et al.* 1956), a Late Palaeozoic accretionary complex and its metamorphic equivalent, and well-bedded clastic sequences of Late Palaeozoic, Triassic and Jurassic ages (Yoshikura *et al.* 1990; Hada *et al.* 2000). The Kurosegawa Tectonic Zone is characterized by the pre-Silurian basement, Siluro-Devonian covering strata, and serpentinite and tectonic blocks included therein. The Late Triassic shallow-marine strata unconformably rest not only on the Permian shallow-marine strata, but also on Late Palaeozoic accretionary complexes. The Kurosegawa rocks are well known in Kyushu and Shikoku (especially central Shikoku).

Basement rocks and Early–Middle Palaeozoic cover in the South Kitakami Belt

In the South Kitakami Belt the basement and Middle Palaeozoic strata are distributed in three separate areas, each showing different lithostratigraphic successions: the Nagasaka–Soma district in the west; the Miyamori–Hayachine–Kamaishi district in the north; and the Hikoroichi–Setamai district in the eastern-central part of the belt (Fig. 2a.19).

Nagasaka–Soma district

The Soma district in the eastern margin of the Abukuma Massif is isolated *c.* 150 km to the south of the Nagasaka district of the Southern Kitakami Massif, due to Early Cretaceous left-lateral strike-slip faulting (Otsuki & Ehiro 1992). The basement exposed in the Nagasaka-Soma district comprises the Matsugadaira–Motai Metamorphic Complex and the Shoboji Diorite (Fig. 2a.20).

The Matsugadaira–Motai Metamorphic Complex is of high-pressure/low-temperature type, containing alkali-amphiboles and pumpellyite (Kanisawa 1964; Maekawa 1988), and individual

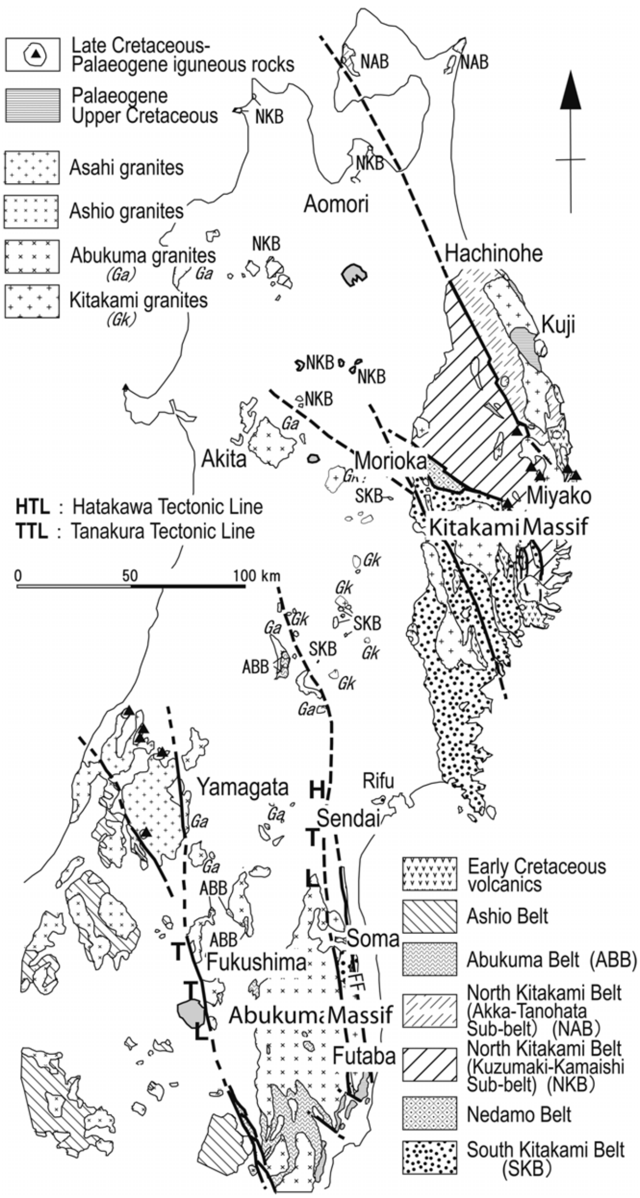

Fig. 2a.18. Geotectonic division map of the pre-Neogene of NE Japan.

metamorphic units have been given various local names such as Motai and Unoki in the Nagasaka district and Matsugadaira, Suketsune, Yamagami and Wariyama in the Soma district (Figs 2a.18 & 2a.19). These metamorphic outcrops comprise amphibolites (basalt and gabbro origin), greenschist, pelitic schist and serpentinized ultrabasic rocks, with subordinate amounts of siliceous and psammitic schists. The chemical composition of the basaltic rocks is similar to that of MORB (mid-oceanic ridge basalts) (Tanaka 1975; Kawabe *et al.* 1979) and the siliceous schist is considered to be derived from chert. The metamorphic lithologies are a mixture of rocks of both continental and oceanic origin and some parts show block-in-matrix mélange structure (Maekawa 1981). The metamorphics are therefore considered to be of accretionary complex origin (Umemura & Hara 1985; Ehiro & Okami 1991). The Matsugadaira Unit is unconformably overlain by the Upper Devonian Ainosawa Formation (Ehiro & Okami 1990). A pre-Late Devonian age is also indicated in the Nagasaka district by the common occurrence of metamorphic clasts in the Upper Devonian Tobigamori Formation (Ehiro & Okami 1991). Sasaki *et al.* (1997) reported an unconformity outcrop between Motai ultrabasic rocks and covering middle Palaeozoic sediments of the Tobigamori Formation (see below) at Natsuyama. Finally, K–Ar hornblende metamorphic ages of *c.* 500 Ma have been obtained from banded amphibolites (Kanisawa *et al.* 1992), indicating the basement rocks to be at least Cambrian in age.

The Shoboji Diorite is sporadically distributed as small bodies near and to the east of Shoboji, Nagasaka district, and is in fault contact with Motai metamorphic rocks and the lower part of the Tobigamori Formation. It is composed of nearly massive diorite to gabbro, and not thought to have been overprinted by the Matsugadaira–Motai high-pressure/low-temperature metamorphism. N-MORB normalized incompatible trace element patterns show that these plutonic rocks were derived from subduction zone magma (Kanisawa & Ehiro 1997), with an intrusion age considered to be latest Ordovician (K–Ar hornblende ages: *c.* 440 Ma; Kanisawa & Ehiro 1997) or latest Cambrian (U–Pb zircon age: 493 Ma; Isozaki *et al.* 2015).

The overlying Middle Palaeozoic strata comprise the Upper Devonian (to lowermost Carboniferous) Tobigamori (Nagasaka district) and Ainosawa (Soma district) formations, meaning that Silurian–Middle Devonian rocks are missing. The Tobigamori Formation consists mainly of mudstone and thin alternating beds of sandstone and mudstone, with purple-coloured tuff and tuff breccia interbedded in the middle part. Metabasite clasts similar to the Motai amphibolites, along with minor amounts of schist clasts, are abundant in the breccia (Fig. 2a.21). The Ainosawa Formation consists of tuffaceous mudstone and mudstone above a basal pebbly sandstone that contains schist and granitic clasts and unconformably covers the underlying metamorphic Matsugadaira Unit (Ehiro & Okami 1990).

Yabe & Noda (1933) reported the Late Devonian (Famennian) brachiopod *Spirifer verneuili* (*Cyrtospirifer tobigamoriensis*; Noda & Tachibana 1959) from the Tobigamori Formation, these being the first Devonian fossils discovered in Japan. The middle part of the formation yielded the land plant *Leptophloeum rhombicum* (Tachibana 1950), whereas in the uppermost part Ehiro & Takaizumi (1992) reported two Late Devonian (and an earliest Carboniferous) ammonoids. The main part of the formation is therefore assumed to be Late Devonian in age, ranging up to earliest Carboniferous. The Ainosawa Formation has also yielded *Cyrtospirifer* (Hayasaka & Minato 1954) and *Leptophloeum* (Koseki & Hamada 1988; Tazawa *et al.* 2006).

Miyamori–Hayachine–Kamaishi district

In this district the Hayachine Complex (Ehiro *et al.* 1988) and its equivalents are widely distributed and comprise the Nakadake Serpentinite, Kagura Unit and Koguro Formation in the eastern part of the Mt Hayachine district (Fig. 2a.20), with similar rocks also cropping out in the Kamaishi and Miyamori districts (Fig. 2a.19). The Nakadake Serpentinite consists of serpentinized peridotite, pyroxenite and hornblendite with subordinate amount of gabbro. The lower part of the Kagura Unit consists of schistose gabbro, peridotite and small amounts of dolerite, whereas the upper part is composed of a dolerite and trondhjemite sheeted dyke complex (Fig. 2a.22) sometimes including gabbro. The Koguro Formation consists mainly of dolerite and basalt, with intercalated thin beds of tuffaceous sandstone and mudstone, and reddish hematite-quartz rock in the upper part. The Hayachine Complex was once inferred to have been formed in a rift zone (Osawa 1983; Ehiro *et al.* 1988) but N-MORB normalized patterns indicate island arc basalt (Mori *et al.* 1992), an origin also inferred by the presence of abundant hornblende in the mafic and ultramafic rocks (Ozawa 1984). Since the overlying Palaeozoic strata yield Silurian fossils, the Hayachine

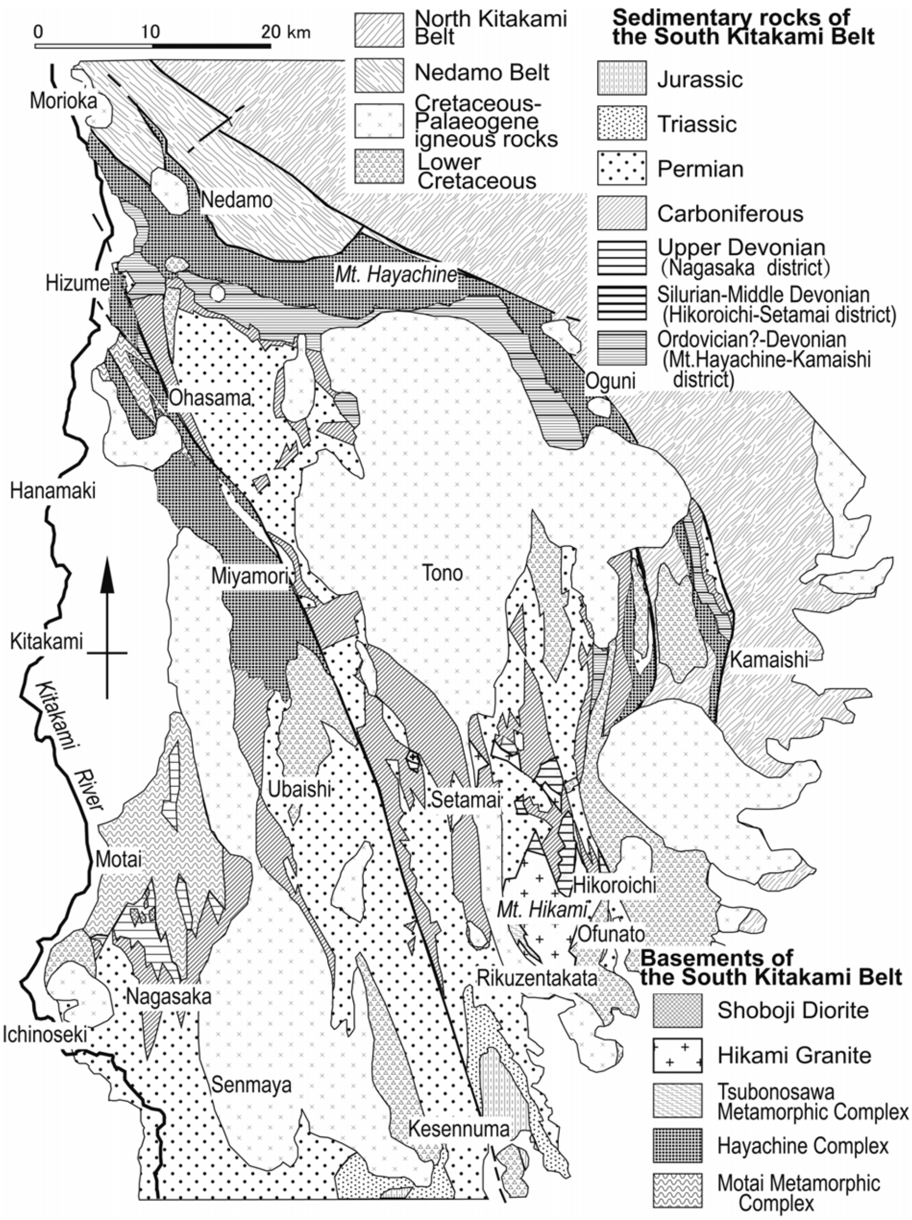

Fig. 2a.19. Geological map of the South Kitakami Belt in the northern half of the Southern Kitakami Massif, NE Japan.

