Chapter 7

Late Cenozoic Tectonic Events and Intra-Arc Basin Development in Northeast Japan

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Additional information is available at the end of the chapter

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1. Introduction

On 11 March 2011, a catastrophic earthquake (Mw 9.0) took place along the Japan Trench off the Pacific coast of northeast Japan and about 20 thousands people were lost mainly by Tsunamis. The 2011 great earthquake occurred along a subduction zone where the Pacific Plate subducts perpendicularly toward the North American Plate at a rate of 8~9 cm/year (Figure. 1). Coseismic slip caused by the 2011 earthquake exceeded 50 m around the epicenter [1]. There has been a discrepancy between short-term (geodetic) and long-term (geologic) strain rates in both horizontal and vertical directions over the northeast Japan arc [2-3]. Geodetic observations [1, 3] have revealed horizontal shortening rate at around several tens mm/yr across northeast Japan arc, which is almost half of the subduction rate, whereas geological observations [4] have revealed horizontal shortening rate at around 3~5 mm/yr, which is one order of magnitude slower than geodetic observations [3]. Only a fraction (~10%) of plate convergence is, therefore, accommodated within the northeast Japan arc as long-term deformation. The 2011 great earthquake was one of the decoupling events that effectively released the accumulated elastic strain due to a plate coupling[3]. Study on long-term (geologic) deformation within the northeast Japan arc was thus proved to be crucial for assessing such extraordinary large decoupling events because they are too rare events (~one per kyr) to be detected by short-term (geodetic) observations.

The Japanese Islands are divided into northeast and southwest Japan by an inter-arc rift system called "Fossa Magna" bounded by the Itoigawa-Shizuoka tectonic line (ISTL; Figure. 1) and the Tonegawa tectonic line (TTL; Figure. 1) that divides northeast Japan from southwest Japan [5]. This chapter focuses on the Northeast Honshu where three lines of ranges, the Kitakami and Abukuma mountains, the Ou backbone Range, and the Dewa Hills from east to west, run parallel to the Japan Trench (Figure. 1). This chapter aims to clarify Late Cenozoic tectonic events in Northeast Japan based on reviews of recent high-resolution studies on modes and development patterns of intra-arc basins.

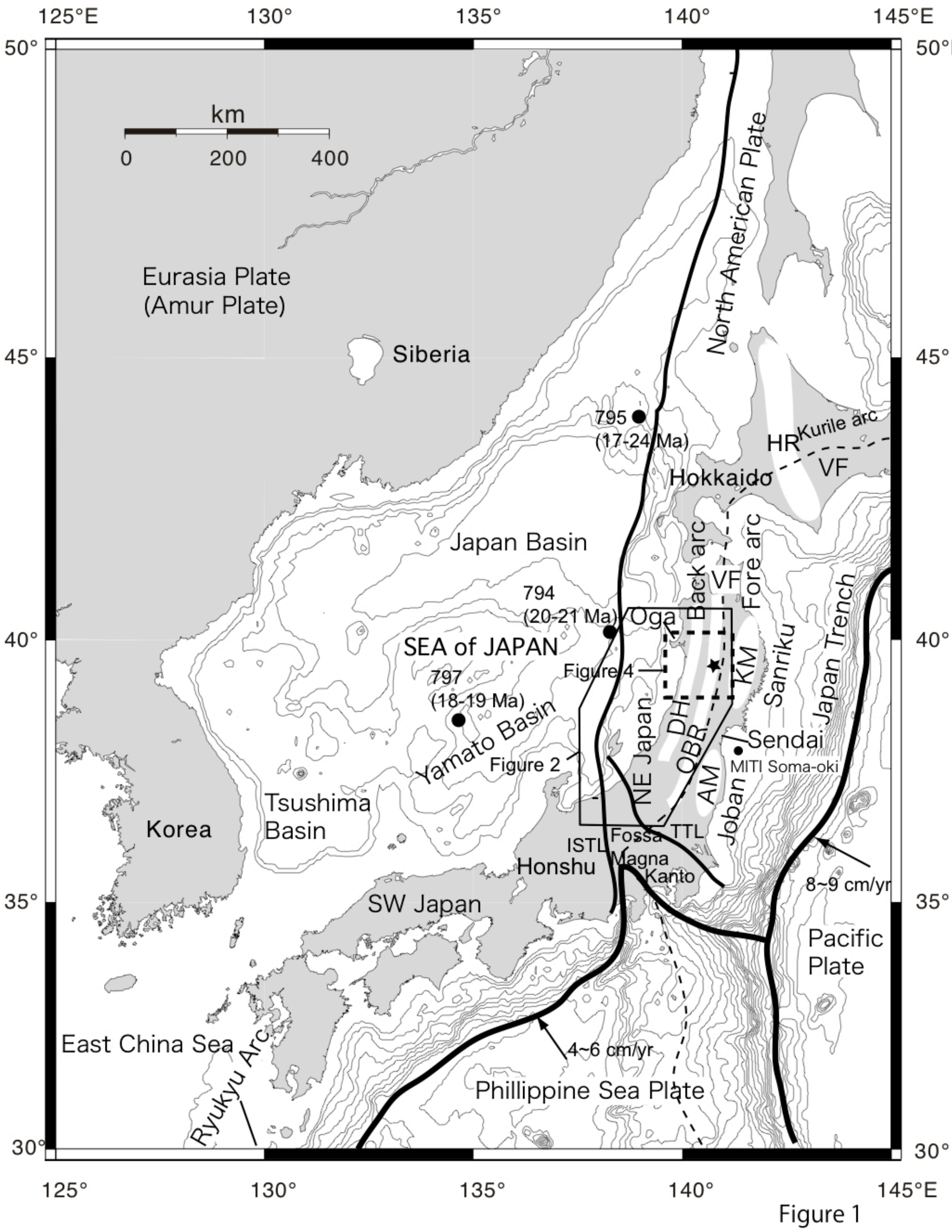

Figure 1. Index map showing the tectonic setting of Northeast Japan. OBR: Ou Backbone Range, DH: Dewa Hills, KM: Kitakami Mountains, AM: Abukuma Mountains, HR: Hidaka Range, VF: Quaternary volcanic front (dotted line). ISTL: Itoigawa-Shizuoka Tectonic Line, TTL: Tonegawa Tectonic Line. Solid circle represents ODP sites with ages of basement rocks and MITI soma-oki well. Star represents location of the Yuda Basin. A hexagone and a dotted square denote the areas shown in Figure 2 and Figure 4, respectively.

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2. Opening of the Sea of Japan

The Sea of Japan is a large marginal sea formed in the backarc side of the Southwest and Northeast Japan and comprises component backarc basins, the Japan, Yamato and Tsushima basins (Figure. 1). The Japan Basin is underlain by an oceanic-type crust 11-12 km thick, whereas the Yamato Basin is floored by a crust 17-19 km thick and is unlikely to be oceanic [6]. The component structure of the Sea of Japan likely constitutes a multi rift system [7]. The Japan Basin was dated around 24-17 Ma based on ^{40}Ar-^{39}Ar dating of basement basalt in Site 795 of ODP Leg. 127 (Figure. 1)[8]. The Yamato Basin yields younger age of 21-18 Ma in Site 794 and 797 of ODP Leg. 127 (Figure. 1)[8]. However, the timing and processes of the opening of the Sea of Japan has been a matter of debate. Two mechanical models for the formation of the Japan arc – the Sea of Japan system has been proposed; a double-door opening model and a pull-apart basin model.

A double-door opening model is primarily based on the paleomagnetic evidence on the Japan arc. The model includes a clockwise rotation of the southwest Japan [9] and a counter-clockwise rotation of the northeast Japan [10-11]. This model implies a simultaneous occurrence of the rotation of the Japan arc and the opening of the Sea of Japan. Paleomagnetic data suggest a rapid 50° clockwise rotation of the Southwest Japan at around 15 Ma [12] and rather prolonged counter-clockwise rotation of the Northeast Japan from 17 to 14 Ma [13]. In terms of discrepancy between the timing of the opening of the Sea of Japan and that of the rotation of SW and NE Japan, Nohda [14] reassessed the ^{40}Ar-^{39}Ar age data of basement basalt with reference to Nd-Sr isotopic data in Site 797 of ODP Leg. 127. The basalts from the upper part of basements in Site 797 have not been dated and were overlain by a felsic tuff dated to be 14.86 Ma. He concluded that the upper basalts at Site 797 may be inferred to be younger than the lower basalts and that the inferred timing of volcanic activity in the Sea of Japan region (ca. 21-15 Ma) is consistent with the timing of rotational crustal movements inferred from palaeomagnetic studies in the Japanese arcs.

A pull-apart basin model is primarily based on the structural studies. In this model, the opening of the Sea of Japan was attributed to pull-apart basins formed by lateral displacement of the Japan arc associated with dextral transcurrent fault systems [15-17]. Jolivet et al. [18] revised their previous model and incorporated paleomagnetic evidence of rotation into their model; strike-slip displacement of the Japan arcs was associated by clockwise and counter-clockwise rotations of numerous blocks in southwest and northeast Japan, respectively.

There has been still no general agreement on the timing and processes of the opening of the Sea of Japan. However, Takahashi [5] proposed that the Early Miocene volcanic front of the northeast Japan was displaced more than 200 km toward east from that of southwest Japan because of the right lateral strike-slip motion of the Tonegawa tectonic line (TTL; Figure. 1). This suggests a differential rotation between southwest and northeast Japan accompanied by southward displacement of northeast Japan.

3. Intra-arc rifting

3.1. Incipient rifting

In response to opening of the Sea of Japan, both Northeast and Southwest Japan Arcs were subjected to intra-arc rifting. Recent studies demonstrated that the two-stage intra-arc rifting occurred along the eastern margin of the Sea of Japan [19-21]. Kano et al. [21] reviewed that Late Eocene and Oligocene (ca. 35-23 Ma) marine sediments distribute sparsely along the eastern margin of the Sea of Japan from Sakhalin to Kyushu through Honshu. It is suggested that rifting started and incipient rift system was formed by the Oligocene time on the back-arc side prior to the opening of the Japan and Yamato basins. During Oligocene-Early Miocene (34-21 Ma) volcanic rocks accumulated in southwest Hokkaido and Oga Peninsula with petrological and geochemical features similar to those of the volcanic rocks from continental rifts, suggesting that the former volcanic rocks were formed under rifting in the Eurasian continental arc during the pre-opening stage of the Sea of Japan [21, 23]. The early phase of incipient rifting during Late Eocene to Oligocene was characterized by slow subsidence (< 800m/m.y.) in the rifted zones [21]. The incipient rifting was interrupted by a regional unconformity prior to the succeeding rapid rifting [21].

Figure 2. Index map showing the Eastern Japan Sea Rift System [24].

3.2. Rapid rifting

Besides large backarc basins such as the Japan and Yamato basins in the Sea of Japan, the Eastern Japan Sea Rift System [24] was generated along the Sea of Japan coast of the northeast Honshu (Figure. 2) as a series of the NE-SW trending rift basins at around 16 Ma [25], which corresponded to the final phase of the backarc opening. The Eastern Japan Sea Rift System consists of several composite basins, such as the Niigata Basin in the south and the Akita Basin in the north (Figure. 2). In the Uetsu district between the Niigata and Akita basins (Figure. 2), many half grabens trending NNE-SSW to NE developed [26]. Subsidence analysis for syn-rift basins showed that rapid rifting started around 18 Ma and ceased around 15 Ma [26]. The maximum subsidence rate exceeded 1 km/m.y., much faster than that in major continental rifts [26]. Intra-arc rifting in outer arc and in most of inner arc of northeast Honshu ended at around 15 Ma [25-26].

