ЛЭПОМ

Данная монография основана на статье профессора Шигенори Маруяма (Япония) и его соавторов «Pacific-type Orogens: New concepts and Variations in Space and Time from Present to Past», опубликованной на японском языке в 2011 году в журнале Chigaku Zasshi (Journal of Geography. 2011. Vol. 120. P. 115-223), на материалах лекций, прочитанных профессором Маруяма на геолого-геофизическом факультете Новосибирского государственного университета в 2017-2019 годах, а также на научных результатах, полученных в Лаборатории эволюции палеоокеанов и мантийного магматизма в рамках реализации проекта № 14.Y26.31.0081 «Мультидисциплинарное изучение складчатых поясов тихоокеанского типа и создание согласованной модели эволюции океанов, их активных окраин и мантийного магматизма» по гранту Правительства Российской Федерации для государственной поддержки научных исследований, проводимых под руководством ведущих ученых в российских образовательных организациях высшего образования.

Shigenori MARUYAMA, Inna SAFONOVA OROGENY AND MANTLE DYNAMICS

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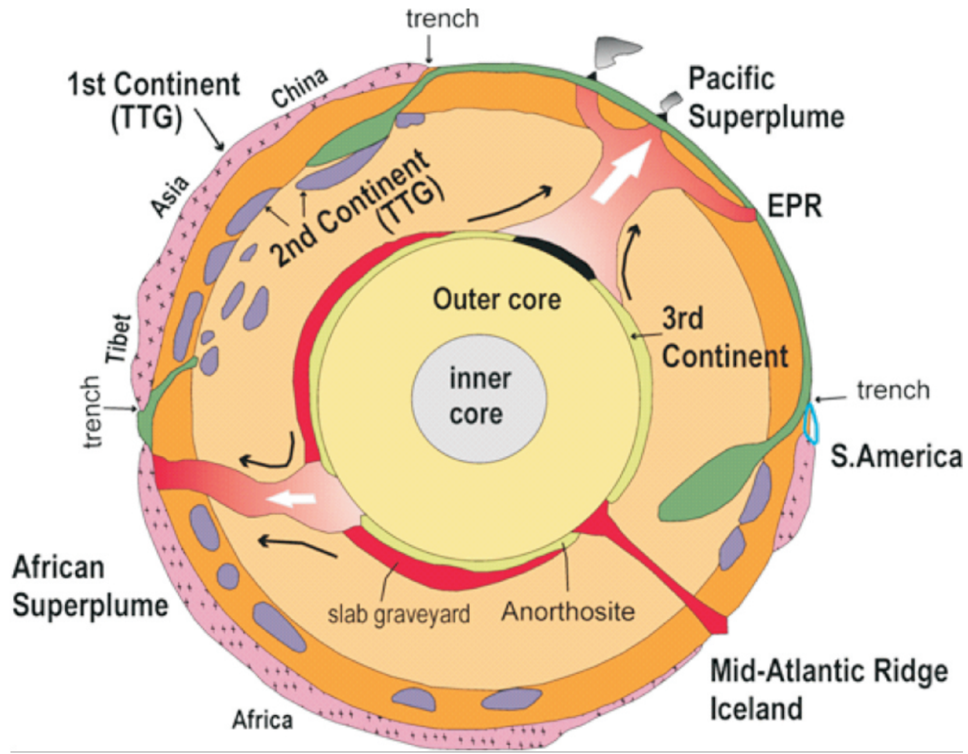
OROGENY AND MANTLE DYNAMICS: role of tectonic erosion and second continent in the mantle transition zone



Novosibirsk
2019

MINISTRY OF SCIENCE AND HIGHER EDUCATION
OF THE RUSSIAN FEDERATION
Novosibirsk State University

Shigenori MARUYAMA, Inna SAFONOVA

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УДК 551.14+551.2.03+551.21+551.24+552.11
ББК 26.324+26.325+26.21
М 296

Maruyama, S.