Complex has been considered to be of Ordovician age (Okami *et al.* 1986; Ehiro *et al.* 1988). K–Ar ages of the complex show a wide range of 244–484 Ma (Ozawa *et al.* 1988; Shibata & Ozawa 1992), probably due to the thermal effect of the Early Cretaceous granites. Shimojo *et al.* (2010) reported Middle and Late Ordovician zircon LA-ICP-MS U–Pb ages from the upper part of the complex (466 $\pm$ 5 Ma) and the lowermost part of the overlying Yakushigawa Formation (457 $\pm$ 10 Ma).

The Lower–Middle Palaeozoic cover strata in the eastern Mt Hayachine district comprise the Yakushigawa and Odagoe formations in ascending order. The Yakushigawa Formation conformably overlies the Koguro Formation and consists of mudstone and sandstone in association with tuff and basalt in the lower part. The Odagoe Formation conformably covers the Yakushigawa Formation and is composed of mudstone and siliceous mudstone with impure limestone in the main part and basalt, tuff and limestone in the upper part. A Silurian brachiopod was reported from the lower part (Ehiro *et al.* 1986). In the western Mt Hayachine district, the cover strata comprise the fault-bounded Nameirizawa and Orikabetoge formations and undivided Devonian (in ascending order). The Nameirizawa Formation is lithologically similar to the Yakushigawa Formation, whereas the Orikabetoge Formation is composed of conglomerate, arkose and mudstone with subordinate amounts of limestone and tuff, and a rich fauna of Silurian corals and trilobites (Okami *et al.* 1986). The conglomerate contains many granitic clasts lithologically similar to those exposed in the Hikami granitic complex of the Hikoroichi–Setamai district (see following section). Finally, undivided Devonian strata in this district comprise tuff, mudstone, sandstone and conglomerate.

In the Kamaishi district the uppermost Silurian–Upper Devonian Senjyogataki Formation rests on the Kentosan Unit (equivalent of the Kagura Unit) of the Hayachine Complex, although their stratigraphic relationship is unclear. The Senjyogataki Formation consists, from base to top, of basalt and dolerite, tuff and alternating beds of tuffaceous and siliceous mudstones, and mudstone (Okami *et al.* 1987). The tuffaceous-siliceous mudstone yielded latest Silurian

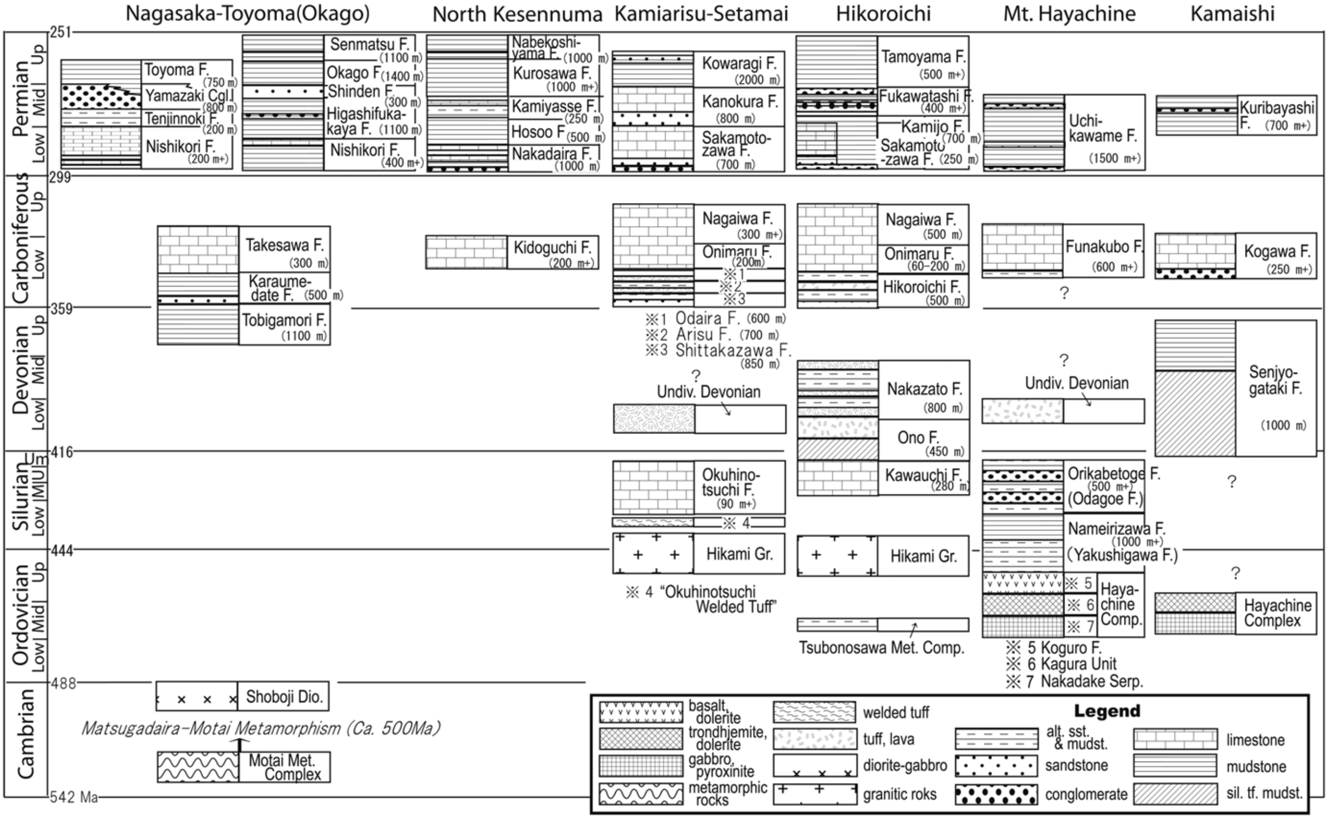

Fig. 2a.20. Stratigraphy of the Palaeozoic of the South Kitakami Belt, NE Japan. The Thicknesses of the strata are given in parentheses. alt., alternating; Comp., Complex; Dio., Diorite; F., Formation; GR., Granite; mudst., mudstone; Met., Metamorphic; Serp., Serpentinite; sil., siliceous; sst., sandstone; tf., tuff.

(Pridolian) (Suzuki *et al.* 1996) to Early Devonian radiolarians (Umeda 1998*a*), and the uppermost part of the formation contains the Late Devonian (Famennian) plant *Leptophloeum* (Okami *et al.* 1987).

Hikoroichi–Setamai district

The Hikami Granite (Murata *et al.* 1974) and Tsubonosawa Metamorphic Complex (Ishii *et al.* 1956) form the basement in this district. The largest body of the Hikami Granite, which consists

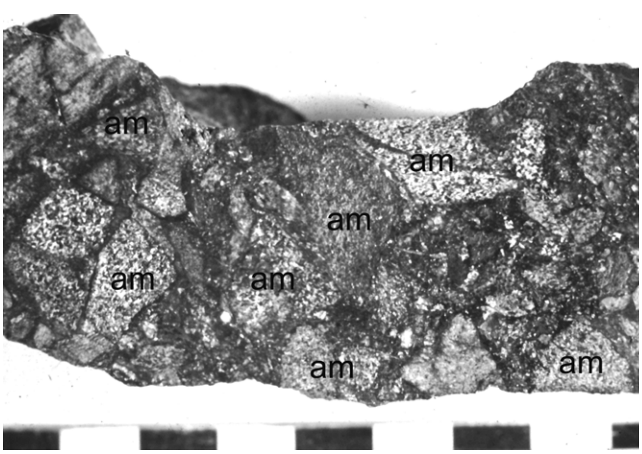

Fig. 2a.21. Polished specimen of tuff breccia from the middle part of the Upper Devonian Tobigamori Formation, showing the common occurrence of amphibolite clasts (am).

of massive or schistose granite, granodiorite and tonalite, crops out around Mt Hikami (Fig. 2a.19), with other bodies forming rather small and scattered outcrops in the Komatsu-Pass, Shiraishi-Pass, Yokamachi, Ezo, Hirasawa and Okuhinotsuchi areas. All of these exposures, except for that at Hirasawa, are associated with Middle Palaeozoic sedimentary rocks of Silurian–Middle Devonian in age.

Ishii *et al.* (1956) divided the Hikami outcrop into 'Hikamisan-type' and 'Ono-type', while Asakawa *et al.* (1999) later divided it into 'D-type', most of which show schistosity, and massive 'S-type', with roughly coincides with the Ono-type. Kobayashi & Takagi (2000) considered that the massive Ono-type granodiorite intruded after the schistose granites had been mylonitized. The radiometric ages obtained from the Hikami Granite are 442 Ma (Watanabe *et al.* 1995; SHRIMP U–PB age) and 440 Ma (Asakawa *et al.* 1999; Rb–Sr whole-rock isochron age). Some younger ages (Devonian–Permian) had been reported (e.g. CHIME ages: Suzuki & Adachi 1991; Suzuki *et al.* 1992; Adachi *et al.* 1994; LA-ICP-MS U–Pb age: Shimojo *et al.* 2010) but these are in conflict with the fact that the Hikami plutons are unconformably overlain by Silurian strata (see the following paragraph). Moreover, a granitic clast included in the lower part of the Ono Formation (Pridolian) has been dated at *c.* 250 Ma by Suzuki *et al.* (1992), so it seems likely that the post-Silurian ages obtained from the Hikami Granite may be related to later thermal events. Finally, the Tsubonosawa Metamorphic Complex comprises contact metamorphic lithologies derived from mudstone and sandstone protoliths of probable Cambro-Ordovician age, and included in the Hikami Granite (Hikamisan-type) as xenoblocks of various sizes. Detrital zircon grains from these metasediments have ages in the range

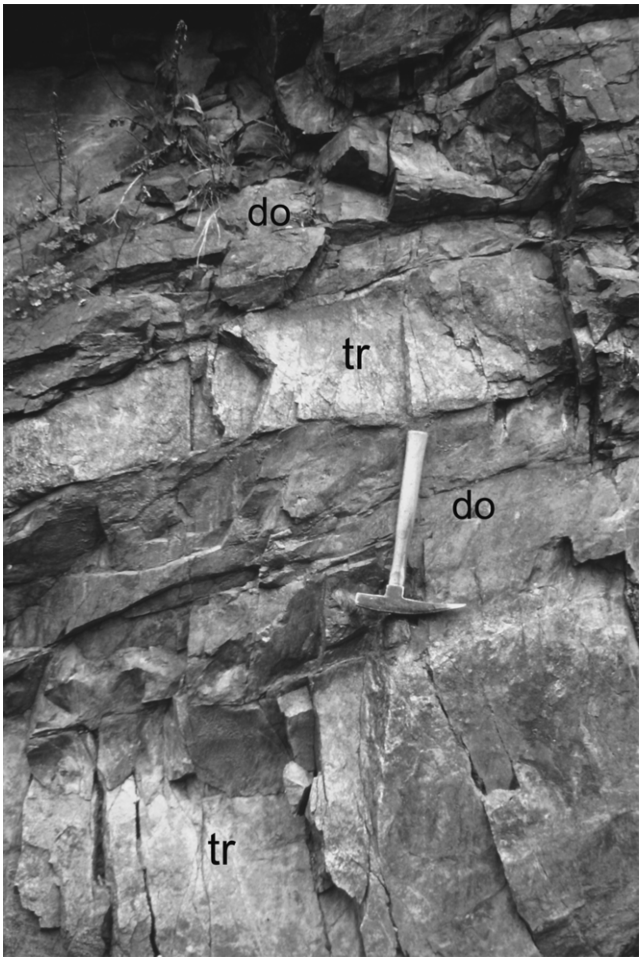

Fig. 2a.22. Photograph of an outcrop showing the sheeted dyke complex of the Kagura Unit of the Hayachine Complex. tr, trondhjemite; do, dolerite.