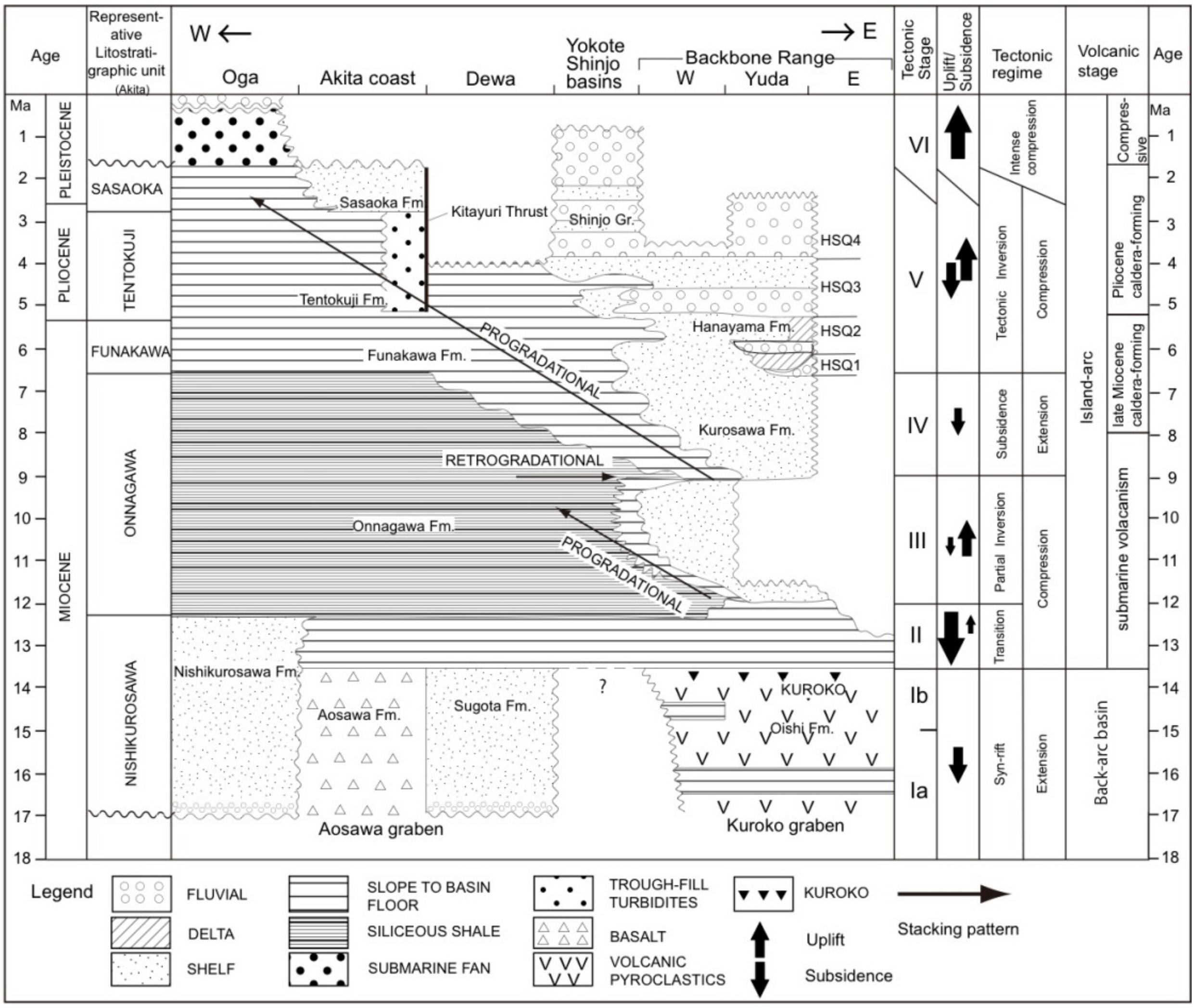

Figure 3. Tectonostratigraphic stage division chart of the Akita Basin showing division of lithostratigraphic units, major depositional systems, stacking patterns, uplift and subsidence patterns, tectonic regimes, and volcanic regimes. Compiled from [22, 27-33].

In the center of the Akita Basin as well as of the Niigata Basin, thick piles of basalts accumulated in grabens [34-35]. The Middle Miocene (16-13 Ma) northeast Japan was characterized by a distinct bimodal volcanism in which basalts dominated in back-arc side whereas fore-arc side was dominated by felsic rocks [23]. Figure 3 shows a tectonostratigraphic stage division chart of the Akita Basin. The Akita Basin was further subdivided into the coastal basin (basin center) and Yokote and Shinjo marginal basins by an intervening ridge, which have uplifted to form the Dewa Hills at present (Figure. 2). During rapid rifting, syn-rift successions filled horsts and grabens formed in the Akita Basin. Two grabens; the Aosawa graben [34] along the Akita coast and the Kuroko graben in the backbone range [36] were formed in the Akita Basin (Figure. 3). The Dewa Hills and Oga Peninsula constituted the intervening structural highs (horsts). The Aosawa graben is a large-scale graben formed by pull-apart tectonism, and was filled with thick piles (~2,000 m) of graben-fill basalt lavas (Aosawa Formation) [34]. The Kuroko graben is composed of component half grabens, which were filled with thick piles (~1,000 m) of felsic volcanic and pyroclastic rocks interbedded with mudstones (Oishi Formation) [31, 36]. In contrast, the Nishikurosawa and Sugota formations, which are only 30–400 m thick, are distributed on the Oga and Dewa ridges, respectively, with unconformable contacts with the underlying volcanic successions, and consist of shallow-marine sandstones, calcareous sandstones, and conglomerates, which were deposited on a shelf environment (Figure. 3)[22, 27]. These lithofacies and thickness contrasts are likely attributed to differential subsidence between horsts and grabens during the rifting under extensional tectonics [37]. Within the Aosawa and Kuroko grabens, syn-rift volcanism commenced at around 16.5 Ma and lasted until 13.5 Ma, prolonged than that in outer arc and in other inner arc areas in northeast Hounshu [33-34, 38]. Syn-rift volcanism (16.5 – 13.5 Ma) is further divided into early syn-rift volcanism (16.5 ~ 15 Ma) characterized by basaltic volcanism in the Aosawa graben and late syn-rift volcanism (15 ~ 13.5 Ma) characterized by felsic volcanism in the Kuroko graben [33, 38]. The Kuroko ore deposits were formed at the end of syn-rift stage at around 13.5 Ma, which may have been caused by the tectonic conversion from a back-arc to an island-arc setting [33, 36]

3.3. Evolution of the Sea of Japan rift system

The Sea of Japan rift system referred herein includes all component rift systems such as the Japan and Yamato basins, and the Eastern Japan Sea Rift System associated with the opening of the Sea of Japan. Evolution of the Sea of Japan rift system is summarized as follows. Incipient rift system commenced at around 35 Ma within a narrow zone along the present eastern margin of the Sea of Japan. Geochemical features of volcanic rocks similar to continental rifts suggest a rifting in the Eurasian continental arc. During the late incipient rifting in northeast Japan, the Japan Basin started opening at around 24 Ma followed by the opening of the Yamato Basin at around 21 Ma. The spreading of the Sea of Japan may have occurred in two stages. During the first stage, the Japan and Yamato basins spread from 24 – 21 Ma to 18 ~ 17 Ma. During the second stage, the upper basalts of the Yamato Basin may have accumulated at around 16 Ma [14]. In the meantime, rapid intra arc rifting started in the Eastern Japan Sea Rift System and surrounding areas at around 18 Ma and subsidence rate attained maximum (> 1km/m.y.) at around 16 – 15 Ma with the formation of graben fill basalts in the Akita and Niigata basins. Synrift volcanism and areas of intense subsidence migrated toward east and lasted until 13.5 Ma mainly in the

Kuroko graben along the eastern margin of the Eastern Japan Sea Rift System with minor amount of basalt activity in the Aosawa graben. The overall evolution of the Sea of Japan rift system may be interpreted as temporal progression from core-complex mode to wide-rift mode to narrow-rift mode by a simplified model of crustal extension [39]. Incipient rift system (35 – 24 Ma) may be assigned to core-complex mode while spreading of the Sea of Japan (24 – 15 Ma) is assigned to wide-rift mode. During the late stage of wide-rift mode (18-15 Ma), rifting extended to the Eastern Japan Sea Rift System. Late syn-rift system (15– 13.5 Ma) developed mainly in the Kuroko graben at the eastern margin of the Eastern Japan Sea Rift System and may be assigned to narrow-rift mode because the lithosphere became cold and strong [37, 39].

4. Intra-arc basin development in response to tectonic events: A case study from the Ou Backbone Range

This section focuses on the post-rift tectonic events based on a case study of intra-arc basin development from the Yuda Basin in the Ou Backbone Range (Figure. 2).

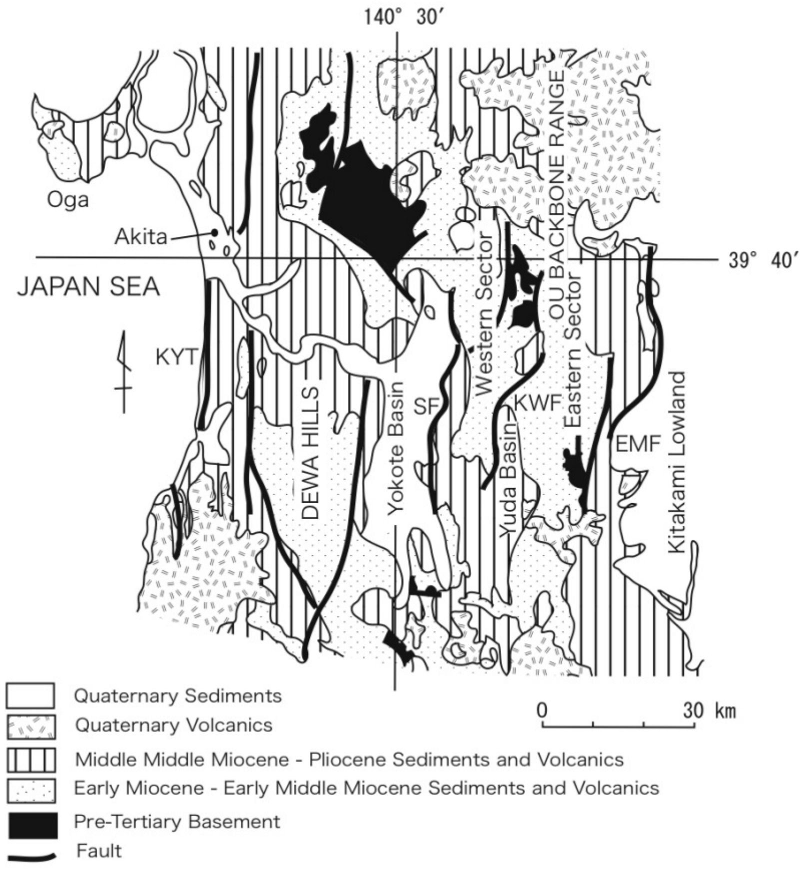

Figure 4. Index map of the Akita Basin and the Ou Backbone Range showing regional geology. SF: Senya Fault, KWF: Kawafune-Warikurayama Fault, EMF: Eastern Marginal Fault, KYT: Kitayuri Thrust.