3800–500 Ma and euhedral zircon grains yield ages of *c.* 440 Ma (CHIME ages: Suzuki & Adachi 1991; SHRIMP U–Pb ages: Watanabe *et al.* 1995).

The Middle Palaeozoic strata in this district comprise the Silurian Kawauchi Formation (and its equivalents), the latest Silurian–Early Devonian Ono Formation and the Middle Devonian Nakazato Formation; Upper Devonian rocks are absent (Fig. 2a.20). Onuki (1937) first reported Silurian fossils from this district, this being the first record of Silurian fossils in Japan. The Kawauchi Formation consists mainly of limestone with basal arkose, and is associated with calcareous mudstone in the upper part. It has yielded many corals (*Schedohalysites*, *Falsicatenipora*, *Favosites*, *Heliolites*, etc.) and trilobites (*Encrinurus*), and is considered to be Middle–Late Silurian in age (Kato *et al.* 1980). The basal arkose unconformably covers the Hikami plutonic basement (Murata *et al.* 1974, 1982; Kitakami Paleozoic Research Group 1982). In the Okuhinotsuchi district, the Early–Late Silurian Okuhinotsuchi Formation (Kawamura 1980) rests unconformably on a welded tuff (the Okuhinotsuchi Welded Tuff; Murata *et al.* 1982), which in turn unconformably covers the Hikami Granite.

The Ono Formation is divided into the Oh1, Oh2 and Oh3 members (Minato *et al.* 1979). The Oh1 Member, which has yielded Pridolian radiolarians (Umeda 1996*b*), is a slump bed which incorporates variously sized clasts of granite, arkose and limestone in a tuffaceous and siliceous mudstone. These clasts are lithologically similar to the underlying Hikami Granite and basal arkose and limestone of the Kawauchi Formation, respectively. The Oh2 Member is composed of acidic tuff and alternating beds of acidic tuff and tuffaceous-siliceous mudstone, and the Oh3 Member similarly consists mainly of tuff with subordinate amount of tuffaceous sandstone and mudstone. The Nakazato Formation conformably overlies the Ono Formation and is divided into the N1, N2, N3 and N4 members (Minato *et al.* 1979). The N1 Member consists mainly of basic tuff and lapilli tuff in association with mudstone; the N2 Member is composed of alternating beds of acidic tuff and mudstone; and the N3 Member comprises alternating beds of sandstone and mudstone, a part of which is tuffaceous. The lower part of the N4 Member consists of sandstone and pebbly sandstone, and the upper part of it consists of sandstone and mudstone with subordinate amounts of tuff. The N3 Member has yielded Middle Devonian trilobites (Kobayashi & Hamada 1977; Minato *et al.* 1979) and Middle Devonian (Givetian) radiolarians (Umeda 1996*b*).

Basement rocks and Middle Palaeozoic cover in the Kurosegawa Belt

Basement rocks

The basement rocks of the Kurosegawa Belt in Shikoku are composed of the Terano Metamorphic Complex (including the Miyagadani Metamorphic Complex) and Mitaki Igneous Complex (Ichikawa *et al.* 1956). The Terano Metamorphic Complex comprises medium-pressure type (Karakida 1981) biotite gneiss and amphibolites with radiometric ages ranging from Ordovician to Jurassic (mostly clustering around 400 Ma; Yoshikura *et al.* 1990), and is in fault contact with the Mitaki Igneous Complex. U–Pb ages of detrital zircons from the Terano Metamorphic Complex show peaks at 450–500 and *c.* 600 Ma (Yoshimoto *et al.* 2013). The garnet-clinopyroxene granulite and amphibolite of the Kurosegawa Belt in Kyushu District have Sm–Nd ages of *c.* 420, and 490 and 540 Ma, respectively (Osanai *et al.* 2000). The Mitaki lithologies – which include the Yokokurayama Granite (Yasui 1984) or Gomi Granite (Yoshikura 1985) – are typically medium- to coarse-grained, commonly sheared, calc-alkaline granitic rocks ranging from granite to diorite (mostly granodiorite) with U–Pb zircon ages of *c.* 440–442 Ma (Hada *et al.* 2000).

Middle Palaeozoic covering strata

Middle Palaeozoic strata in central Shikoku (Fig. 2a.23) are represented by the Siluro-Devonian Yokokurayama Group which is

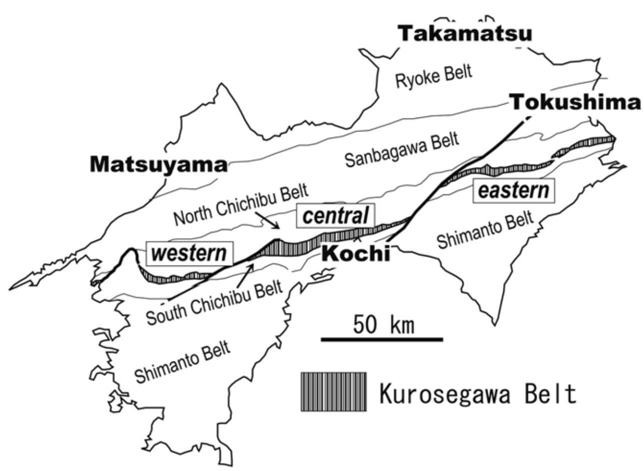

Fig. 2a.23. Map showing the distribution of the Kurosegawa Belt in Shikoku, SW Japan.

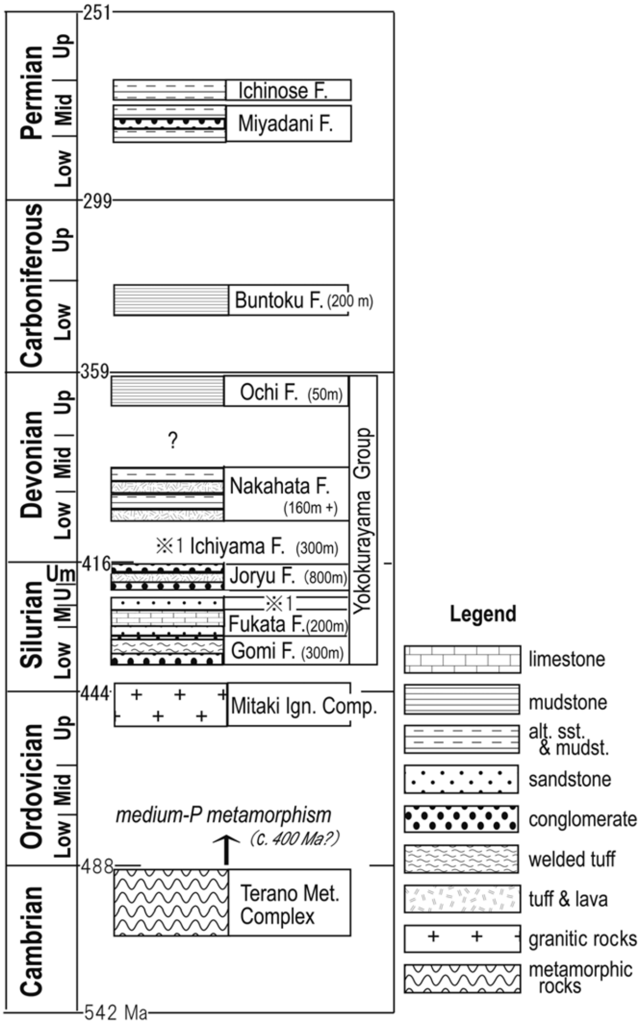

Fig. 2a.24. Stratigraphy of the Palaeozoic of the Kurosegawa Belt in the central part of Shikoku, Southwest Japan. Thicknesses of the strata are given in parentheses. alt., alternating; Comp., Complex; Ign., Igneous; F., Formation; mudst., mudstone; Met., Metamorphic; sst., sandstone.

divided into the Gomi, Fukata, Ichiyama, Joryu, Nakahata and Ochi formations, in ascending order (Umeda 1998*b*; Fig. 2a.24). The stratigraphic relationships among the first three formations are conformable, whereas there are hiatuses between the other formations, based on the ages of radiolarians (Umeda 1998*b*, *c*). The Okanaro Group (Ichikawa *et al.* 1956; Umeda 1994) in western Shikoku, the Suberidani Group (Hirayama *et al.* 1956) in eastern Shikoku and the Gionyama Formation (Hamada 1959*c*) in Kyushu are equivalent strata, but exclude the equivalent of the Ochi Formation.

The lower part of the Gomi Formation consists of basal conglomerate, which unconformably covers the Yokokurayama Granite (Yasui 1984), overlain by conglomeratic sandstone and sandstone. The main part is composed of welded tuff and intercalated acidic tuff, tuffaceous sandstone, sandstone and mudstone (Yoshikura & Sato 1976). A Wenlockian U–Pb SHRIMP age (427 Ma) has been reported from the welded tuff (Aitchison *et al.* 1996). The Fukata Formation (Umeda 1998*b*) is subdivided into a Lower Member, comprising alternating beds of sandstone and mudstone, and mudstone, with calcareous sandstone and conglomerate in its lower and upper parts, and an Upper Member comprising mainly limestone with limestone conglomerate, tuffaceous sandstone and mudstone. This formation yields late Wenlockian–early Ludlovian corals, trilobites (Hamada 1959*c*; Kobayashi & Hamada 1985), conodonts (Niko *et al.* 1989) and radiolarians (Wakamatsu *et al.* 1990). The corals are rich in halysitids such as *Halysites*, *Schedohalysites* and *Falsicatenipora*, and favositids. The Ichiyama Formation consists of conglomerate, tuffaceous sandstone and mudstone, and acidic tuff. Based on the radiolarians, this formation dated as early–middle Ludlovian (Umeda 1998*b*). The Joryu Formation consists of conglomerate in the lower part and tuffaceous sandstone, acidic tuff and conglomerate in the upper part. The acidic tuff contains rich Pridolian radiolarians (Umeda 1998*b*, *c*). The Nakahata Formation is dominated by acidic tuff and tuffaceous sandstone, including basal conglomerate-sandstone with abundant granite, rhyolite and acidic tuff clasts. This formation yields an abundant radiolarian fauna, ranging in age from middle Early Devonian to early Middle Devonian (Umeda 1998*b*, 1998*c*).

The Ochi Formation (Hirata 1966; Yoshikura 1982) mainly comprises alternating beds of sandstone and mudstone, with intercalated conglomerate, sandstone, mudstone and acidic tuff. It is in fault contact with the other formations (Yoshikura 1985), although Yasui (1984) has stated that there is a possibility that this formation unconformably rests on the underlying acidic-tuff-dominated Devonian. The Late Devonian plant *Leptophloeum rhombicum* has been reported (Hirata 1966), and a similar *Leptophloeum*-bearing clastic facies is known from Upper Devonian rocks in eastern Shikoku where conglomerate, sandstone and mudstone cover Bed G4 of the Suberidani Group in probable unconformity (Yasui & Okitsu 2007). In Kyushu, the Naidaijin Formation has also yielded *Leptophloeum* (Miyamoto & Tanimoto 1993; Saito *et al.* 2003).

Upper Palaeozoic strata of the South Kitakami Belt

Carboniferous and (especially) Permian shallow-marine strata are widely distributed in the South Kitakami Belt (Fig. 2a.19).

Carboniferous

The Carboniferous succession ranges in age from Tournaisian to Moscovian, lacks Late Carboniferous rocks, and can be divided into two lithologically distinct lower and upper sequences. The lower (Tournaisian–upper Visean) consists mainly of volcaniclastics with minor amounts of limestone, whereas the upper (upper Visean–Moscovian) is dominated by limestone and partly intercalated with tuff (Kawamura & Kawamura 1989*a*). The volcanics in these strata, as in the underlying Devonian sequence, show bimodal SiO_2 contents and belong to a volcanic-arc to back-arc tectonic setting (Kawamura & Kawamura 1989*b*).