4.1. Geology of the Yuda Basin

The Yuda Basin is located in the axis of the northern part of the Ou Backbone Range (Figure. 1). This portion of the Backbone Range is divided into western and eastern sectors, the altitudes of which are up to 1,000 m and 9,00 m, respectively. The two sectors are bounded by three fault systems; the Senya (bounding the western margin of the western sector), the Kawafune-Warikurayama (dividing the two sectors) and the Eastern Marginal Fault systems, from west to east (Figure. 4). The Kawafune-Warikurayama and Eastern Marginal Fault systems were originally formed as normal faults bounding eastern margins of half grabens during Early - Middle Miocene rifting/backarc opening. These normal faults have been reactivated (inverted) as thrusts during post-rift stages, resulting in a "pop-up" uplift of the present Ou Backbone Range (Figure. 4) [40]. The Yuda Basin is a depression between the eastern and western sectors (Figure. 4), and is 15 km long in a N-S direction and 5 km wide in the E-W direction. Pre-Tertiary basement rocks and Early – Middle Miocene syn-rift volcanic rocks are distributed in the axis of the western and eastern sectors, while Middle Miocene to Pleistocene marine and non-marine deposits are distributed in the surrounding lowlands including the Yuda Basin (Figure. 4).

The stratigraphy of the study area is divided into the Oishi, Kotsunagizawa, Kurosawa, Sannai, Hanayama and Yoshizawa Formations in ascending order (Figure. 5). The basin structure shows a simple synclinal structure bounded by the Kawafune-Warikurayama Fault at its western margin (Figure. 6). The deposits become younger toward the basin center, reflecting a synclinal structure.

The Oishi Formation consists of syn-rift volcanics, volcaniclastic rocks and mudstone. This formation had been formed by syn-rift felsic volcanism within half-grabens bounded by the Kawafune-Warikurayama and Eastern Marginal Fault systems in the Kuroko graben during the syn-rift stage (16 – 13.5 Ma) (Figure. 3).

The Kotsunagizawa Formation consists of mudstone, fine-grained sandstone and felsic tuffs. The formation conformably overlies the Oishi Formation and intercalates the Okinazawa basalt member, consisting of basaltic tuff breccia and lapilli stone in the middle part of the formation (Figures 5, 7). The uppermost several meters of the formation grades from mudstone to sandstone upward and is overlain by the Ochiai volcanic breccia bed (OB: 7 m thick)[43]. The age of the Kotsunagizawa Formation spans from 13.5 Ma to ca. 12 Ma based on biostratigraphy and fission-track datings (Figure, 8)[30-31, 45].

The Kurosawa Formation consists of shallow marine sandsone and tuffaceous mudstone with intercalations of felsic tuffs and conglomerate. Three key tuff beds, the Tsukano Tuff beds (TN: 20–50 m thick), Ohwatari Tuff beds (OW: 25 m thick) and the Torasawa Tuff bed (TS: 3–15 m thick) are intercalated in the middle and upper parts of the formation, respectively [30, 43] (Figures. 5-7). The Kurosawa Formation unconformably overlies the Kotsunagizawa Formation in the eastern margin of the Yuda Basin. The age of the Kurosawa Formation in the Yuda Basin was dated to be 9-6.5 Ma based on fission-track dating with a notable age gap of 2 – 3 Ma at the unconformity between the Kotsunagizawa and the Kurosawa formations (Figure. 8) [31, 43]. The Kurosawa Formation is, however, continuous from the Kotsunagizawa Formation in the western margin of the Yuda Basin (Core 41PAW-1 in Figure. 7) and in the western sector (Figure. 7).

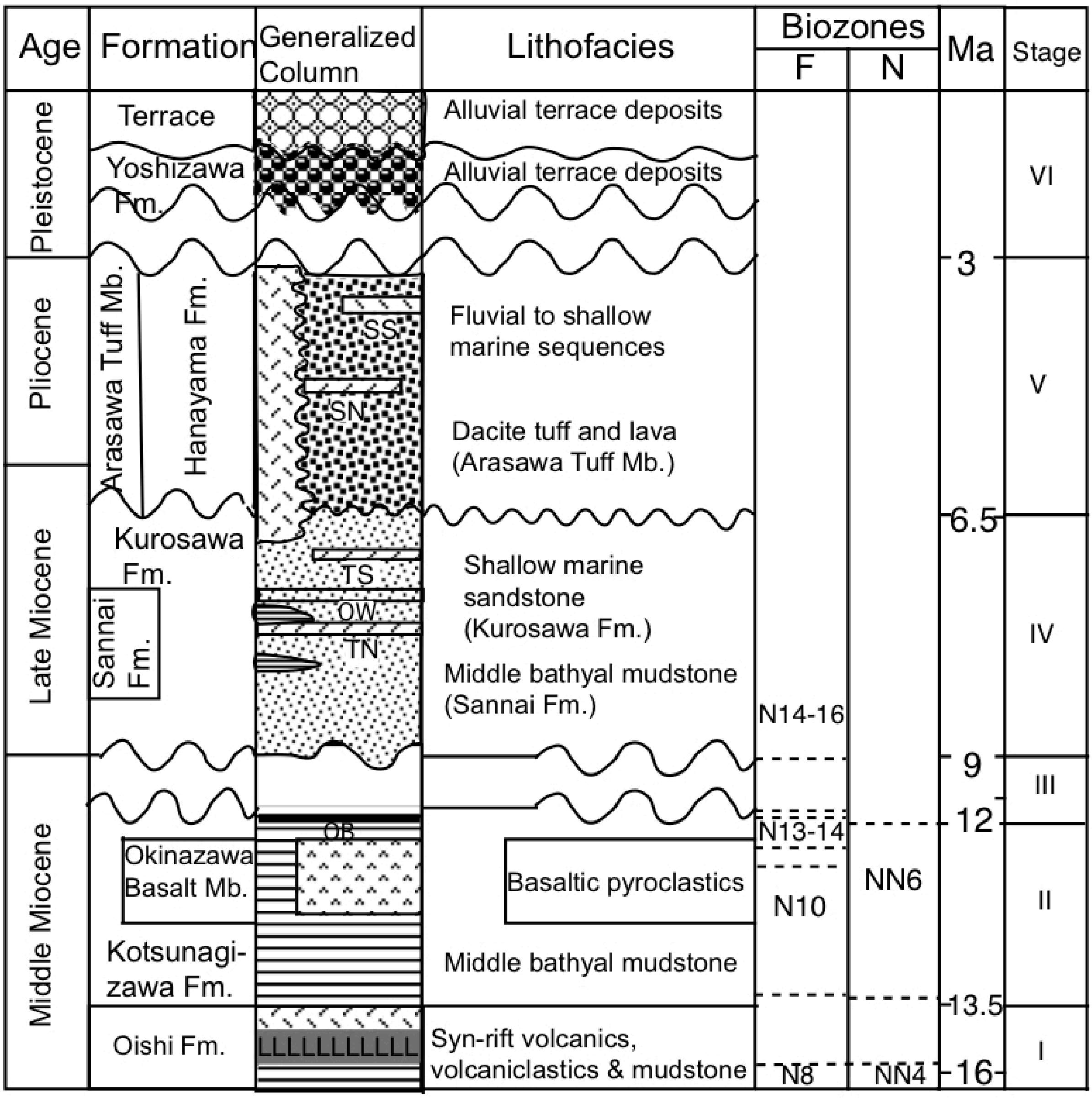

Figure 5. Generalized stratigraphy of the Yuda Basin with approximate ages and tectonic stages (I – VI). Key tuff bed names: SS: Sasoh tuff, SN: Sawanakagawa tuff, TS: Torasawa tuff, OW: Ohwatari tuff, TN; Tsukano tuff. F: planktonic foraminifer zonation after [41]. N: nannofossil zonation after [42]. Modified from [43].

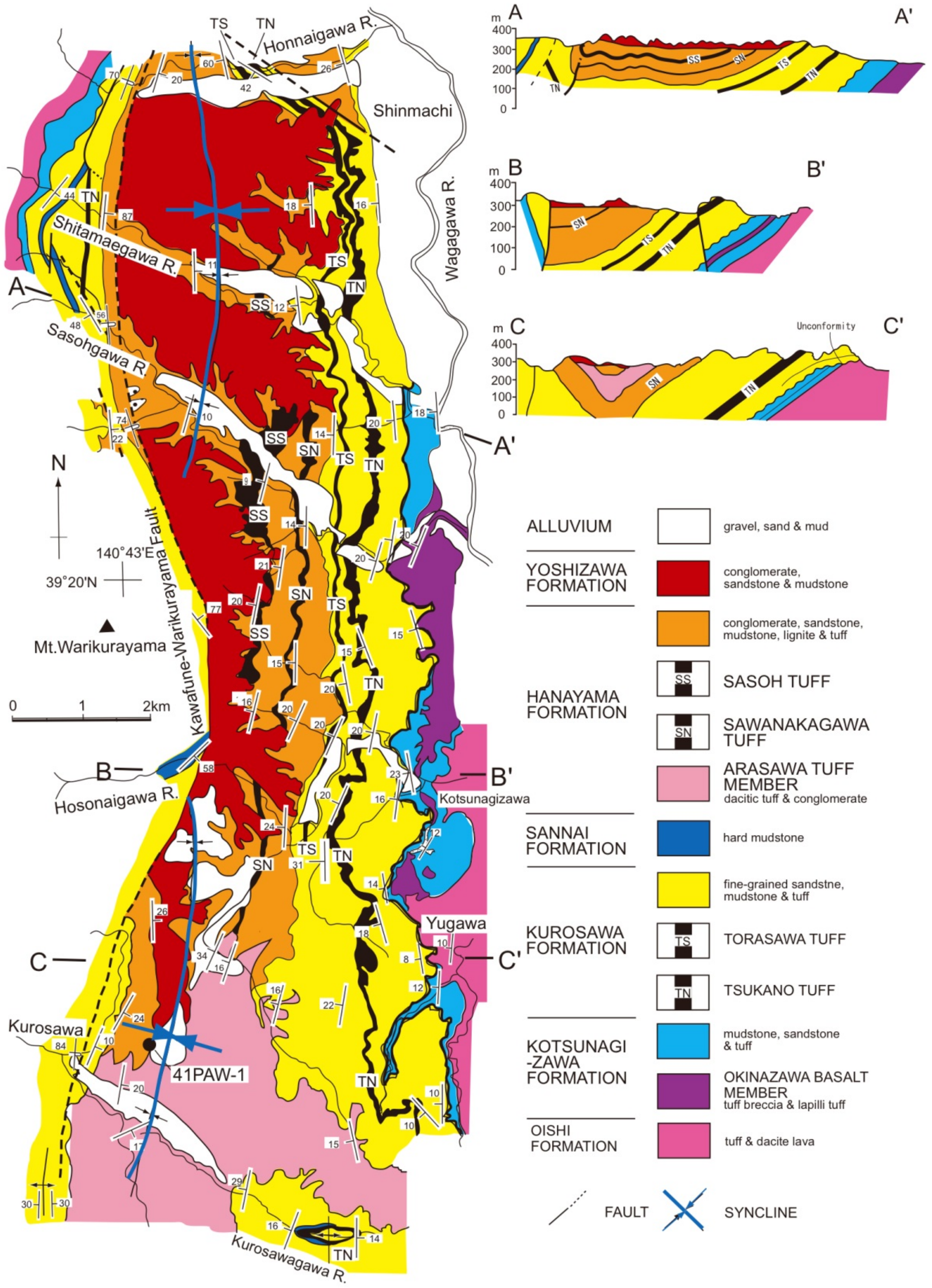

Figure 6. Detailed geological map and geological cross-sections of the Yuda Basin. A-A', B-B' and C-C' are lines of cross-sections. 41PAW-1 denotes the location of the drilling hole [44] in the southwestern part of the Yuda Basin. Modified from [31]

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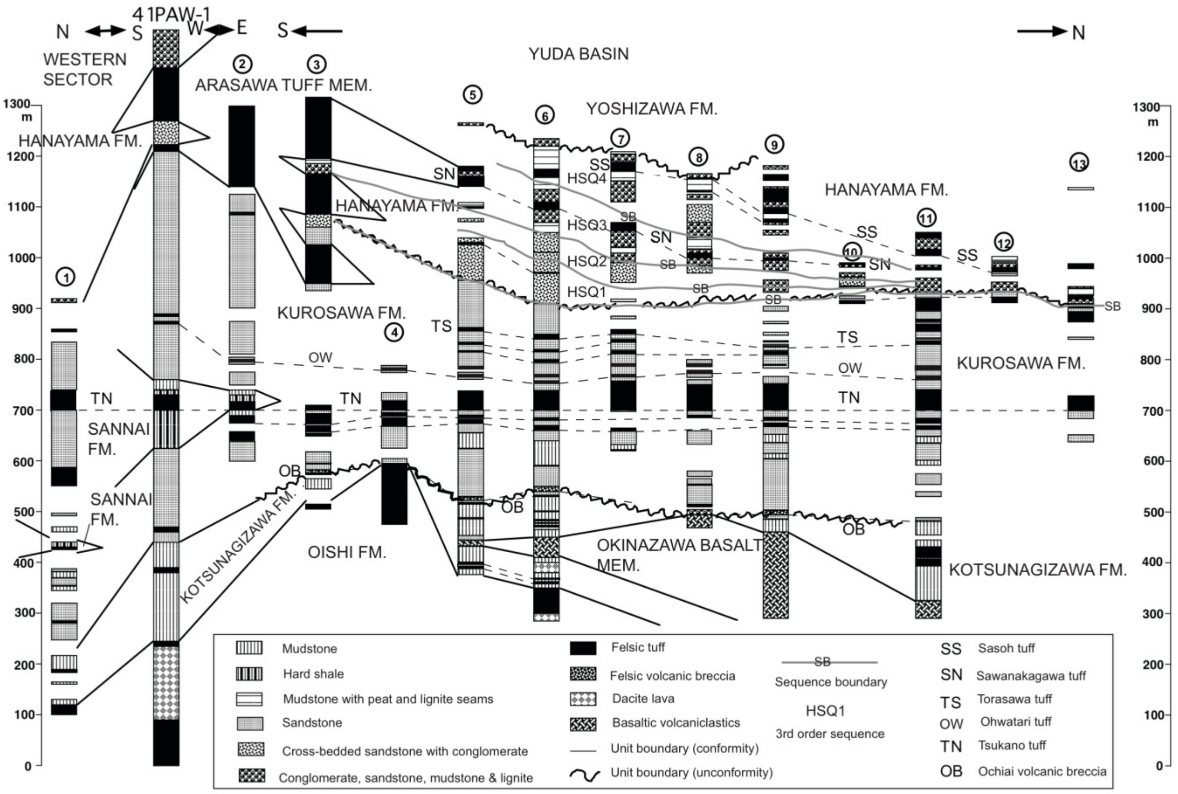

Figure 7. Correlation of lithologic columns in the Yuda Basin (modified from [43]). Numbers in open circles denotes locations of columns shown in Figure 10.

The Sannai Formation consists of alternating hard mudstone and black shale associated with felsic tuffs. The formation interfingers with the Kurosawa Formation (Figure. 7) and is distributed in the western sector and in the southern margin of the Yuda Basin, while it is absent in the eastern margin of the Yuda Basin (Figure. 7). This formation is equivalent to the Onnagawa Formation, a siliceous shale unit in the center of the Akita Basin (Figures. 2&3).

The Hanayama Formation distributed along the syncline in the axis of the Yuda Basin is divided into a main part and the Arasawa Tuff Member distributed in the southern Yuda Basin (Figure. 7). The main part of the formation unconformably overlies the Kurosawa Formation, and consists of fluvial, marsh, deltaic and shallow marine deposits [30-31]. The Hanayama Formation includes four 3rd order depositional sequences overlapped by numerous high frequency sequences (Figure. 7). Two key tuff beds, the Sawanakagawa Tuff beds (SN: 4–30 m thick) and the Sasoh Tuff bed (SS: 10–50 m thick) are intercalated in the middle and upper Hanayama Formation (Figures. 5-7)[30]. The Arasawa Tuff Member chiefly consists of dacite pumice tuff associated with dacite lava and conglomerate. This member interfingers with upper part of Kurosawa Formation and also with the main part of the Hanayama Formation in the southern Yuda Basin (Figures. 5-7). The age of the main part of the Hanayama Formation

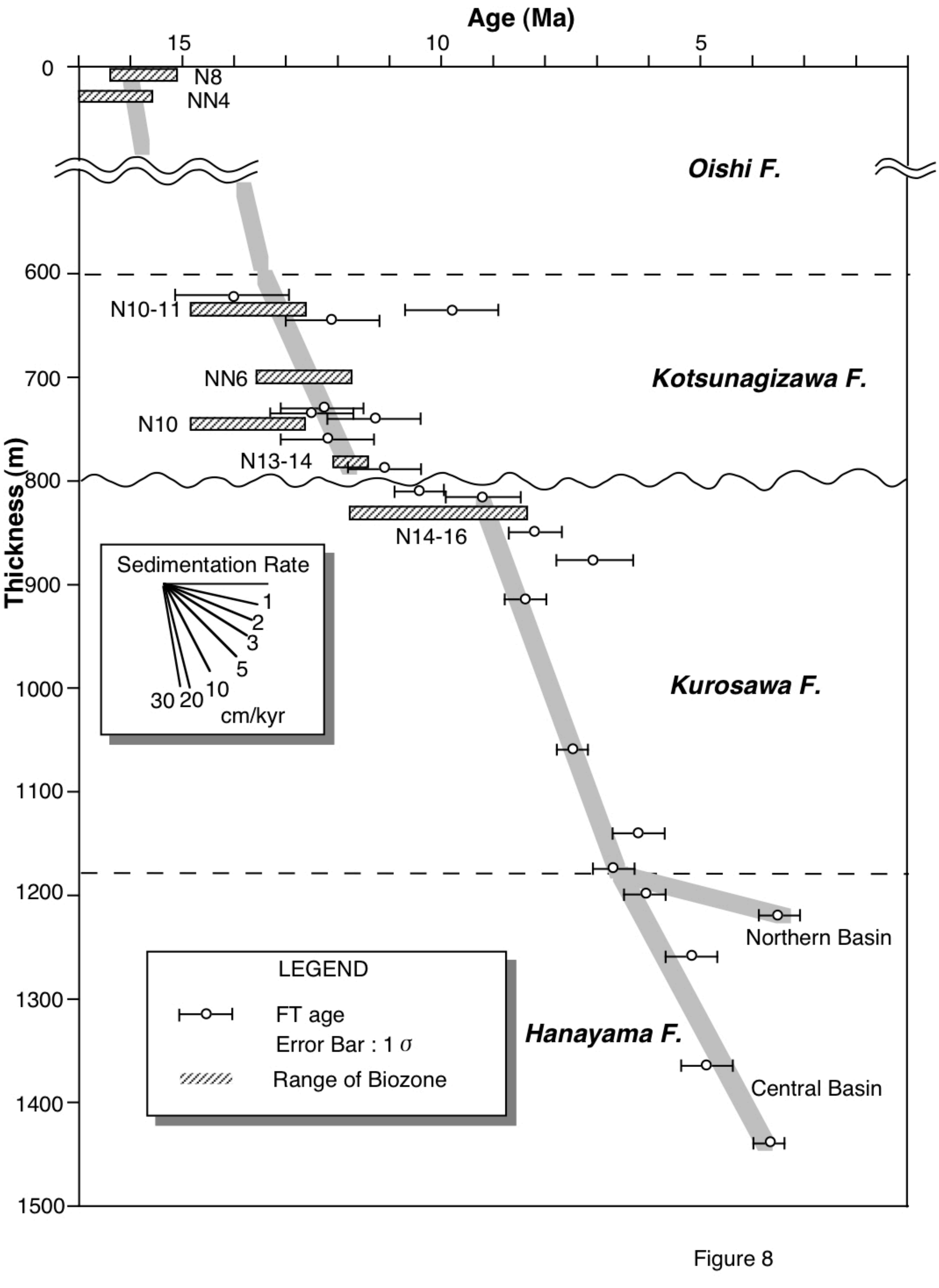

Figure 8. Age-thickness diagram in the Yuda Basin after [31].

was estimated from 6.5 to less than 3 Ma based on fission-track dating (Figure 8) [30, 45] although the upper boundary has not yet been precisely dated.

The Yoshizawa Formation overlies the Hanayama and Kurosawa formations with a notable angular unconformity and is distributed in the western part of the Yuda Basin along the Kawafune-Warikurayama Fault (Figure. 4). This formation consists of alternation of conglomerate, sandstone and mudstone associated with lignite seams and peat. This formation may have been deposited on alluvial fans that were formed along the eastern margin of the western sector [30, 45]. The Yoshizawa Formation may be early Pleistocene in age based on the stratigraphic relationships between this formation and the Hanayama Formation and fluvial terraces [30].

4.2. Styles and origins of unconformities in the Yuda Basin

Three major unconformities were formed around 10 Ma, 6.5 Ma and after 3 Ma in the eastern margin of the Yuda Basin. These unconformities have significant implications for post-rift tectonics not only in the Ou Backbone Range but also in Northeast Japan and thus the styles and origins of these unconformities are described and discussed herein

4.2.1. 10 Ma unconformity

The unconformity between the Kotsunagizawa Formation and the Kurosawa Formation at around 10 Ma is a partial unconformity within the basin and shows significant horizontal change in terms of the style [31]. The unconformity apparently eroded the Kotsunagizawa Formation at two hills in the eastern margin of the Yuda Basin (Figures. 5-7). At the road cut section north of the Kotsunagizawa (Figure. 6), the unconformity eroded the upper part of the Kotsunagizawa Formation and the Okinazawa basalt member and the shallow marine sandstone of the Kurosawa Formation overlies the Okinazawa basalt member with the basal conglomerate bed composed of boulders of basalts eroded from the underlying Okinawaza basalt member (Figure. 9)[43]. In this section, the upper 40 meters of the Kotsunagizawa Formation and the Ochiai volcanic breccia bed were eroded as well (Figure 9). The lower Kurosawa Formation in this section yields fission-track ages of 9.2 and 7.1 Ma (Figure. 10), significantly younger age than the top of the Kotsunagizawa Formation dated at around 12-11 Ma (Figure 8)[31]. This suggests an age gap of 3 – 2 Ma interval at the unconformity (Figure 8) [43].