The lower sequence comprises the Hikoroichi Formation (Hikoroichi district), the Shittakasawa, Arisu and Odaira formations (Setamai–Yokota district), the Karaumedate Formation (Nagasaka district), Mano Formation (Soma district) and their equivalents in other areas (Fig. 2a.20). They differ in their lithological component ratio and thickness, and are considered to have been deposited in different basins (Kawamura & Kawamura 1989*a*, *b*). Those in the Hikoroichi and Setamai–Yokota districts are dominated by volcanic rocks and are very thick, whereas those in the Nagasaka–Soma district are mainly composed of clastic rocks with minor amounts of tuff. In contrast, the Mano Formation in the Soma district is very thin. In most districts in the South Kitakami Belt, the Lower Carboniferous strata rest unconformably above Middle Devonian strata. In the Nagasaka–Soma district, however, the Tobigamori Formation ranges from Upper Devonian to lowermost Carboniferous and is covered conformably by the Lower Carboniferous Karaumedate Formation. The Hikoroichi Formation and its equivalents yield a rich fauna of corals, brachiopods and trilobites, and its lower–middle part and upper part are dated as Tournaisian–lower Visean and upper Visean, respectively (Mori & Tazawa 1980; Kawamura 1983; Tazawa 1984).

The upper sequence comprises the Onimaru and Nagaiwa formations in the Hikoroichi, Setamai and Yokota districts, and their equivalents in other areas. The Onimaru Formation is composed of bedded limestone and muddy limestone, and the Nagaiwa Formation consists of massive limestone or limestone with siliceous nodules and sometimes intercalated pyroclastic rocks. The Onimaru Formation is rich in corals belonging to the *Kueichouphyllum* fauna. It is biostratigraphically divided into the *Kueichouphyllum glacial–Actinocyathus japonicus* Zone, *Saccaminopsis* Zone, *Arachnolasma–Palaeosmilia regia* Zone and Barren Zone, and dated as upper Visean (Niikawa 1983*a*, 1983*b*). The Nagaiwa Formation is divided into the *Millerella* and *Profusulinella* zones (fusulinoideans; Kobayashi 1973) or *Declinognathodus noduliferus* and *Idiognathoides sulcatus* zones (conodont; Minato *et al.* 1979), and dated as Serpukhovian–Moscovian.

Permian

The Permian succession in the South Kitakami Belt consists mainly of shallow-marine clastic sediments and limestone, almost entirely lacks pyroclastic rocks (unlike the Devonian and Carboniferous sequences), and has been divided into the Sakamotozawan, Kanokuran and Toyoman series (Fig. 2a.20) in ascending order (e.g. Minato *et al.* 1978, 1979).

The Sakamotozawa Formation (the type of the Sakamotozawan Series) in the Hikoroichi district consists of basal conglomerate, sandstone and alternating beds of sandstone and mudstone in the lower part, massive limestone with mudstone in the main part and sandstone and mudstone in the uppermost part (Kanmera & Mikami 1965*a*). Equivalent strata of the Sakamotozawa Formation in other areas also comprise basal conglomerate or sandstone, sandstone, mudstone and limestone, although the amounts of these different components varies from place to place (Ehiro 1989, 2016). The formation unconformably overlies the Carboniferous Nagaiwa Formation or its equivalents but without any notable structural break, except in the Setamai district where it progressively oversteps various horizons of the Carboniferous from the uppermost Shittakasawa Formation to the Nagaiwa Formation (Saito 1968). The Sakamotozawa Formation and its equivalents yield a rich fauna of fusulinoideans and corals. Kanmera & Mikami (1965*b*) established five fusulinoidean zones – the *Zellia nunosei*, *Monodiexodina langsonensis*, *Pseudofusulina vulgaris*, *P. fusiformis* and *P. ambigua* zones – and dated the formation as Sakmarian–Artinskian. Ueno *et al.* (2009) extended the uppermost part into the Kungurian stage, based on the occurrence of *Misellina* sp. The Sakamotozawan Nishikori Formation in the Toyoma district yielded plant fossils belonging to the Cathaysian flora such as *Cathaysiopteris*, *Sphenopteris*, *Odontopteris*, *Pecopteris*, *Taeniopteris* and *Cordites* (Asama 1956, 1967).

The Kanokura Formation (the type of the Kanokuran Series) in the Setamai district rests conformably on the Sakamotozawa Formation, and comprises a lower sequence of conglomeratic sandstone, sandstone and mudstone overlain by muddy limestone and lenticular limestones. The upper part of this formation consists mainly of massive limestone, but changes laterally to calcareous sandstone or mudstone (Ehiro 1989, 2015). Thick conglomerate beds, called the Usuginu-type Conglomerate, are sometimes intercalated in the Kanokuran Series (especially in the upper part) and the overlying Toyoman Series. The conglomerate contains well-rounded pebbles to boulders rich in granitic clasts in a muddy or sandy matrix. Permian–Triassic radiometric ages have been known from these granitic clasts (K–Ar ages of 276–225 Ma, Shibata 1973; CHIME ages of 257–244 Ma, Takeuchi & Suzuki 2000). Various fossils such as fusulinoideans, corals, brachiopods and molluscs have been reported from the Kanokura Formation (e.g. Minato *et al.* 1978), allowing Choi (1973) to establish three fusulinoidean zones: the *Monodiexodina matsubaishi* Zone (lower part, which has also yielded *Cancellina*), *Colania kotsuboensis* Zone (uppermost part of the lower) and *Lepidolina multiseptata* Zone (upper part). Based on these fusulinoideans, the Kanokura Formation is dated as Roadian–Capitanian.

The Permian in the northern Kesennuma district comprises the Nakadaira, Hosoo, Kamiyasse, Kurosawa and Nabekoshiyama formations in ascending order. The Hosoo Formation is dominated by mudstone, and its upper and uppermost parts yield Roadian and Wordian ammonoids, respectively (Ehiro & Misaki 2005). The Kamiyasse Formation consists of calcareous sandstone and mudstone with limestone, and contains Wordian–Capitanian fusulinoideans and ammonoids (Ehiro & Misaki 2005) with a rich fauna of brachiopods. Finally, the lower part of the Kurosawa Formation, in which mudstone is dominant, contains Capitanian ammonoids (Ehiro & Araki 1997) along with the agassizodontid shark *Helicoprion* sp. (Araki 1980).

The Toyoman Series includes the Toyoma Formation (Toyoma, Ogatsu and southern Kesennuma districts), Okago and Senmatsu formations (Okago district), the main part of the Suenosaki and the Tanoura formations (Utatsu district), the upper part of the Kurosawa and the Nabekoshiyama formations (northern Kesennuma district) and the main part of the Kowaragi Formation (Karakuwa district), as well as various unnamed beds in the Ofunato district and other places. Compared to the Lower–Middle Permian successions of the South Kitakami Belt, these younger formations are lithologically monotonous and dominated by mudstone with minor sandstone and rare limestone and conglomerate. The mudstones of these strata typically show strong slaty cleavage, except for those in the Okago district, and so have been used as roofing slate. The Toyoma and equivalent formations yield Wuchiapingian ammonoids (Murata & Bando 1975; Ehiro & Bando 1985; Ehiro 2010), whereas the Senmatsu, Tanoura and Nabekoshiyama formations, as well as unnamed beds in the Ofunato district, contain Changhsingian fusulinoideans, smaller foraminifers and ammonoids (Ishii *et al.* 1975; Tazawa 1975; Murata & Shimoyama 1979; Ehiro 1996; Kobayashi 2002).

Upper Palaeozoic covering strata in the Kurosegawa Belt

Carboniferous

In the Kurosegawa Belt in Shikoku Carboniferous rocks are represented by the shallow-marine Buntoku Formation (Nakai 1980), fragments of which are dispersed within a fault zone in Yokokurayama district, central Shikoku. It consists mainly of mudstone intercalated with basic tuff containing limestone lenses which have yielded Late Visean corals such as *Lithostrotion, Palaeosmilia, Diphyphyllum* and *Carcinophyllum*. Equivalent strata also crop out in Kyushu (Yuzuruha Formation, Miyamoto & Tanimoto 1993; Kakisako Formation, Kanmera 1952) where they are dominated by limestone with a late Visean *Kueichouphyllum*-coral fauna (Kanmera 1952; Miyamoto & Tanimoto 1993; Kido *et al.* 2007).

Permian

Permian strata in central Shikoku are represented by the shallow-marine Miyadani and Ichinose formations (Hada *et al.* 1992), these having equivalents in western Shikoku as the Miyanaro and Doi formations, respectively. They are all in fault contact with other strata including Jurassic accretionary complexes. The Miyadani and Miyanaro formations mainly consist of sandstone, mudstone and alternating beds of sandstone and mudstone, in association with conglomerate and rare acidic tuff. The latter also includes small limestone lenses with early–middle Middle Permian fusulinoideans (Hada 1974) and siliceous mudstones with late Early Permian radiolarians (Hada *et al.* 1992). The Ichinose and Doi formations are composed of sandstone, conglomerate and mudstone, with subordinate amount of acidic tuff and limestone lenses. Limestone lenses yield the late Middle Permian fusulinoid *Lepidolina*, and radiolarians from these formations are also late Middle Permian in age (Hada *et al.* 1992). Similarly, the middle–late Middle Permian ammonoid *Cibolites* has been reported from the Katsura Sandstone of the Ichinose Formation (Group) (Koizumi *et al.* 1994). The Haigyu Group (Hirayama *et al.* 1956) in eastern Shikoku is also Middle Permian in age, whereas to the west in Kyushu coeval siliciclastic strata comprise the Early Permian Tsurukoba Formation (Miyamoto *et al.* 1997), Middle Permian Kozaki Formation (Kanmera 1961, 1963) and the late Middle–Late Permian Kuma Formation (Kanmera 1953, 1954).

Late Palaeozoic–Mesozoic metamorphic rocks and serpentinite in the Kurosegawa Belt

Various kinds of metamorphic rocks (excluding the Terano Metamorphic Complex) and serpentinites occur in the Kurosegawa Belt, although their relationships with the basements and their covering strata are not clear. The metamorphic rocks were known as the Ino Formation, which consists of low-grade metamorphic rocks and tectonic blocks of high-grade metamorphic rocks in serpentinite and Ino Formation. Wakita *et al.* (2007) referred to the former as the Younger Ino Metamorphic Complex and the latter as the Older Ino Metamorphic Complex.

High-grade metamorphic rocks

The Older Ino Metamorphic Complex occurs as lenticular small bodies in and around the Younger Ino Metamorphic Complex and is composed mainly of mafic schist with minor amount of pelitic schist. It records an epidote-glaucophane schist subfacies to albite-epidote-amphibolite subfacies metamorphism (Wakita *et al.* 2007). The metamorphic age is estimated to be Carboniferous based on the K–Ar ages of phengite from mafic schists (327–317 Ma: Ueda *et al.* 1980; 350 Ma: Wakita *et al.* 2007). High-grade metamorphic rocks in serpentinite consist of basic, pelitic and psammitic schists and show glaucophene schist to pumpellyite-actinolite facies metamorphism (Nakajima & Maruyama 1978); a basic and siliceous schist block preserving jadeite-glaucophene facies metamorphism is also known. K–Ar ages of phengite from the schist are 240 and 208 Ma (Maruyama *et al.* 1978).

Younger Ino Metamorphic Complex

The Younger Ino Metamorphic Complex (Ino Formation) is rather widely distributed in the Kurosegawa Belt of central Shikoku and composed of basic, pelitic, psammitic and siliceous schists. It records pumpellyite-actinolite subfacies metamorphism and K–Ar ages of phengite in the range 185–148 Ma (Wakita *et al.* 2007). The unmetamorphosed part of the complex shows an Oceanic Plate Stratigraphy including the Upper Carboniferous–Upper Triassic bedded chert, Upper Triassic–Lower Jurassic siliceous mudstone and Lower Jurassic mudstone (Hori & Wakita 2004).