The amount of erosion varies over a short distance, but tends to increase where the boundary shifts to the east, towards the uplift axis of the eastern sector (Figures. 6 & 10). The maximum amount of the erosion attains more than 100 m at another hill west of Yugawa (Figure. 6), where the unconformity eroded the entire Kotsunagizawa Formation and the Kurosawa Formation directly overlies dacite tuff breccia of the Oishi Formation (Figure. 10). At the boundary, the Kurosawa Formation onlaps the Oishi Formation with lower angle dip of 8° W than the underling Oishi Formation with a dip of 16° W (Figure. 6)[43]. This indicates angular unconformity between the Oishi and Kurosawa formations. The basal 5 m of the Kurosawa Forma-

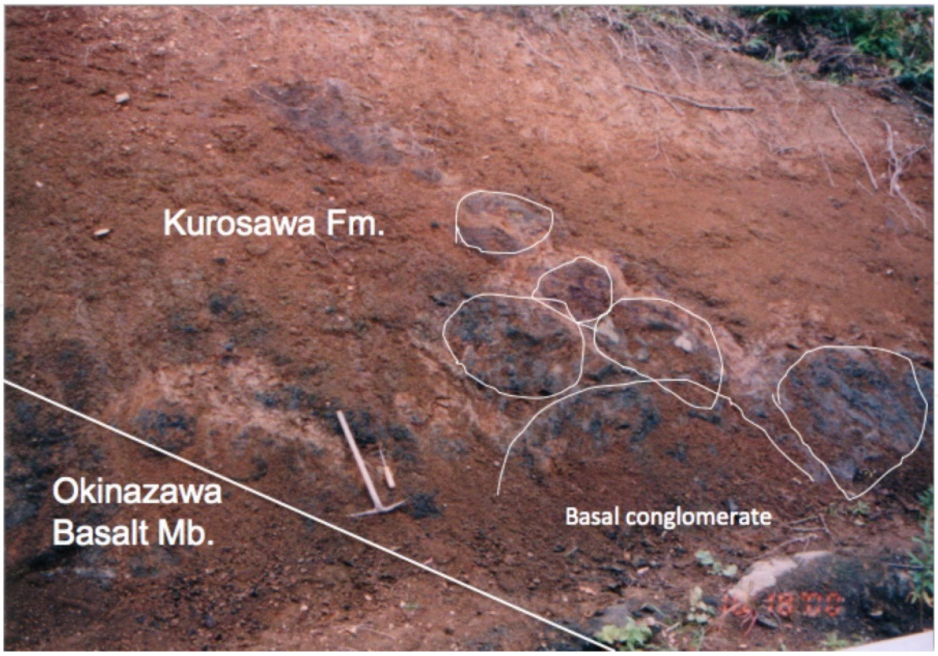

Figure 9. A photo showing the unconformity between the Kotsunagizawa and Kurosawa formations in column K of Figure 10. White circles denote basal conglomerates derived from the Okinazawa Basalt Member of the Kotsunagizawa Formation. Modified from [43].

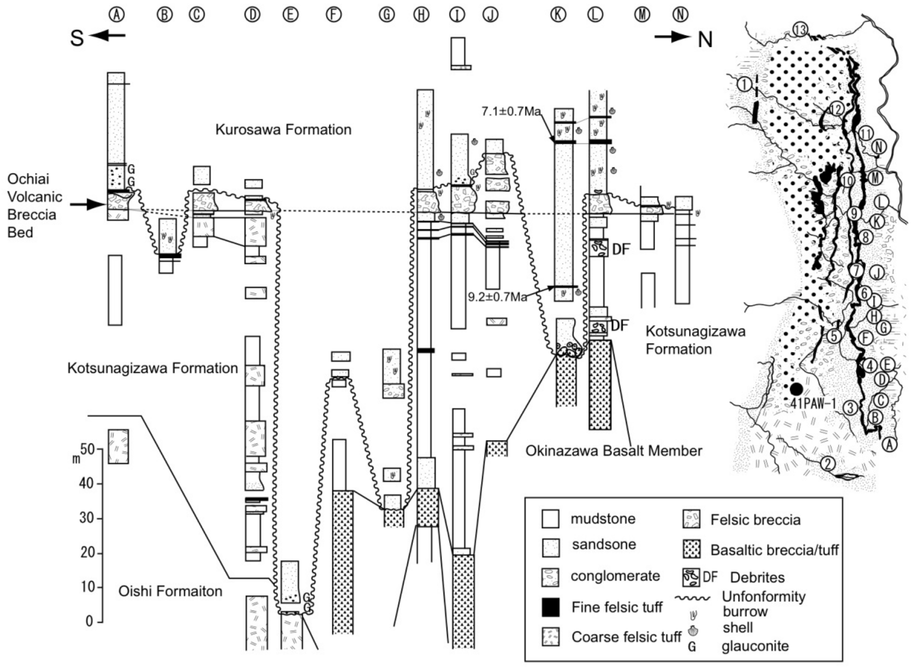

Figure 10. Detailed correlated columns showing the unconformity between the Kotsunagizawa and Kurosawa formations in the Yuda Basin. Inset shows locations of lithologic columns in Figure 7. Modified from [31].

tion in this section contains glauconite sandstone (Figure. 10), indicating that significant hiatus or low sedimentation rate event took place at the time of the unconformity[46].

In other parts of the eastern margin of the Yuda Basin, however, this boundary is a paraconformity. Bathyal mudstones in the upper part of the Kotsunagizawa Formation grade upward into shallow marine sandstones in the top ~10 m of the formation. The Ochiai volcanic breccia bed [43] of up to 15 m in thickness rests on sandstones at the top of the formation and is a distinct key correlation bed in the uppermost Kotsunagizawa Formation (Figure. 10). The breccia bed was covered by a fine tuff bed (<1 m in thickness), which is intensely burrowed and bioturbated. This fine tuff bed is sharply overlain by shallow-marine, massive sandstone of the basal Kurosawa Formation (Figure. 10). The contact between intensely burrowed fine tuff at the top of the Kotsunagizawa Formation and the shallow marine sandstone of the basal Kurosawa Formation can be traced at the same stratigraphic position <1 m above the Ochiai breccia bed over 10 km from the south to the north except for the two hills where the unconformity incised deeply the underlying strata (Figure. 10). Although no significant erosion at the boundary was suggested, the significant age gaps between two formations indicate hiatus or slow deposition at the boundary.

In the western margin of the Yuda Basin (Core 41PAW-1 in Figures. 6, 7) and in the western sector (Figure. 7), the boundary between the Kotsunagizawa and Kurosawa formations is continuous. Correlation between the Yuda Basin and the westerm sector (Figure. 7) indicates that the equivalent time interval of unconformity in the eastern Yuda Basin corresponds to the deposition of the lower Kurosawa Formation in the western sector. Based on the above consideration, there was a westward submarine paleo-slope in the basin during ~12–9 Ma. In summary, the unconformity dated around 10 Ma was attributed to the uplift and westward tilting of the eastern sector, starting around 12 Ma and followed by the cessation of subsidence until 9 Ma and by subsequent onlapping of the Kurosawa Formation from west because of the resumption of the subsidence. The overall observation suggests that the eastern sector started to uplift and emerge earlier than the western sector. However, the western sector may have also slightly uplifted in this stage because the middle bathyal Kotsunagizawa Formation changed upward into the upper bathyal to outer shelf Kurosawa Formation (Figure 7).

4.2.2. 6.5 Ma unconformity

The base of the main part of the Hanayama Formation was defined by the first occurrence of gravelly sandstone deposited in fluvial channels. The relationship between the Kurosawa and Hanayama formations is thus unconformity formed by fluvial erosion. Nakajima et al. [30] interpreted the unconformity as a sequence boundary formed by a relative sea-level lowering. Although the erosion by the unconformity is not significant (Figure. 7) and no age gap was suggested between two formations [30-31], this unconformity may have reflected major tectonic changes at around 6.5 Ma for the following reasons [43]. After the 6.5 Ma unconformity was formed, 4th or 5th order high frequency depositional sequences comprising of fluvial, deltaic and shallow water deposits with significant amount of conglomerate were deposited [30]. This contrasts with the Kurosawa Formation, which consist totally of shallow marine sandstones. This suggests that supply of coarse sediments began to exceed the accommodation space possibly because of the uplift of surrounding backbone range. In addition, this unconformity marks the major changes from uniform sedimentation rate in north-south direction in the Kurosawa Formation to differential sedimentation rate in the Hanayama Formation (Figure. 7). Those changes in depositional style may have resulted from tectonic conversion from extension to compression as discussed later.

4.2.3. < 3 Ma unconformity

The angular unconformity between the Hanayama and Yoshizawa formations is the most significant structure among three unconformities described herein. The Kurosawa and Hanayama formations form a syncline with dips of 20 – 30 ° and are unconformably overlain by almost flat Yoshizawa Formation (Figure. 6). The formation of syncline and subsequent unconformity may have reflected uplift of the eastern and western sectors of the Ou Backbone Range during and after deposition of the Hanayama Formation. The unconformity was formed after 3 Ma although the precise age has not been dated.

4.3. Evolution of the Yuda Basin since Middle Miocene

Evolution of the Yuda Basin since Middle Miocene is summarized here, based on correlation of tuff beds and unconformities (Figures 7 & 10) and basin subsidence analysis (Figure. 11). Tectonic history of the Yuda Basin and the surrounding Ou Backbone Range was divided into six tectonic stages according to the basin subsidence pattern (Figure. 12)[31].

4.3.1. Stage I (Syn-rift stage; 16-13.5 Ma)

This stage was characterized by a rapid subsidence (600 m/m.y.) with accumulation of thick volcanic and volcaniclastic successions in the Yuda Basin (Figure. 11). The amount of tectonic subsidence attains no less than 1,000 m even if altitude of sea-level rise at around 15 Ma [47] is subtracted (Figure 11). The Eastern Marginal Fault and Kawafune-Warikurayama Fault systems may have been activated as normal faults and formed half grabens in the eastern and western sectors, respectively as a result of crustal stretching under extensional tectonics during Early-early Middle Miocene rifting (Figure 12) [40].

4.3.2. Stage II (Post-rif transition stage; 13.5–12 Ma)

This stage was characterized by the cessation of syn-rift volcanism and accumulation of hemipelagic sediments at a slower rate (10 cm/k.y.) in the Yuda Basin (Figures. 8 & 12). The subsidence reconstruction of the Yuda Basin (Figure. 11) indicates subsequent reduction of tectonic subsidence rate although precise estimation is difficult due to unreliable estimation of paleo-depth under bathyal environment.