Serpentinite

The serpentinite is distributed along the fault zones as sheet-like to lenticular bodies up to several kilometres wide and their protoliths are considered to be dunite and harzburgite with lesser amount of lherzolite, based on relict minerals and pseudomorphic textures (Hada *et al.* 2001). Their chemistry is similar to subduction-related ultramafic rocks from the Mariana and Tonga trenches (Yoshikura & Miyaji 1988).

Triassic–Lower Cretaceous strata in the South Kitakami Belt

Triassic to lowest Cretaceous strata in the South Kitakami Belt were deposited in a shallow marine or alluvial environment and are mainly composed of clastic rocks in association with rare limestone and tuff. They are distributed in the southern part of the Southern Kitakami Massif (Figs 2a.18 & 2a.25) and in the eastern margin of the Abukuma Massif (Soma district), forming three large, south-plunging synclinal structures (Figs 2a.25 & 2a.26): the Shizugawa (Shizukawa)–Hashiura Sub-belt (Western Sub-belt); Karakuwa–Oshika Sub-belt (Central Sub-belt); and the Ofunato Sub-belt (Eastern Sub-belt) (Takizawa 1977, 1985). The stratigraphy, lithology and thickness of the strata are different in the three synclines.

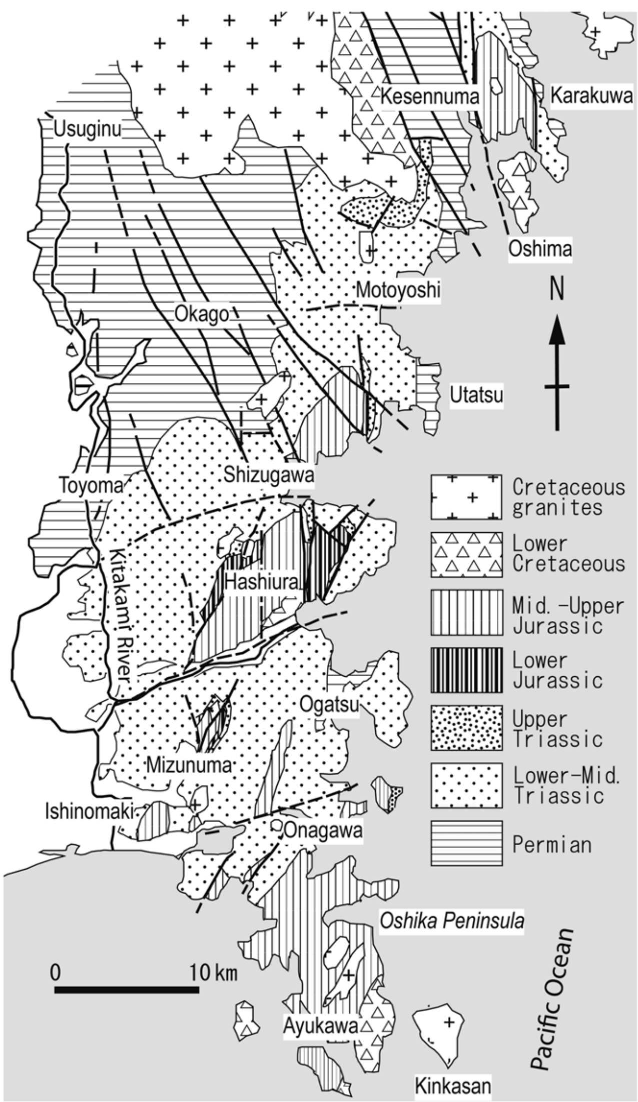

Fig. 2a.25. Geological map of the southern part of the Southern Kitakami Massif, showing the distribution of the Mesozoic strata of the South Kitakami Belt.

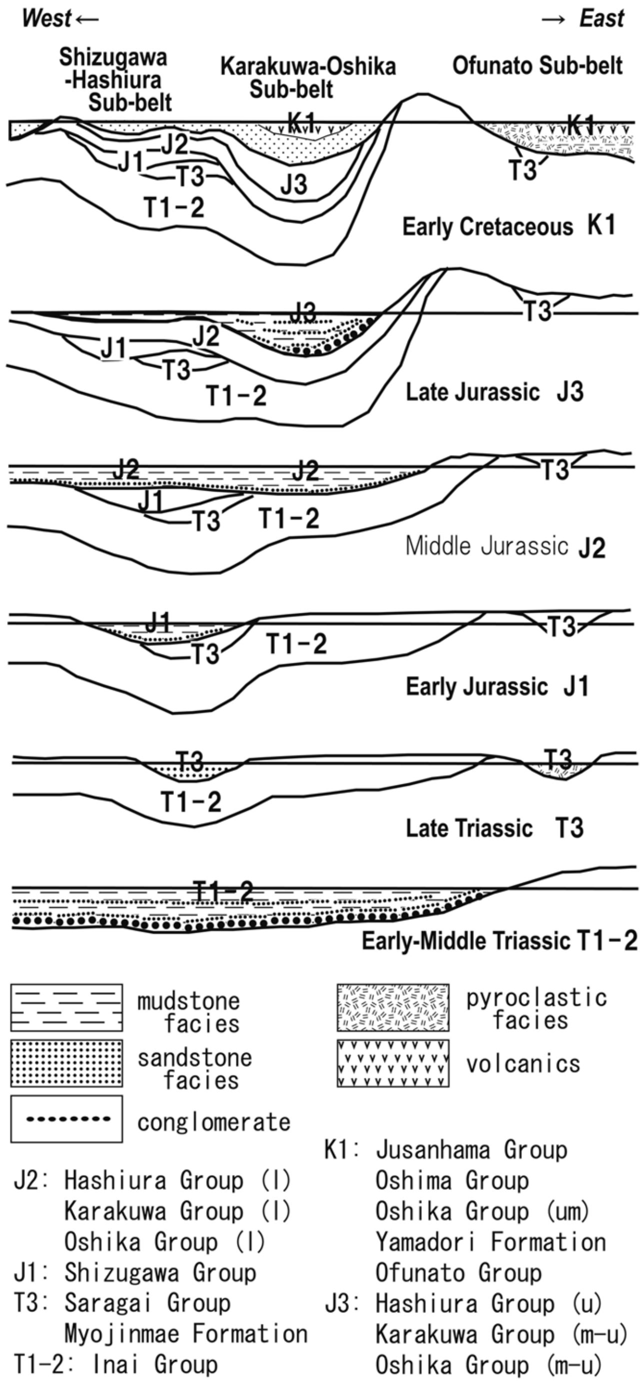

Fig. 2a.26. Tectonic history of the Mesozoic sedimentary basin in the South Kitakami Belt (modified from Yamashita 1957). l, lower part; m, middle part; u, upper part; um, uppermost part.

Lower–Middle Triassic

Lower–Middle Triassic rocks are represented by the Inai Group which crops out in the Shizugawa–Hashiura Sub-belt and Karakuwa–Oshika Sub-belt of the Southern Kitakami Massif and in the Rifu district to the west of the former. The group comprises the Hiraiso, Osawa, Fukkoshi and Isatomae formations in conformable ascending order (Onuki & Bando 1959). The Hiraiso Formation rests unconformably on the Upper Permian formations and consists of a basal conglomerate overlain by bedded calcareous sandstones followed by alternating beds of calcareous sandstone and mudstone in the upper part. The Osawa Formation consists mainly of calcareous laminated or massive mudstone, but sometimes contains lenticular slump beds of sandstones and conglomerates in the middle part (e.g. Kamada 1980, 1983). The Fukkoshi Formation is made up of sandstones and mudstones, and the thick Isatomae Formation (>1500 m) consists of laminated sandy mudstone to muddy sandstone, intercalated with thick sandstone beds.

The Osawa Formation has yielded the Ichthyosaurian fossil *Utatsusaurus hataii* in association with abundant ammonoids belonging to the *Columbites–Subcolumbites* fauna, and is dated as late Olenekian (Spathian) (Bando & Shimoyama 1974; Bando & Ehiro 1982). Bando (1970) described an early Induan (Griesbachian) ammonoid *Glyptophiceras* cf. *gracile* from the Hiraiso Formation, but it was probably derived from the Osawa Formation and the specific assignment needs to be re-examined (Ehiro 2002). The Hiraiso Formation rests unconformably on the various horizons of the Upper Permian (Wuchiapingian–Changhsingian), and a stratigraphic interval from the lower Induan (uppermost Changhsingian?) to lower Olenekian is missing (Ehiro 2002). The Fukkoshi and Isatomae formations yield Anisian ammonoids (Mojsisovics 1888; Diener 1916; Shimizu 1930; Onuki & Bando 1959; Bando 1964), with the work of Mojsisovics (1888) providing the first description of ammonoid fossils from Japan.

Upper Triassic and lowest Middle Jurassic

Upper Triassic rocks are represented by the Saragai Group which is limited to narrow outcrops within the Shizugawa–Hashiura and Ofunato sub-belts. In contrast, Lower to lower Middle Jurassic rocks are placed within the Shizugawa Group which is found only in the Shizugawa–Hashiura Sub-belt, and Upper Triassic–Lower Jurassic sequences are totally absent from the Karakuwa–Oshika Sub-belt (Fig. 2a.26).

The Upper Triassic Saragai Group comprises the Shindate and Chonomori formations (Shizugawa area; Onuki & Bando 1958) and Uchinohara Formation (Hashiura–Mizunuma area; Takahashi & Onuki 1959) in the Shizugawa–Hashiura Sub-belt, and the Myojinmae Formation in the Ofunato Sub-belt (Kanagawa & Ando 1983). The Shindate Formation, resting unconformably on the Isatomae Formation, is composed of thick sandstone beds with subordinate amounts of mudstone and rare coaly mudstone. The Chonomori Formation conformably covers the Shindate Formation and consists of alternating beds of sandstone and mudstone. Naumann (1881) reported an occurrence of bivalve fossil *Monotis* from the Chonomori Formation, which is the first report of the Triassic fossils from Japan. The Chonomori Formation yields a rich *Monotis* fauna in association with ammonoids *Placites* and *Arcestes* (Shimizu & Mabuti 1932; Onuki & Bando 1958; Nakazawa 1964; Ando 1987), and is dated as being of late Carnian–Norian age. The Uchinohara Formation, unconformably overlying the Isatomae Formation, consists mainly of massive arkose and lacks age-diagnostic fossils, although some authors (e.g. Takizawa *et al.* 1984, 1990) have treated this formation as of Early Jurassic age. The Myojinmae Formation, lying unconformably on Permian strata, consists of pyroclastic rocks, volcanic conglomerate and tuffaceous sandstone, with an occurrence of *Monotis ochotica* having been reported by Kanagawa & Ando (1983).

Lowest Middle Jurassic rocks are represented by the Shizugawa (Shizukawa) Group which comprises the Niranohama Formation and overlying Hosoura Formation. The Niranohama Formation rests unconformably on the Saragai Group and consists of sandstone and sandy mudstone deposited in a brackish–littoral environment (Hayami 1961*a*) and yielded abundant bivalves and middle–upper Hettangian ammonoids (Matsumoto 1956; Hayami 1961*b*; Sato

1962; Takahashi 1969; Sato & Westermann 1991). A diverse belemnite fauna has been reported from this formation (Iba *et al.* 2012). The Hosoura Formation conformably overlies the Niranohama Formation and consists of laminated and massive mudstone, and thin alternating beds of sandstone and mudstone. Many ammonoid species ranging from Sinemurian to Aalenian in age have been obtained from this formation (Sato 1962; Takahashi 1969; Sato & Westermann 1991).

Middle Jurassic–Lower Cretaceous

The Middle Jurassic–Lower Cretaceous sequence in the Shizugawa–Hashiura Sub-belt comprises the Hashiura Group and the overlying Jusanhama Group. The Hashiura Group in the Shizugawa area is, in ascending order, divided into the Aratozaki, Arato and Sodenohama formations. The Aratozaki Formation correlates in other areas with the Nakahara Formation (Hashiura area) and the Kojima Formation (Mizunuma area), and the Arato Formation correlates with the Nagao Formation (Hashiura area) and the Owada Formation (Mizunuma area). The Aratozaki Formation and its equivalents rest unconformably on the Hosoura Formation or the Isatomae Formation, consist mainly of arkose in association with conglomerate and alternating beds of sandstone and mudstone, and yield bivalve fossils such as *Inoceramus*, *Trigonia* and *Vaugonia* (Hayami 1961*b*). The Arato Formation and its equivalents conformably rest on the Aratozaki and equivalent formations and comprise bedded mudstone with alternating beds of sandstone and mudstone. The Arato Formation yielded upper Bajocian ammonoids (Sato 1962; Takahashi 1969). The Nagao Formation yields middle Bajocian–Lower Cretaceous ammonoids (Takahashi 1969; Kase 1979), whereas the lower and middle–upper parts of the Owada Formation contain Bajocian (Suzuki *et al.* 1998) and Oxfordian (Takahashi 1969) ammonoids, respectively. The Sodenohama Formation conformably overlies the Arato Formation, consists of sandstone and laminated mudstone, and contains ammonoid fossils of Oxfordian–Kimmeridgian age (Takahashi 1969).

The Jusanhama Group crops out only in the Hashiura area (Fig. 2a.25) where it has been divided into the Tsukihama and overlying Tategami formations (Takahashi 1961, but see Kase 1979; Takizawa *et al.* 1990 for alternative lithostratigraphic divisions). According to Takahashi (1961), the Tsukihama Formation unconformably covers the Nagao Formation and is mainly composed of massive or bedded, quartzose or arkosic sandstones with intercalated mudstones. The Tategami Formation rests conformably on the Tsukihama Formation, consists of alternating beds of quartzose sandstone and mudstone, and has yielded Late Jurassic–Early Cretaceous brackish water molluscs (Hayami 1960). Given its stratigraphic position above the Nagao Formation, the Jusanhama Group is considered to be Early Cretaceous in age.

In the Karakuwa district of the Karakuwa–Oshika Sub-belt, Mesozoic rocks are represented by the Karakuwa Group and the Oshima Group within the wide, south-plunging Tsunakizaka Syncline (Fig. 2a.24). The Middle Jurassic–lowermost Cretaceous Karakuwa Group is divided into the Kosaba, Tsunakizaka, Ishiwaritoge, Mone, Kogoshio and Isokusa formations in conformably ascending order (Shiida 1940; Ehiro 1974; Takizawa 1985). The Kosaba Formation consists of basal conglomeratic sandstone resting unconformably on the Isatomae Formation, then sandstone intercalated with mudstone that yielded Bajocian molluscs (Hayami 1961*a*). The Tsunakizaka Formation consists of mudstone in its main part and alternating beds of sandstone and mudstone in the uppermost part. It is about 400 m in thickness at the NE part of its outcrop, thins to about 200 m towards the west and south, and has yielded a rich fauna of Bajocian ammonoids (Sato 1962, 1972). The overlying Ishiwaritoge Formation (20–150 m) is composed of conglomerate with alternating beds of sandstone and mudstone deposited in a fluvial–alluvial environment (Takizawa 1985), and also thins towards the west and south (Ehiro 1974). The conglomerate is dominated by granitic pebbles to boulders, thought to be derived from the pre-Silurian Hikami Granite (Kano 1959). The Mone Formation consists of bedded mudstone overlain by alternating beds of arkosic sandstone and mudstone, which is in turn capped by bedded sandstone. It has yielded the bivalve *Myophorella* (*Haidaia*) (Hayami 1961*c*) and ammonoid *Perisphinctes* (Kato *et al.* 1977). The Kogoshio Formation is rich in sandstone and mudstone with rare conglomerate. The sandstone is arkosic, and those in the lower part are thick, coarse-grained and quartzose. This formation has yielded shallow-marine bivalves (Hayami 1961*a*), although the middle part is thought to be an alluvial deposit (Takizawa 1985). The Isokusa (Nagasaki) Formation crops out on Oshima Island in Kesennuma Bay. It consists mainly of sandy mudstone and has yielded Early Cretaceous (Berriasian–Valanginian) ammonoids (Sato 1958; Takahashi 1973; Nara *et al.* 1994). The Oshima Group mostly crops out on Oshima Island and comprises the Kanaegaura and overlying Yokonuma formations (Onuki 1969). The Kanaegaura Formation is very thick (1200 m) and consists mainly of andesite lavas, andesitic pyroclastic rocks and volcanic conglomerates. The covering Yokonuma Formation is composed of sandstone, mudstone and alternating beds of sandstone and mudstone with rare limestone and tuffaceous sandstone; since it has yielded the ammonoid *Crioceratites* (*C.*) *ishiwarai* (Yabe & Shimizu 1925) along with corals and bivalves, it is dated as Hauterivian–Barremian (Obata 1988).

The Jurassic–lowest Cretaceous Oshika Group and overlying Lower Cretaceous Yamadori Formation, widely distributed across the Oshika Peninsula, belong to the Karakuwa–Oshika Sub-belt. The Oshika Group comprises the Tsukinoura, Oginohama and Ayukawa formations in ascending order (Takizawa *et al.* 1974). The Tsukinoura Formation unconformably overlies the Isatomae Formation, and consists of conglomerate and sandstone in the lower part and monotonous mudstone in the upper part. The lower part yields Bajocian ammonoids (Sato 1972) in association with rich bivalves (Hayami 1959, 1961*a*). The Oginohama Formation, conformably covering the Tsukinoura Formation, is composed of sandstone and alternating beds of sandstone and mudstone (Fig. 2a.27) with intercalated conglomerate, and has been divided into the Kitsunezaki, Makinohama, Kozumi and Fukiura members (Takizawa *et al.*

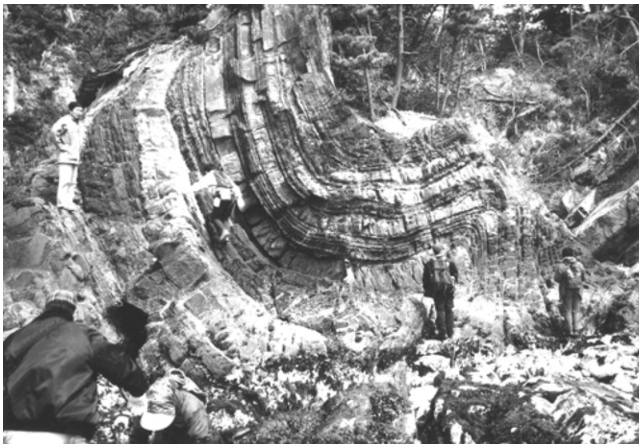

Fig. 2a.27. Photograph showing the folded alternating beds of sandstone and mudstone of the Oginohama Formation (Oshika Group).

1974). The Oginohama Formation yields Oxfordian–Tithonian ammonoids (Inai & Takahashi 1940; Sato 1962; Takahashi 1969; Takizawa *et al.* 1974) and abundant plant fossils belonging to the Ryoseki Flora (Kimura & Ohana 1989*a*, *b*). The Ayukawa Formation conformably covers the Oginohama Formation, consists of arkose and mudstone with rare conglomerate, and is divided into the Kiyosaki, Kobitawatashi, Futawatashi and Domeki members in ascending order (Takizawa *et al.* 1974). The lower part of the Kobitawatashi Member has yielded the Berriasian ammonoid *Berriasella* (Takizawa 1970), whereas the upper part (and Futawatashi Member) yields Valanginian ammonoids (Takizawa 1970; Obata 1988). The Kiyosaki and Domeki members are considered to be continental deposits because they have no marine fossils and contain rich plant fragments and coaly mudstones. The Yamadori Formation unconformably overlies the Ayukawa Formation and consists of andesitic to dacitic pyroclastic rocks in the lower part and basalt and basaltic pyroclastic rocks in the upper part (Takizawa *et al.* 1974).

The Somanakamura Group, distributed in the Soma district of the Abukuma Massif, also belongs to the Karakuwa–Oshika Sub-belt. It comprises the Awazu, Yamagami, Tochikubo, Nakanosawa, Tomizawa and Koyamada formations in ascending order (Yanagisawa *et al.* 1996). These units are dominated by clastic rocks, except for the Nakanosawa Formation which includes thick limestone beds. Based on their ammonoid fauna, the Awazu, Nakanosawa and Koyamada formations have been given ages of Bajocian–Bathonian, Kimmeridgian–Tithonian and Tithonian–Valanginian, respectively (Sato 1962; Mori 1963; Sato *et al.* 2005; Sato & Taketani 2008). The Tochikubo Formation is of particular interest in yielding a rich flora of Ryoseki-type plants (Kimura & Ohana 1989*a*, *b*; Takimoto *et al.* 2008), and the Tomizawa Formation has been interpreted as an alluvial deposit (Takizawa 1985).

In the Ofunato Sub-belt the Lower Cretaceous Ofunato Group is widely distributed, rests unconformably on the Permian strata, and has been divided into the Hakoneyama, Funagawara, Hijochi, Kobosoura and Takonoura formations in conformable ascending order (Onuki & Mori 1961). (Note however that Kanagawa & Ando (1983) interpreted the Hakoneyama Formation, which consists of volcanic conglomerate, as a southern extension of the Upper Triassic Myojinmae Formation.) The Funagawara Formation is composed of conglomerate, sandstone, mudstone and tuffaceous sandstone, and has yielded late Hauterivian ammonoids (Obata & Matsumoto 1977; Matsumoto *et al.* 1982) and late Hauterivian–early Barremian, brackish to shallow-marine bivalves (Kozai & Tashiro 1993). The Hijochi Formation consists of alternating beds of sandstone and mudstone in association with conglomerate and tuff breccias, with the Barremian ammonoid *Holcodiscus* having been reported from its upper part (Obata & Matsumoto 1977). The Kobosoura Formation consists of pyroclastic rocks, tuffaceous sandstone and conglomerate with intercalated sandstone and mudstone, and contains the bivalve *Pterotrigonia* (*Pterotrigonia*) dated as Hauterivian–Aptian (Tashiro & Kozai 1989). The Takonoura Formation consists mainly of alternating beds of sandstone and mudstone with conglomerate and tuff.

Lower Cretaceous volcanic-dominated strata are widely distributed throughout the Kitakami Massif, resting on Palaeozoic formations in the Southern Kitakami Massif and on Jurassic accretionary complexes in the Northern Kitakami Massif. The volcanic rocks are subduction-related, range from basalt to rhyolite, and include a volcano-plutonic complex with Early Cretaceous (120–110 Ma) granitic rocks (Kanisawa 1974). The abundant presence of Early Cretaceous adakitic volcanic and granitic rocks implies the subduction of young, hot oceanic plate (Tsuchiya & Kanisawa 1994; Tsuchiya *et al.* 2005).

Mesozoic covering strata in the Kurosegawa Belt

In central Shikoku, a shallow-marine Mesozoic cover sequence is represented by the Middle Triassic Zohoin Formation, Upper Triassic Kochigatani Group, Triassic–Jurassic Naruo Formation and the Jurassic Keta Formation (Hada *et al.* 1992).

The Zohoin Formation consists of mudstone with sandstone and rare tuffaceous mudstone, is in fault contact with other strata, and yields bivalves and ammonoids of Ladinian age (Bando 1964). It is therefore coeval with the *Daonella*-bearing mudstone-dominated Ladinian Usugatani Formation which crops out in eastern Shikoku (Hirayama *et al.* 1956). The Kochigatani Group, which is also found in western Shikoku, is of Carnian–Norian age (Bando 1964) and divided into lower and upper subgroups. The lower subgroup is composed of sandstone, mudstone and alternating beds of sandstone and mudstone, with conglomeratic sandstone, and has yielded the bivalves *Halobia*, *Tosapecten* and the ammonoid *Paratrachyceras*. The upper subgroup is composed of mudstone and sandstone and contains the bivalve *Monotis* and ammonoid *Sirenites* (Tsujino *et al.* 2013). The Sabudani Formation in eastern Shikoku is an equivalent of this group, which rests unconformably on the Permian accretionary complex and Devonian shallow-marine strata (Ichikawa *et al.* 1956; Yoshikura *et al.* 1990).