4.3.3. Stage III (partial inversion stage: 12–9Ma)

This stage was represented by temporal uplift of the Ou Backbone Range and associated unconformity in the Yuda Basin. The amount of tectonic uplift in the Yuda Basin at around 12-11 Ma was estimated at more than 500 m. This uplift was followed by a cessation of

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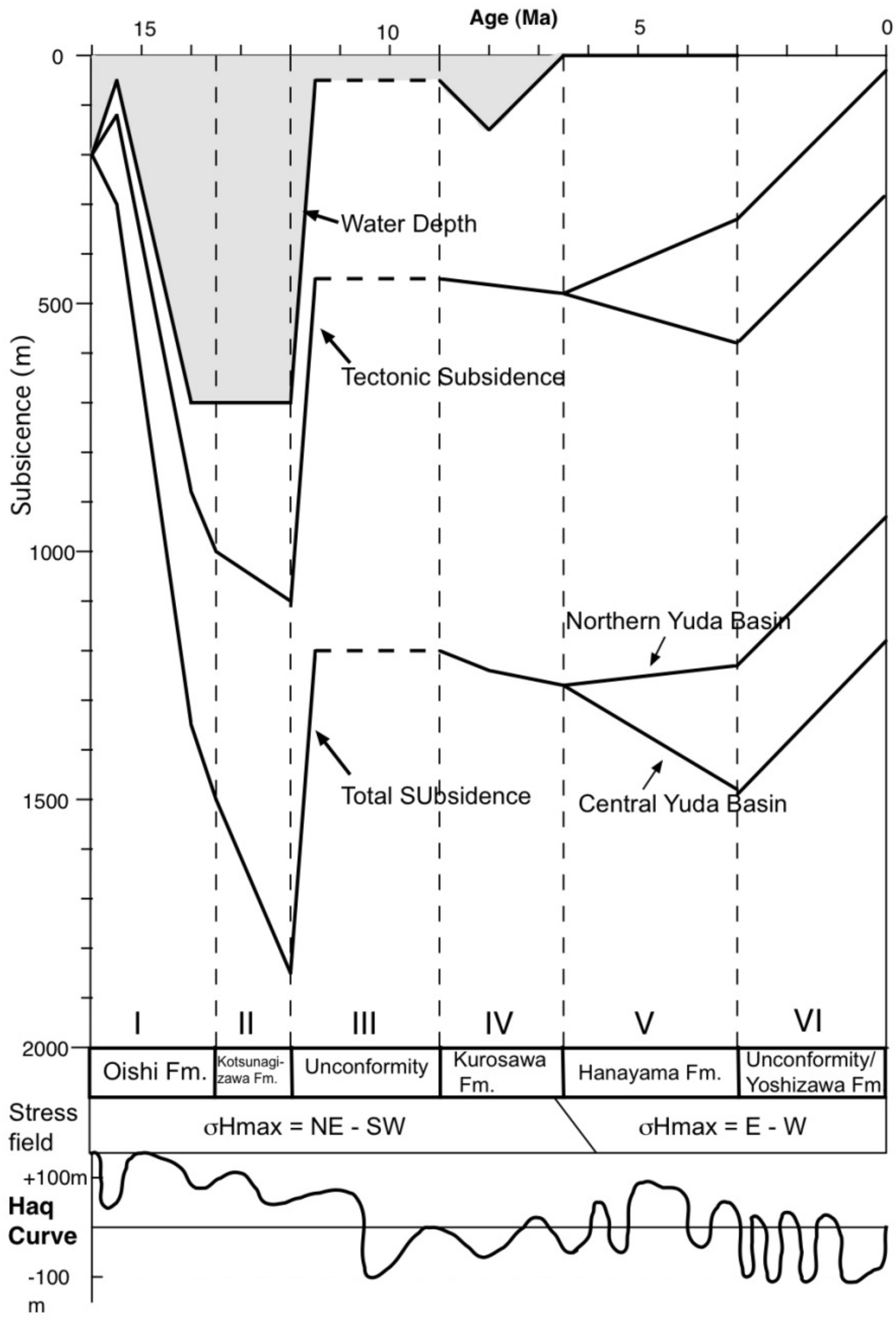

Figure 11. The total and tectonic subsidence curves of the Yuda Basin with tectonic stages (I-VI), inferred stress field [30] and a eustatic curve proposed by [47]. Modified from [31, 43].

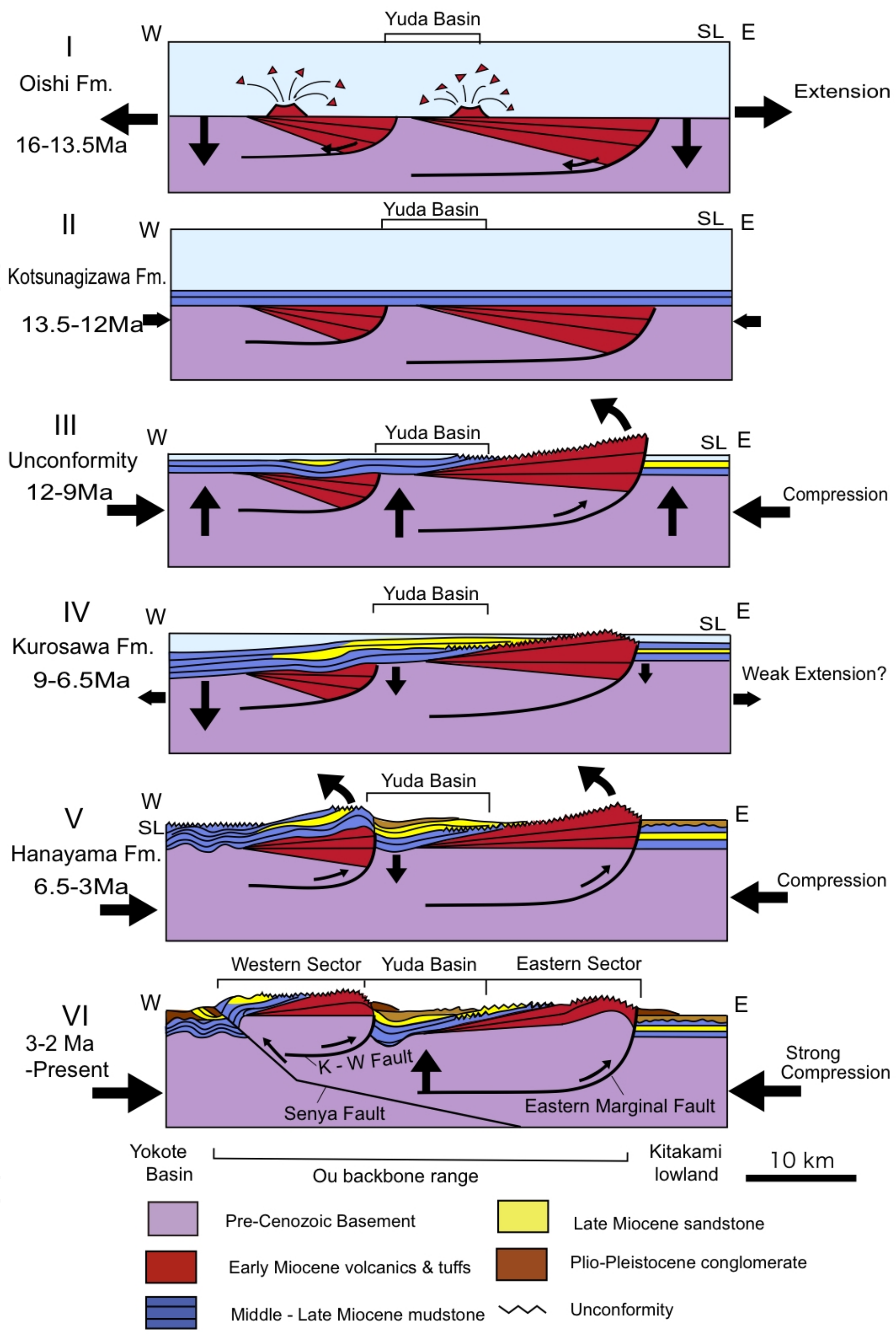

Figure 12. A cartoon illustrating the tectonic evolution of the Ou Backbone Range (modifed from [31]). K-W Fault: Kawafune-Warikurayama Fault. See text for explanation.

subsidence and uplift within the basin until about 9 Ma (Figure. 11), leading to a hiatus in the eastern margin of the basin. Correlation of key beds and unconformities (Fig. 7) clearly demonstrates that intensely uplifted areas at the eastern margin of the basin were more subject to intense truncation. The uplift was more intense in the eastern sector than in the western sector, and was accompanied by the westward tilting of the eastern sector (Figure 12). The eastern sector began to emerge and became a sediment source to the surrounding basins by subaerial erosion. The uplift of the eastern sector of the backbone range may be attributed to a subsequent inversion of a half graben formed during Stage I because of an increase in horizontal compressional stress (Figure. 12)[31]. The western sector of the backbone range also uplifted from middle bathyal to shelf environments, although the bounding normal fault does not seem to have been inverted. This stage was also characterized by reduced volcanic activity around the basin as suggested by scarcity of tuffs intercalated in the lower Kurosawa Formation in the western sector (Figure. 7).

4.3.4. Stage IV (subsidence stage: 9–6.5 Ma)

This stage was characterized by slow subsidence and deposition of sand in shallow marine environments in the Yuda Basin. The subsidence resumed at around 9 Ma and shallow-marine sandstones were deposited over the unconformity in the eastern margin of the Yuda Basin, while the eastern sector of the backbone range remained as a sediment source. The lower Kurosawa Formation thins in intensely uplifted and truncated areas such as west of Yugawa (Figures. 6 & 7), which suggests that the Kurosawa Formation onlaps these uplifted topographic highs. While subsidence rates in the eastern part of the Yuda Basin were uniform in north-south direction (Figure. 7) within the magnitude of eustatic sea-level fluctuation (Figure. 11), the western sector subsided more rapidly. Total tectonic subsidence in the western Yuda Basin was estimated at least 600 m during this stage from the thickness of the Sannai and upper Kurosawa Formations in the core 41PAW-1 (Figure. 7)[31]. This stage was also characterized by increased felsic volcanism, as represented by increased felsic tuffs such as the Tsukano and Torasawa Tuff beds in the Kurosawa Formation in the Yuda Basin (Figure. 7). Occurrence of Northeast-Southwest-trending minor normal faults in the Kurosawa Formation indicates Northwest-Southeast-trending extensional stress field [30], which resulted in regional slow subsidence.

4.3.5. Stage V (basin inversion and compression stage: 6.5 – 3~2 Ma)

Stage V represents differentiation of uplifted and subsided areas within the Yuda Basin. Subsidence pattern changed from the preceding stage IV, and the northern Yuda Basin turned into an uplifted area (Figure. 11). Moreover, conglomeratic deposits within the basin indicate uplift of the surrounding mountains. The high frequency depositional sequences were developed in this stage, suggesting that supply of coarse sediments began to exceed the accommodation space. The western sector may have also uplifted because marine environments gradually retreated from the Yuda Basin by ~4 Ma [30], which indicates the emergence of the western sector by that time. The Northeast-Southwest trending minor normal faults found in the Kurosawa Formation disappeared at the base of the Hanayama Formation (~6 Ma) in the Yuda Basin [30]. After 6 Ma, only North-South-trending reverse faults were formed in the Yuda Basin. The observation indicates that stress changed from extension to E-W compression at the beginning of this stage. The compressive stress field resulted in basin inversion and uplift of both the eastern and western sectors of the Ou Backbone Range (Figure. 12).

4.3.6. Stage VI (Intense compression stage; 3~2 Ma - Present)

Stage VI represents the uplift of the whole Backbone Range and the formation of an angular unconformity between the Hanayama and Yoshizawa Formations within the Yuda Basin (Figures 11, 12). The uplift of the Backbone Range during this stage resulted from "pop-up" of the sectors of the Backbone Range by activation of the Senya Fault system as well as reactivation of the Eastern Marginal Fault and Kawafune-Warikurayama Fault systems under intense compression (Figure. 12)[40].

5. Late Cenozoic tectonic events in northeast Japan

In this section, Late Cenozoic tectonic events deduced from developments of other fore-arc, intra-arc and back-arc basins across the northeast Japan transects were correlated with those in the Yuda Basin (Figure 13). Based on the results, Late Cenozoic tectonics in northeast Japan was clarified and was divided into seven stages from 0 to VI as described below. Then, the author will discuss the origin of regional tectonic events. The reviews presented herein could provide a revised Late Cenozoic tectonics model in northeast Japan.

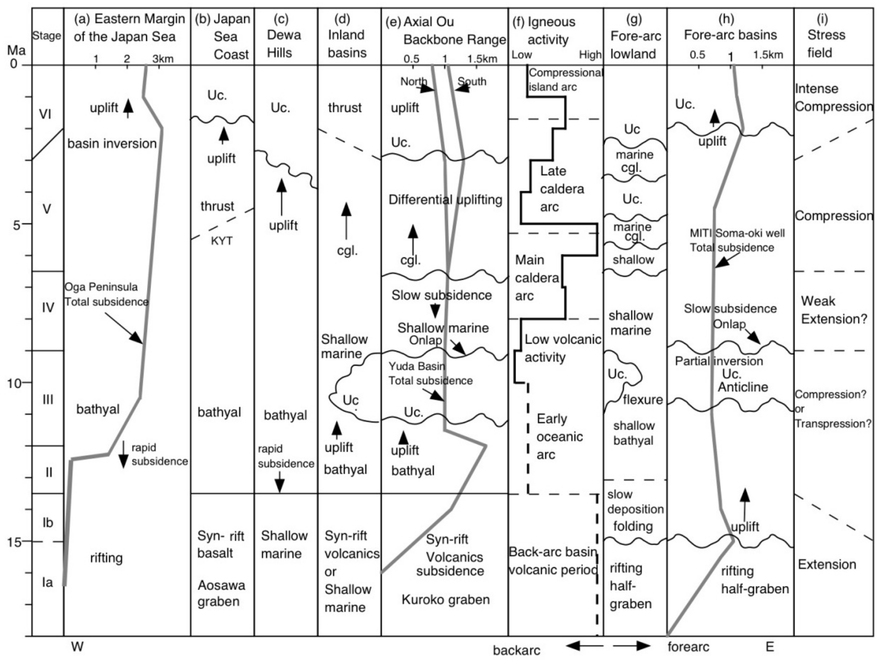

Figure 13. A compiled chart showing correlation of tectonic movements across northeast Japan transect since 18 Ma (revised after [31]). Stages I-VI are tectonic stages defined in this study. Uc.: unconformity, cgl.: conglomerate. Compiled after (a) [48] revised after new age model of [22], [49]; (b) [27-29]; (c) [27-29, 32]; (d) [32, 51-54]; (e) [31, 38, 55-56]; (f) [16, 33, 57-58]; (g) [56, 59-65]; (h) [31, 66-68]; (i) [31]

5.1. Stage 0 (Incipient rift stage; 35 – 20 Ma)

Stage 0 represents formation of the incipient rift system along the eastern margin of the Sea of Japan as already noted in section 3. Rift basins in this stage were relatively small and marine incursion was limited within a narrow zone along the present Sea of Japan coast [21]. The timing of marine incursion varied among rift basins [21], suggesting that individual marine basin was relatively short lived. The incipient rift basins had not directly developed into later syn-rift large basins, but were interrupted by regional unconformity during the late Stage 0

5.2. Stage I (Syn-rift stage: 20 - 13.5 Ma)

Syn-rift stage has been divided into two substages: IA and IB on the basis of tectonic conversion from wide rift mode to narrow rift mode [39], associated with a stress change from regional extension to coexistence of extension and compression at around 15 Ma.

5.2.1. Substage IA (20 – 15 Ma)

After the incipient rifting stage, dacitic volcanics with some amount of basaltic volcanics together with conglomerates and sandstones deposited in terrestrial environments at around 20 Ma in the Akita Basin [22, 34]. These stratigraphic units may represent slow subsidence [22] prior to rapid rifting after 18 Ma as described in section 3.2. They were unconformably overlain by non-marine to marine successions deposited during rapid rifting. The stratigraphic units deposited during the rapid rifting (ca. 18 – 15 Ma) represent regional marine transgression in northeast Japan as a result of rapid subsidence under extensional tectonics associated with rifting [26, 69]. The equivalent stratigraphic units during Substage IA also deposited within rotated half grabens in the eastern margin of the Sea of Japan [49]. Rapid subsidence associated with rifting and half grabens also took place in the fore-arc side of northeast Japan [56, 66, 70] and in the Kanto Plain (Figure 1)[5, 71]. In terms of igneous activity, this stage was assigned to a backarc basin volcanic period (Figures, 3, 13)[33, 58] and was characterized by intense basaltic volcanism within rift grabens in the Akita and Niigata basins [34, 36, 58].

5.2.2. Substage IB (15 – 13.5 Ma)

Substage IB is characterized by shrinkage of rift zones and by uplift with a notable unconformity in fore-arc side and fore-arc basins in northeast Japan. Formation of half grabens in the Kanto Plain (Figure 1) was suddenly terminated by rapid uplift with formation of a notable unconformity (the Niwaya unconformity) at 15.3 – 15.2 Ma [5, 71]. Marine fine-grained sediments with low sedimentation rates unconformably overlie the successions in rotated half grabens [71]. This change in the style of subsidence and deposition was attributed to tectonic conversion from extensional to strong compression stress, followed by relatively quiet tectonics [71]. Sedimentation rates in the post rift successions over the Niwaya unconformity in the northern Kanto Plain had been suppressed until 14 – 12.5 Ma, suggesting compression lasted until ca. 13 Ma [64]. The Joban forearc basin (Figure 1) was also inverted and uplifted at around 15 Ma (Figure 13) with a notable unconformity being formed [68]. This tectonic change was accompanied by NW – SE trending folding along the coast of Sendai [56] and in the Joban Basin [68], which suggests that the stress field in the forearc switched to a compressional or transpressional regime at around 15 Ma (Figure 13). The shallow marine successions in Substage IB in the Sendai Plain (Figure 1) comformably overlie the underlying conglomerates of ca. 16 Ma [60, 63]. However, sedimentation rates had been suppressed until about 13 Ma in the glauconite bed at the base of the shallow marine successions [60], similar to the observation in the northern Kanto Plain. Rifting associated with activities of half grabens also ceased at about 15 Ma in the Niigata Basin [69] and in the Uetsu district [26] within the Eastern Japan Sea Rift System (Figure 2). This suggests that extensional tectonics ended at ca. 15 Ma in the back-arc side as well as fore-arc side of northeast Japan. However, rifting associated with activities of half grabens lasted until 13.5 Ma in the Kuroko graben, as suggested by the rapid subsidence of the Yuda Basin (Figures 11 & 12) and voluminous felsic volcanism in the Kuroko graben [31, 33, 36]. Sato [72] also concluded that synsedimentary faults bounding the Aosawa and Niigata grabens had been active until 13.5 Ma based on the isopach map of Substage IB. These observations clearly indicate that extensional tectonics continued within some parts of the Eastern Japan Sea Rift System, particularly in the Kuroko graben as a narrow rift mode [39] during Substage IB. For this reason, both strong compression and extension coexisted in northeast Japan during Substage IB. The origin of strong compression in the fore-arc side of northeast Japan may be attributed to rapid counter-clockwise rotation of northeast Japan at around 15 Ma as a result of the opening of the Sea of Japan (See section 2). The rapid counter-clockwise rotation of northeast Japan might accelerate the relative convergence rate of the Pacific Plate at the Japan Trench, which must have resulted in increased compressional stress. Collision and transpressional movements of northeast and southwest Japan arcs along the TTL (Figure 1) as a result of differential rotation of both arcs [5] might also contribute to strong compression along the border of the two arcs such as in the Kanto Plain (Figure 1).

5.3. Stage II (Post-rift transition stage: 13.5 – 12 Ma)

Stage II was characterized by the cessation of all rifting including the Aosawa and Kuroko grabens, and by marine transgression associated with rapid subsidence of the Dewa and Oga ridges in the Akita Basin (Figure 3). Although this stage had been previously regarded as quiet thermal subsidence stage under neutral stress regime [56, 72, 73], the Dewa ridge subsided rapidly from inner sublittoral to middle bathyal environments at 13.5 Ma, followed by rapid subsidence of the Oga ridge from upper bathyal to lower bathyal environments at 12.3 Ma [22, 27]. This subsidence mode cannot solely be attributed to a post-rift thermal subsidence, because the subsidence rate far exceeded the estimated rate (~70 m/m.y.) of the thermal subsidence [31], and the timing of rapid subsidence was out of phase between the Dewa and Oga ridges. In terms of volcanism, Stage II was represented by a major change from back-arc basin stage to island-arc stage [33]. These changes in tectonic and volcanic styles suggest a stress change from extension to compression [31]. Subsidence resumed and sedimentation rates increased at about 13 Ma in the northern Kanto Plain and in the Sendai Plain, where deposition had been suppressed during the previous Substage IB because of strong compression [60, 64].

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5.4. Stage III (Partial inversion stage: 12 – 9 Ma)

Stage III was represented by the temporal uplift and associated unconformity caused by partial inversion in the Backbone Range. The temporal uplift and associated unconformity at around 10 Ma occurred not only in the Yuda Basin, but also in other sections of the Backbone Range from south to north along the axis of the Backbone Range [38, 55, 56]. Moreover, contemporary uplift and associated unconformity also occurred in inland basins and forearc lowlands along the western and eastern margins of the Backbone Range, respectively (Figure 13). For example, northwestern margin of the western sector at the eastern margin of the Yokote Basin (Figure 4), rapidly uplifted (>700 m) from middle bathyal to terrestrial environments with a notable unconformity at ~9 Ma [54]. Although the southern part of the western sector had not been inverted as described in Section 4.3.3, sedimentary successions show westward progradation from the backbone range to the Dewa Hills as a local response to excess sediment supply over the rate of creation of accommodation space (Figure 3). The contemporaneous uplift and resultant unconformity also took place at ~10 Ma in the margins of the Yonezawa and Aizu Basins (Figure 2), ~200 km south of the Yokote Basin [51, 53]. Notably, the amount of erosion at the unconformities increased toward the east (toward the Backbone Range), while deposition continued in the west of the basins [51, 53]. An angular unconformity was also formed at 11.5 – 9 Ma in response to the NW-SE trend flexure activity in the Sendai Plain (Figure. 1), east of the Backbone Range [62, 63]. Sedimentary successions in the northern Kanto Plain, east of the Backbone Range showed upward shallowing successions since 10 Ma, which was followed by an unconformity at ~9 Ma [64]. A regional angular unconformity (~11–9 Ma) occurred in the sedimentary sequence in the Joban forearc basin [67]. The basement subsidence reconstruction at the MITI Soma-oki well (Figure. 1) in the Joban forearc basin suggests uplift until this time interval (Figure 13). Seismic reflection profiles across the well show that erosional surfaces within this interval truncated the top of the Soma-oki anticline and the subsequent sedimentary sequence of ~9 Ma onlaps the anticline (Figure 13)[66]. This indicates activity along a N-S trending anticline in this stage. These observations on partial inversion in this stage have been attributed to an increase in horizontal compression stress [31]. In contrast to intense tectonic movements in both the fore-arc side and Backbone Range, the Akita coastal area and the Dewa Hills are considered to have remained bathyal environments (Figure 3). Siliceous shale of the Onnagawa Formation began to deposit in the Akita Basin at 12.3 Ma and yields high TOC-content and constitute major hydrocarbon source rocks (Figure 3)[22]. However, paleogeography of the Niigata Basin (Figure 2) reconstructed from wells [74] suggests that uplift of the eastern part of the Niigata Basin at the base of siliceous shale (~12 Ma) resulted in westward shift of the basin. The eastern margin of the Sea of Japan showed only minor deformation during this stage [49]. These observations suggest attenuation of compression stress and its related deformation toward the Sea of Japan [31]. This time interval (~10-8 Ma) was also characterized by the minimum of volcanic activity in northeast Japan (Figure 13)[16, 33, 57, 75-76]. The reduced volcanic activity during Stage III seems to have been attributed to an increase in horizontal compression stress [31, 33], because compressional stress may prevent the ascent of magma through the upper brittle crust to the surface [16].