The Naruo and its equivalent Nakanose (western Shikoku) formations are composed of clastic rocks and contain latest Triassic–Early Jurassic radiolarians (Hada *et al.* 1992). The Keta Formation, and its equivalent Kagio Formation in western Shikoku, are divided into the lower (sandstones and conglomerates) and upper (mudstone) members. The Kagio Formation unconformably covers the Mitaki Igneous Complex (Hada 1974; Tominaga *et al.* 1979). Radiolarians from these formations are of early Middle Jurassic age (Matsuoka 1985; Hada *et al.* 1992).

Lower Cretaceous Miyako Group, Upper Cretaceous and Palaeogene strata

The Aptian–Albian Miyako Group is distributed along the Pacific coast of the Kitakami Massif, and Upper Cretaceous and associated Palaeogene strata are distributed in the Kuji district (Kuji and Noda groups) and some narrow areas in the northern Kitakami Massif, and the Futaba district (Futaba and Shiramizu groups) are distrubted in the SE Abukuma Massif (Fig. 2a.18). Although these strata were deposited after the Early Cretaceous amalgamation of the South Kitakami and North Kitakami belts and are therefore not strictly part of the South Kitakami Belt, they are briefly described to understand the geological development of the Kitakami Massif.

The Miyako Group, which rests on Jurassic accretionary complexes and Early Cretaceous volcanic and granitic rocks with remarkable unconformity (see Fig. 2a.28), comprises mainly storm-dominated shallow-marine sequences (Mochizuki & Ando 2003). The group has yielded abundant fossils such as ammonoids, molluscs and corals, and is dated as late Aptian–early Albian based on ammonoids (Matsumoto 1953, 1963; Hanai *et al.* 1968).

The Upper Cretaceous Kuji Group (Shimazu & Teraoka 1962) in the Kuji district was deposited in an alluvial-shallow marine basin (Minoura & Yamauchi 1989; Terui & Nagahama 1995), and consists of clastic rocks with minor amounts of tuff. Based on the ammonoids and inoceramids, Matsumoto *et al.* (1982) dated the group as late Coniacian–Campanian, whereas Futakami *et al.* (1987) considered the group to be Santonian–early Campanian. The Futaba Group (Matsumoto 1943) in the Futaba district is composed mainly of clastic rocks deposited during three sedimentary cycles, each of which begins with fluvial conditions and finishes with a lagoonal or shallow-marine facies (Ando *et al.* 1995). It

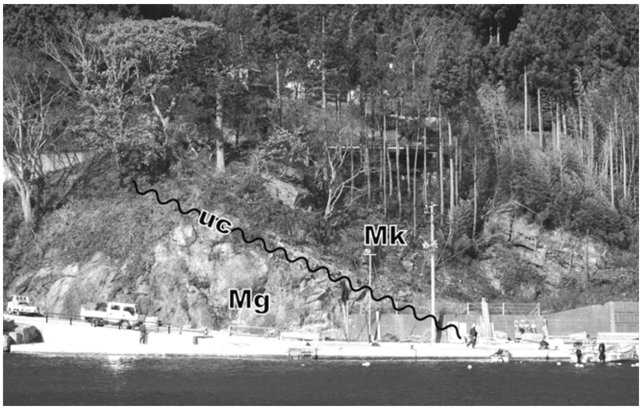

Fig. 2a.28. Photograph showing the unconformity between the Jurassic accretionary complex of the North Kitakami Belt and Lower Cretaceous Miyako Group at Raga, Tanohata, Iwate Prefecture. Mg, sandstone of the Jurassic accretionary complex (Magisawa Unit), dipping vertically; Mk, alternating beds of sandstone and mudstone of the Miyako Group, dipping gently eastwards.

has yielded a rich fauna of ammonoids and inoceramids of Coniacian–early Santonian age (Obata & Suzuki 1969; Matsumoto *et al.* 1990; Kubo *et al.* 2002), including the plesiosaur *Futabasaurus suzukii* (Sato *et al.* 2006).

The Palaeogene Noda Group in the Kuji district unconformably covers the Upper Cretaceous succession and consists of clastic rocks deposited in an alluvial environment showing four sedimentary cycles, each starting with conglomerate and ending with coaly mudstone rich in land plants (Shimazu & Teraoka 1962). Shimazu & Teraoka (1962) correlated the whole group with the Oligocene, whereas Horiuchi & Kimura (1986) & Uemura (1997) considered that the lower part of the group is Paleocene in age. The upper Eocene–lower Oligocene Shiramizu Group (Kubo *et al.* 2002) in the Futaba district overlies unconformably both metamorphic rocks of the Abukuma Belt and Upper Cretaceous strata, and comprises clastic rocks deposited in a fluvial to shallow-marine environment with a coal bed in the lower part (Kubo *et al.* 2002; Suto *et al.* 2005).

Geological development and palaeogeography of the South Kitakami–Kurosegawa Belt

Although fragmental and scattered in fault-bounded small outcrops, the rocks of the Kurosegawa Belt closely resemble those of the South Kitakami Belt; both include Cambro-Ordovician granitic and metamorphic basement, Early Silurian welded tuff, halysitid-bearing Silurian limestone, Siluro-Devonian volcaniclastics, *Leptophloeum*-bearing Late Devonian and late Early Carboniferous *Kueichouphyllum*-bearing limestone (Figs 2a.19 & 2a.23). The Kurosegawa Belt is therefore interpreted to have been a part of the South Kitakami Belt ('South Kitakami Palaeoland'; Ehiro & Kanisawa 1999) during Palaeozoic times (Umeda 1996*a*; Ehiro 2000).

Early–Middle Palaeozoic

The South Kitakami Cambro-Ordovician basement comprises high-*P* (or medium-*P*) metamorphic rocks of accretionary complex origin and subduction-related intrusives. These rocks represent the continental crust of a 'South Kitakami Palaeoland' created along the active continental margin of an ancient continent (Ehiro & Kanisawa 1999), a tectonic setting that persisted through Devonian and Carboniferous times (Kawamura & Kawamura 1989*b*). Silurian palaeomagnetic data from the Kurosegawa Belt indicate low latitudes (5–15°; Shibuya *et al.* 1983). Since no valid palaeomagnetic data have been obtained from the Pre-Cretaceous rocks of the South Kitakami Belt due to the thermal effect by the Early Cretaceous granitic intrusives, the palaeogeography of the South Kitakami Palaeoland has been reconstructed mainly by using palaeobiogeographic data. Kato (1990) argued that the Silurian–Devonian coral faunas from the South Kitakami Belt and Kurosegawa Belt bear a close resemblance to those from eastern Australia and South China, especially to the former. On the other hand, Kido & Sugiyama (2011) stressed that the Silurian corals from the Kurosegawa Belt are more closely comparable with those from South China, often down to individual species level, than to those from Australia. It is therefore highly probable that the South Kitakami Palaeoland was established and developed in a marginal part of South China or nearby continental area during Early–Middle Palaeozoic times.

Late Palaeozoic

In contrast to the Carboniferous strata, Permian outcrops almost completely lack pyroclastic sediments, a fact leading Minoura (1985) to deduce that the South Kitakami Belt was a passive environment from Permian until earliest Cretaceous times when arc volcanism resumed. The Middle Permian Usuginu-type conglomerates contain abundant granitic clasts dated as 257–244 Ma (CHIME ages: Takeuchi & Suzuki 2000), but there is no outcrop of the source granitic rocks in the South Kitakami Belt. Kano (1971) and Iwai & Ishizaki (1966) considered that the granitic clasts were derived from the dissected, lost uplift zones in the belt, but Yoshida *et al.* (1994) and Yoshida & Machiyama (1998) concluded that the hinterland was a magmatic arc situated to the west of the belt. Based on the CHIME ages of the granitic clasts, Takeuchi & Suzuki (2000) speculated that the source granitic rocks were intruded just before the deposition of the Usuginu-type conglomerate, uplifted rapidly, and eroded and transported in the sedimentary basin.

The Australian affinity of the corals of the South Kitakami Palaeoland persisted until early Visean times when the coral faunas changed drastically to have affinity only with those of the South China area (Kato 1990; Niikawa 1994). The Early Carboniferous flora also belongs to the Cathaysian Floral Province (Asama *et al.* 1985; Kato *et al.* 1989). The southwards drift of Australia (Eastern Gondwana) took place during Late Devonian or Early Carboniferous times (Powell & Li 1994), and the Palaeotethys Sea which separated the Cathaysian and Gondwana continents already existed in the Devonian Period (e.g. Metcalfe 1996). This palaeobiogeographic change is therefore considered to be due to the separation of South China (along with South Kitakami) from the Gondwana continent during Devonian times.

The Permian flora of the South Kitakami Belt similarly belongs to the Cathaysian Floral Province (Asama 1956, 1967, 1974; Asama & Murata 1974), and so also differs from that of the Australia area (Gondwanan Province). It is noteworthy that in North China the species of the genus *Gigantopteris*, one of the typical constituents of Cathaysian flora and widespread in South China, have only been reported from the southernmost areas, namely in Xu-Huai-Yu subprovince (Mei *et al.* 1996) or the concurrent *Gigantonoclea–Gigantopteris* region (Zhang *et al.* 1999), as well as the South Kitakami Belt. The Permian bivalve (Fang 1985; Nakazawa 1991; Fang & Yin 1995) and ammonoid faunas (Ehiro 1997; Ehiro *et al.* 2005) are closely allied to those of South China, as are Middle–Late Permian fusulinoideans (Ozawa 1987). In the same way, according to Wang *et al.* (2006), Early–Middle Permian coral faunas closely resemble those of South China (sometimes down to species level), and Kawamura & Machiyama (1995) observed that the biotic composition of the reefal Middle Permian Iwaizaki Limestone is typical tropical and closely allied to those of South China and Indochina.

Although Tazawa (1991, 2002) and Shi *et al.* (1995) used Middle Permian brachiopod fauna to argue that the South Kitakami Belt was located near the NE margin of North China, the majority of palaeobiogeographic data indicate that the 'South Kitakami Palaeoland' was located at or near the eastern margin of South China during Late Palaeozoic time (Ehiro 2001).

Mesozoic

Following marine regression at the Permian–Triassic boundary, the sea re-entered the South Kitakami Belt in late Early Triassic times (Ehiro 2002), spreading uniformly over the Shizugawa–Hashiura and Karakuwa–Oshika sub-belts by Middle Triassic times so that there are only minor differences in stratal lithofacies and thicknesses. After a regressive phase at the Middle–Late Triassic boundary, subsequent transgression was restricted only to the narrow area represented by the Shizugawa–Hashiura and Ofunato sub-belts. The deposition of Lower–lower Middle Jurassic strata was restricted only to the Shizugawa–Hashiura Sub-belt, and Upper Triassic–Lower Jurassic rocks are totally absent from the Karakuwa–Oshika Sub-belt. Whereas the thicknesses of the Middle Jurassic strata of the Shizugawa–Hashiura and Karakuwa–Oshika sub-belts are similar, Upper Jurassic–Lower Cretaceous strata are very much thicker in the Karakuwa–Oshika Sub-belt. We may therefore deduce that the sedimentary basin depocentre lay initially in the Shizugawa–Hashiura Sub-belt (Late Triassic–Early Jurassic) but shifted east to the Karakuwa–Oshika Sub-belt in Late Jurassic times (Fig. 2a.26; Yamashita 1957; Takizawa 1977, 1985).