Contemporaneous tectonic events at ~10 Ma have been reported not only from northeast Japan but also from Hokkaido (Figure 1) and from further south. Regional unconformity was formed in the western part of Hokkaido at about 12 Ma [77]. Duplex structures developed in the fore-arc basin off Hidaka, north of the Sanriku fore-arc basin (Figure 1) and have been attributed to westward shift of the outer Kurile arc and resultant uplift of the Hidaka Range (Figure 1) [78]. Concurrent deformation (uplift, faulting and folding) took place in the southern Japan Sea and the southern Korea at ~11–10 Ma [48, 79]. Synchronous tectonism with regional deformation and unconformity generation took place during the latest Middle Miocene to Late Miocene in the East China Sea and Ryukyu arc area (Figure 1)[80-82]. In the latter region the Lower-Middle Miocene Yaeyama Group dated to be as young as ~13–12 Ma [83] was folded and truncated by an erosional surface, and then covered by the Upper Miocene-Pliocene Shimajiri Group, which dates from N16 zone (11–8 Ma) [84]. Arc-continent collision had also started by 9 Ma around Taiwan [85]. The Middle Miocene unconformity at around 10 Ma was traced further south to the Pattani Trough in the Gulf of Thailand [86]. The simultaneous occurrence of tectonic events at around 10 Ma in the broad zone in the eastern margin of Asia suggests that compressional tectonics of this age may have had a more regional influence in eastern Asia than previously supposed [31].

The origin of regional compressional tectonics at around 10 Ma is still speculative at the moment. Compressional stress field with NE-SW maximum horizontal stress in northeast Japan in Stage III was attributed to westward shift of the outer Kurile arc and resultant collision of the western and eastern Hokkaido along the Hidaka Range [33]. However, the observations that deformation in northeast Japan tended to reduce towards the west while it was persistent towards the south to the northern Kanto Plain and the Joban fore-arc basin suggest that the Pacific Plate might be a key control and that the westward shift of the outer Kurile arc and resultant collision of the western and eastern Hokkaido was a consequence of regional compression rather than a cause. The late Neogene change in relative motion of the Pacific-Antarctic Plates started at about 12 Ma [87]. The increase in the spreading rate at the Pacific-Antarctic Ridge at 12 Ma might accelerate subduction of the Pacific Plate toward northeast Japan [87]. This might be a possible origin for a regional compression intensified at about 12 Ma. A change in Pacific-Antarctic Plates motion would also have affected the motion of the adjacent plates such as Philippine Sea and Indo-Australian Plates. For example, contemporary increase in spreading rates of Indo-Australian Plate at ~10 Ma at the eastern part of the Southeast Indian Ridge and a consequent peak in basin inversion in the West Indonesian Tertiary basins was suggested by [88].

5.5. Stage IV (Subsidence stage: 9 – 6.5 Ma)

At around 9 Ma, subsidence resumed on the fore-arc side, as well as in the axial Backbone Range (Figure 13). A rapid retrogradation occurred at around 9 Ma in the backbone range because of the resumption of subsidence (Figure 3). In the Joban fore-arc basin, a sedimentary sequence dated at ~9 Ma onlaps an eroded anticline structure with a notable hiatus between 11 and 9 Ma [66-67]. This suggests resumption of subsidence at ~9 Ma after the preceding partial inversion stage. Simultaneous resumption of subsidence and deposition occurred in the fore-arc lowland (Sendai Plain) and inland basins where an unconformity had been formed during the preceeding stage III (Figure. 13)[51, 53, 63]. This stage was also characterized by a period of intense felsic volcanism, with increased caldera formation at ~8 Ma on the Ou Backbone Range (Figure. 13)[33, 57-58]. Concurrent upward lithostratigraphic change from siliceous shale of the lower part to alternation of shale and tuffs of the upper part of the Onnagawa Formation (Figure 3) was suggested by analysis of wireline logs in oil-producing wells along the Akita coast [89].

Resumption of regional subsidence with increased volcanic activity in this stage may be attributed to weak extensional stress (Figure. 12). The occurrence of Northeast-Southwest trending minor normal faults in the Kurosawa Formation within the Yuda Basin [30] is consistent with this interpretation. Felsic magma trapped in the magma chamber or laccolith within the upper crust during the previous compressional stage may have been released to the surface under an extensional stress field. Caldera formation may have caused local domelike uplift in the Backbone Range during this stage [56, 72]. However, the origin of reduction in horizontal compression stress during Stage IV is unclear because no relative plate motion changes have been reported at around 9 Ma [87].

5.6. Stage V (basin inversion and compression stage: 6.5 – 3~2 Ma)

Stage V represents the differentiation of uplifted and subsided areas associated with crustal deformation in the back-arc region because of basin inversion and compressive stress field (Figure 13)[31]. The sedimentary successions in the Akita Basin (Figure 3) show basin-scale westward progradational stacking patterns including upward shallowing cycles, which consist of slope to basin-floor, trough-fill-turbidite, shelf, nearshore, delta, and fluvial systems, and likely reflect an increased accumulation rate caused by large amount of sediment supply from the uplifted backbone range and the Dewa Hills in response to the increase in compressional stress. The activity of the Kitayuri Thrust (Figures 3 & 4) extending north–south along the Akita coast started at around 5 Ma and resulted in the deposition of trough-fill turbidite system (Katsurane Facies) on the footwall trough of the thrust as a response to syn-depositional faulting and folding under compressional stress [25, 29, 90]. West of the Shinjo Basin (Figure 2), the Dewa Hills uplifted and emerged first in the southern part at about 5 Ma, followed by emergence of the northern part at about 4 Ma and by emergence of central part at around 3 Ma [32]. Four third-order depositional sequences consisting of shallow-marine to fluvial successions (i.e. Shinjo Group) developed in the Shinjo Basin, which represent gradual retreat of marine environments from the Shinjo Basin in response to the successive uplift of the Dewa Hills [32]. Four third-order depositional sequences accompanied by high-frequency depositional sequences consisting of shallow-marine, deltaic and fluvial successions developed in the Yuda Basin from 6.5 Ma to 3 Ma. The third and fourth 3rd-order depositional sequences in the Yuda Basin are correlated with the first and second 3rd-order depositional sequences in the Shinjo Basin, respectively (Figure 3). The correlation indicates marine incursion in the center of the Backbone Range until around 4 Ma, followed by separation from the Sea of Japan by emergence of the western sector of the Backbone Range [30](Figure 3). In other inland basins (e.g. Yonezawa Basin; Figure 2), the sedimentation of conglomerate increased after ~6 Ma

(Figure 13)[53], which suggests uplift of the surrounding mountains at that time. The origin of uplift of the Backbone Range and the Dewa Hills during this stage may be attributed to basin inversion due to increased compressional stress (Figure 13). A change in regional stress field from tension into E-W compression at 7–6 Ma was suggested by the earlier stress field studies in northeast Japan [91-93]. Similar basin inversion and change in depositional style at 6.5 Ma have been reported from the Neogene Niigata-Shin'etsu Basin in central Japan [94]. A notable angular unconformity was formed at the eastern margin of the Niigata Basin at around 7 ~ 6.5 Ma [95-96]. However, half grabens in the eastern margin of the Sea of Japan had not been inverted until early Pliocene [49]. In the fore-arc lowlands, major unconformities were formed at around 6.5 Ma, 5.5 Ma and 3.5 Ma in Stage V (Figure 13)[63, 65]. The unconformity at 6.5 Ma in the fore-arc lowlands can be correlated with that in the Backbone Range and in the Niigata Basin, which suggests a regional tectonic event.

This Late Miocene tectonic change associated with compressional deformation had a greater regional influence than seen the northeast Japan Arc alone. Ingle [48] pointed out that acceleration of uplift and deformation commenced at ~5 Ma in both northeast Japan and the Kurile Arcs (Sakhalin). Itoh et al. [97] demonstrated that Late Miocene uplift and deformation widely took place in the backarc side of the southwest Japan. The compressional deformation and uplift also occurred at 6.5 Ma in Taiwan [98]. The origin of these regional tectonic events has been attributed to resumption of subduction of the Philippine Sea Plate at ~7 Ma [97, 99-100]. However, contemporaneous motion change of the Pacific Plate commenced at 6 Ma [87, 101-102], suggesting more a regional tectonic event within circum Pacific region. For examples, transpressional tectonics along the San Andreas fanult, California, and the Alpine fault, New Zealand commenced at 6 Ma in relation to this change in the Pacific Plate motion [101]. This change in the Pacific Plate motion might also change the motion and subduction of the Philippine Sea Plate.

5.7. Stage VI (Intense compression stage; 3-2 Ma–Present)

Stage VI represents intense crustal deformation associated with the uplift and emergence of all present land areas because of the increased compressive stress [31, 56]. Major angular unconformities were formed at the base of Stage VI in the Yuda Basin and in the eastern margin of the Backbone Range [103], indicating intense uplift of the Backbone Range (Figure 12). The Akita coastal plain emerged at 1.7 Ma, resulting in westward shift of a sedimentary basin and submarine-fan deposition in Oga, followed by gradual fill of the basin with coarse sediments and by emergence of the basin-fill successions [29](Figure 3). However, the timing of basin inversion and of anticline growth varied from Earlly Pliocene to < 1 Ma according to structures both in the center of the Akita Basin [104] and in the eastern margin of the Sea of Japan [49]. Coeval deformation also occurred in the central and southwest Japan [94, 105]. The cause for the increased compressive stress during Stage VI has been attributed either to a change in the Pacific Plate motion [72] or to a change in the Philippine Sea Plate motion at 3 Ma [106].

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6. Conclusion

In this chapter, Late Cenozoic tectonic events in northeast Japan were reviewed. Both rifting process and post-rifting tectonics in northeast Japan were much more complex than those proposed in previous tectonic models [72]. Processes of the intra-arc rifting and opening of the Sea of Japan were interpreted as progression from core-complex mode (incipient rift system) to wide-rift mode (opening of the Sea of Japan and rapid intra-arc rifting) to narrow-rift mode (Late syn-rift system)[39]. A transition from extensional tectonics to compressional tectonics in fore-arc side of northeast Japan at the end of the wide rift mode may be related to the effect of lateral motions of the island arc; rotation of northeast Japan accelerated relative convergence rate of the Pacific Plate, thereby promoting compressional stress. A case study of intra-arc development from the Ou Backbone Range revealed three steps of uplift in 12 – 9 Ma, 6.5 – 3-2Ma, and 3-2 Ma - Present. These uplift events were correlated with regional tectonic movements not only in northeast Japan but also in other regions and were clarified as regional tectonic events. The origins of post-rift tectonic events in northeast Japan were inferred to have most likely attributed to changes in the Pacific Plate and Philippine Sea Plate motions.

The present review suggests that the tectonic mode in northeast Japan arc transformed from extension / crustal stretching to compression / crustal shortening much earlier (15 ~ 13.5 Ma) than previous models (3.5 Ma). Moreover, this change in tectonic mode was not straightforward but progressed forward and backward. Reactivation of normal faults bounding half grabens as reverse faults may have started earlier in Middle/Late Miocene. For this reason, the history of active faults may have been longer than previous estimates. Activities of active faults and uplift rates estimated assuming constant rate of crustal shortening after 3-2 Ma need to be reassessed. This also indicates that horizontal shortening rate estimated at around 3~5 mm/yr by assuming a constant rate after 2.4 Ma [4] might be overestimated. If so, only several % of plate convergence is accommodated within the northeast Japan arc as long-term deformation. This means that the 2011 great earthquake was inevitable consequence of accumulated elastic strain in northeast Japan arc. This review thus provides important implications for assessing activities of inland active faults, and for recurrence of great subduction zone earthquakes.

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