The South Kitakami Belt continued to be located at or near the eastern margin of South China until Early Cretaceous times, as indicated by Mesozoic faunas and floras in the South Kitakami Belt (Ehiro & Kanisawa 1999; Ehiro 2001). Early Triassic ammonoids in the Osawa Formation belong to the lower latitudinal *Columbites–Subcolumbites* fauna (Ehiro 1997; Brayard *et al.* 2009), although Nakazawa (1991) considered that the Middle Triassic *Daonella* fauna has some affinity with that of Siberia and Ando (1987) stressed that the Late Triassic *Monotis* fauna from the Saragai Group show similarity to that of the Boreal province. Although the Jurassic ammonoid fauna similarly includes Boreal elements (the genus Kepplerites; Takizawa 1977), most Jurassic ammonoids in the South Kitakami Belt are of Tethys-Pacific type (Bando *et al.* 1987). Finally, Late Jurassic–Early Cretaceous flora in East Asia can be divided into the Ryoseki Floral Province, which typically developed in the Outer Zone of SW Japan and South China, and the Tetori Floral Province, distributed at higher latitudes and so now found in the Inner Zone of SW Japan and the NE China–east Siberia area (Kimura 1987). The Late Jurassic–earliest Cretaceous floral assemblages from the South Kitakami and Kurosegawa belts belong to the Ryoseki Floristic Province (Kimura 1987; Kimura & Ohana 1989*a*, *b*). Palaeomagnetic study of the Mesozoic sedimentary rocks of the Kurosegawa Belt in Kyushu District shows palaeolatitudes for the Late Triassic and Early Cretaceous as $3.5 \pm 3.8°$ N and $18.4 \pm 2.5°$ N, respectively, which coincide with those of South China (Uno *et al.* 2011).

The Palaeozoic–Lower Cretaceous (Barremian) strata in the Kitakami and Abukuma massifs, including the Jurassic accretionary complexes, were strongly folded and faulted during the Early Cretaceous Oshima Orogeny (Kobayashi 1941) and then intruded by Early Cretaceous granitic rocks (Fig. 2a.27). NNW-trending left-lateral strike-slip faults, with displacements of several tens to several hundreds of kilometres, were still active during the early phase of the granitic activity, and the granites (120–110 Ma) along the faults consequently display strong mylonitic fabrics (e.g. Otsuki & Ehiro 1978, 1992). In contrast, the Aptian–Albian Miyako Group and Upper Cretaceous strata are only weakly deformed and rest on the pre-Miyakoan folded strata and granites with pronounced unconformity (Fig. 2a.28). The distribution area of these strata on the land surface is rather narrow, due to the shift of the sedimentary depocentre to the east after the Oshima Orogeny. Thick Cretaceous–Palaeogene strata are widely distributed beneath the continental shelf off the coasts of Kitakami (Osawa *et al.* 2002) and Abukuma (Iwata *et al.* 2002), however.

Appendix

English to Kanji and Hiragana translations for geological and place names

Abukuma	阿武隈	あぶくま
Ainosawa	合ノ沢	あいのさわ
Akaiwa	赤岩	あかいわ
Akiyoshi	秋吉	あきよし
Arakigawa	荒城川	あらきがわ
Arato	荒砥 (荒戸)	あらと
Aratozaki	荒砥崎	あらとざき
Arisu	有住	ありす
Awazu	粟津	あわづ
Ayukawa	鮎川	あゆかわ
Buntoku	文徳	ぶんとく
Chonomori	長の森	ちょうのもり
Chugoku	中国	ちゅうごく
Dogo	島後	どうご
Doi	土居	どい
Domeki	百目木	どうめき
Ezo	恵蘇	えぞ
Fujikuradani	藤倉谷	ふじくらだに
Fukata	深田	ふかた
Fukkiura	福貴浦	ふっきうら
Fukkoshi	風越	ふっこし
Fuko	普甲	ふこう
Fukuji	福地	ふくじ
Funagawara	船河原	ふながわら
Funatsu	船津	ふなつ
Futaba	双葉	ふたば
Futawatashi	長渡	ふたわたし
Gionyama	祇園山	ぎおんやま
Gomi	五味	ごみ
Haigyu	拝宮	はいぎゅう
Hakoneyama	箱根山	はこねやま
Happo (Happou)	八方	はっぽう
Hashiura	橋浦	はしうら
Hatakawa	畑川	はたかわ
Hayachine	早池峰	はやちね
Hida	飛騨	ひだ
Hida Gaien	飛騨外縁	ひだがいえん
Higo	肥後	ひご
Hijochi	飛定地	ひじょうち
Hikami	氷上	ひかみ
Hikamisan	氷上山	ひかみさん
Hikoroichi	日頃市	ひころいち
Hiraiso	平磯	ひらいそ
Hirasawa	平沢	ひらさわ
Hitachi	日立	ひたち
Hitoegane	一重ヶ根	ひとえがね
Hosoo	細尾	ほそお
Hosoura	細浦	ほそうら
Ichinose	市ノ瀬	いちのせ
Ichinotani	一の谷	いちのたに
Ichiyama	市山	いちやま

(*Continued*)

English to Kanji and Hiragana translations for geological and place names (Continued)

Inai	稲井	いない
Ino	伊野	いの
Isatomae	伊里前	いさとまえ
Ise	伊勢	いせ
Ishiwaritoge	石割峠	いしわりとうげ
Isokusa	磯草	いそくさ
Itoigawa	糸魚川	いといがわ
Itoshiro	石徹白	いとしろ
Iwaizaki	岩井崎	いわいざき
Iwatsubodani	岩坪谷	いわつぼだに
Izushi	出石	いずし
Joryu	上流	じょうりゅう
Kagero	影路	かげろ
Kagio	嘉義尾	かぎお
Kagura	神楽	かぐら
Kakisako	柿迫	かきさこ
Kamaishi	釜石	かまいし
Kamianama	上穴馬	かみあなうま
Kamioka	神岡	かみおか
Kamiyasse	上八瀬	かみやっせ
Kanaegaura	鼎ヶ浦	かなえがうら
Kanokura	叶倉	かのくら
Karakuwa	唐桑	からくわ
Karaumedate	唐梅舘	からうめだて
Katsura	桂	かつら
Kawauchi	川内	かわうち
Kentosan	犬頭山	けんとうさん
Kesennuma	気仙沼	けせんぬま
Keta	毛田	けた
Kitakami	北上	きたかみ
Kitsunezaki	狐崎	きつねざき
Kiyomi	清見	きよみ
Kiyosaki	清崎	きよさき
Kobitawatashi	小長渡	こびたわたし
Kobosoura	小細浦	こぼそうら
Kochigatani	川内ヶ谷	こうちがたに
Kogoshio	小々汐	こごしお
Koguro	小黒	こぐろ
Kojima	小島	こじま
Komatsu	小松	こまつ
Konogidani	此木谷	このぎだに
Kosaba	小鯖	こさば
Kotaki	小滝	こたき
Kowaragi	小原木	こわらぎ
Koyamada	小山田	こやまだ
Kozaki	小崎	こざき
Kozumi	小積	こづみ
Kuji	久慈	くじ
Kuma	球磨	くま
Kurosawa	黒沢	くろさわ
Kurosegawa	黒瀬川	くろせがわ
Kuruma	来馬	くるま
Kuzuryu	九頭竜	くずりゅう
Maizuru	舞鶴	まいづる
Makinohama	牧の浜	まきのはま
Mano	真野	まの
Matsugadaira	松ヶ平	まつがだいら
Mino	美濃	みの
Misaka	三坂	みさか
Mitaki	三滝	みたき
Miyadani	宮谷	みやだに
Miyagadani	宮ヶ谷	みやがだに

(*Continued*)

English to Kanji and Hiragana translations for geological and place names (Continued)

Miyako	宮古	みやこ
Miyamori	宮守	みやもり
Miyanaro	宮成	みやなろ
Mizunuma	水沼	みずぬま
Mizuyagadani	水屋ヶ谷	みずやがだに
Mone	舞根	もうね
Moribu	森部	もりぶ
Motai	母体	もたい
Myojinmae	明神前	みょうじんまえ
Nabekoshiyama	鍋越山	なべこしやま
Nagaiwa	長岩	ながいわ
Nagao	長尾	ながお
Nagasaka	長坂	ながさか
Nagasaki	長崎	ながさき
Naidaijin	内大臣	ないだいじん
Nakadaira	中平	なかだいら
Nakadake	中岳	なかだけ
Nakahara	中原	なかはら
Nakahata	中畑	なかはた
Nakanosawa	中ノ沢	なかのさわ
Nakanose	中之瀬	なかのせ
Nakazato	中里	なかざと
Nameirizawa	名目入沢	なめいりざわ
Naradani	楢谷	ならだに
Naruo	成穂	なるほ
Natsuyama	夏山	なつやま
Nedamo	根田茂	ねだも
Niranohama	韮の浜	にらのはま
Nishikori	錦織	にしこり
Noda	野田	のだ
Ochi	越智	おち
Odagoe	小田越	おだごえ
Odaira	大平	おおだいら
Oeyama	大江山	おおえやま
Ofunato	大船渡	おおふなと
Ogatsu	雄勝	おがつ
Oginohama	荻の浜	おぎのはま
Oguradani	小椋谷	おぐらだに
Okago	大籠	おおかご
Okanaro	岡成	おかなろ
Oki	隠岐	おき
Okuhinotsuchi	奥火の土	おくひのつち
Omi	青海	おうみ
Onimaru	鬼丸	おにまる
Ono	大野	おおの
Orikabetoge	折壁峠	おりかべとうげ
Osawa	大沢	おおさわ
Osayama	大佐山	おおさやま
Oshika	牡鹿	おしか
Oshima	大島	おおしま
Owada	大和田	おおわだ
Raga	羅賀	らが
Renge	蓮華	れんげ
Rifu	利府	りふ
Rosse	呂瀬	ろっせ
Ryoke	領家	りょうけ
Ryoseki	領石	りょうせき
Sabudani	寒谷	さぶだに
Sakamotozawa	坂本沢	さかもとざわ
Sanbagawa (Sambagawa)	三波川	さんばがわ
Sangun	三郡	さんぐん
Saragai	皿貝	さらがい

(*Continued*)

English to Kanji and Hiragana translations for geological and place names (*Continued*)

Sekinomiya	関宮	せきのみや
Senjogataki	千丈ヶ滝	せんじょうがたき
Senmatsu	千松	せんまつ
Setamai	世田米	せたまい
Shibasudani	子馬巣谷	しばすだに
Shimozaisho	下在所	しもざいしょ
Shindate	新舘	しんだて
Shiraishi	白石	しらいし
Shiramizu	白水	しらみず
Shiroumadake	白馬岳	しろうまだけ
Shittakasawa	尻高沢	しったかさわ
Shizugawa (Shizukawa)	志津川	しづがわ
Shoboji	正法寺	しょうぼうじ
Sodenohama	袖の浜	そでのはま
Soma	相馬	そうま
Somanakamura	相馬中村	そうまなかむら
Sorayama	空山	そらやま
Suberidani	辷谷	すべりだに
Suenosaki	末の崎	すえのさき
Suketsune	助常	すけつね
Takayama	高山	たかやま
Takonoura	蛸浦	たこのうら
Tanba (Tamba)	丹波	たんば
Tandodani	谷土谷	たんどだに
Tanohata	田野畑	たのはた
Tanoura	田の浦	たのうら
Tari	多里	たり
Tategami	立神	たてがみ
Terano	寺野	てらの
Tetori (Tedori)	手取	てとり
Tobigamori	鳶ヶ森	とびがもり
Tochikubo	栃窪	とちくぼ
Tomizawa	富沢	とみざわ
Toyoma	登米	とよま
Tsubonosawa	壺の沢	つぼのさわ
Tsukihama	月浜	つきはま
Tsukinoura	月の浦	つきのうら
Tsunakizaka	綱木坂	つなきざか
Tsurukoba (Tsurunokoba)	鶴木場	つるこば
Uchinohara	内の原	うちのはら
Unazuki	宇奈月	うなづき
Unoki	鵜ノ木	うのき
Usugatani	臼ヶ谷	うすがたに
Usuginu	薄衣	うすぎぬ
Wakamatsu	若松	わかまつ
Wakasa	若桜	わかさ
Wariyama	割山	わりやま
Yaeyama	八重山	やえやま
Yakuno	夜久野	やくの
Yakushigawa	薬師川	やくしがわ
Yamagami	山上	やまがみ
Yokamachi	八日町	ようかまち
Yokokurayama	横倉山	よこくらやま
Yokonuma	横沼	よこぬま
Yokota	横田	よこた
Yoshiki	吉城	よしき
Yuzuruha	湯鶴葉	ゆづるは
Zohoin	蔵法院	ぞうほういん

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