M296 Orogeny and mantle dynamics: role of tectonic erosion and second continent in the mantle transition zone / S. Maruyama, I. Safonova ; Novosibirsk State University. – Novosibirsk : IPC NSU, 2019. – 208 p.

ISBN 978-5-4437-0993-2

В книге рассматриваются ключевые вопросы орогении тихоокеанского и коллизионного типа, образования континентальной коры, тектонической эрозии, субдукции материалов континентальной коры в мантию, циклов суперконтинентов и мантийной геодинамики. Книга представляет интерес для широкого круга исследователей в области наук о Земле. Рекомендуется для студентов старших курсов и аспирантов всех геологических специальностей.

The book reviews and integrates key questions of orogeny, Pacific-type and continent collision type, continental growth, tectonic erosion, subduction of continental crust into mantle, supercontinental cycle and mantle dynamics. Could be interesting to a wide comunity of geoscientists. Also recommended to undergraduate and graduate students of faculties/departments specialized in Earth Sciences.

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Contents

Preface

This Book summarizes recent progress on the understanding of (1) Pacific-type orogeny in the context of global plate boundary process, (2) ubiquitous phenomenon of tectonic erosion at Pacific-type convergent or consuming plate boundaries, and (3) the fate of eroded and subduction material and its potential to grow 2nd Continents in the mantle transition zone (410- 660 km depth) toward the topmost lower mantle < 800km depth (MTZ). The volume of 2nd continents in the MTZ more than 10 times exceeds that of the surface continents (1st Continents). The formation of 2nd continents is controlled by the dynamics of the upper mantle and plate tectonics. The main goal of this Book is therefore to introduce the new concept of 2nd Continents and how this changes the understanding of Earth's dynamics. The history of the Earth from the Hadean, through Archean, Proterozoic, and Phanerozoic is briefly reviewed with a focus to Asian paleogeographic reconstructions including the Central Asian and Alpine-Himalayan orogenic belts in terms of the super-continent cycle and Pacific-type orogeny.

Pacific-type orogeny (PTO) has long been recognized as a contrasting accretionary alternative to continent-continent collisional orogeny. However, since the original concept was proposed, there have many new developments, which make it timely to produce a new reevaluated model, in which we emphasize the following new aspects. First, substantial growth of Tonarite—Trondhjemite—Granite (TTG) crust, and second the reductive effect of tectonic erosion. The modern analog of a Pacific-type orogen developed through six stages of growth exemplified by specific regions. Initial Stage 1 – accretionary growth in the southern end of the Andes. Stage 2 – exhumation of subducted and progressively metamorphosed rocks to midcrustal levels, the Indonesia outer arc. Stage 3 – Barrowian retrograde metamorphism or hydration stage, Kii Peninsula, SW Japan. Stage 4 – the initial surface exposure of the high-P/T regional metamorphic belt, Olympic Peninsula, south of Seattle, USA. Stage 5 – exposure of the orogenic core, Shimanto metamorphic belt, SW Japan. Stage 6 – post-orogenic tectonic erosion at the Mariana and Japan trenches and Nankai trough.

The fundamental framework of a Pacific-type orogen is an accretionary complex, which includes limited ocean floor material, much terrigenous trench sediment, plus island arc, oceanic plateau, and intra-oceanic basaltic material from the ocean. The classic concept of PTO stresses the importance of the addition of new subduction-generated arcs and TTGs to accreted rocks along the continental margins particularly during the Cretaceous. Besides the above additional or positive aspects of PTO, we emphasize the negative effects of previously little-considered tectonic erosion caused by subduction over time. The evaluation of extensive tectonic erosion leads to a prospect of the presence of huge quantities of TTG material in the MTZ, where many subducted slabs have ponded, as illustrated by mantle tomography. This is confirmed by mantle density profiles suggesting abundant TTGs only along the bottom of the upper mantle together with subducted slab peridotite and MORB. The major velocity anomaly in the MTZ is best explained by the presence of SiO_2 phases from TTG rocks, not possible from MORB or ultramafic rocks. Numerical calculations indicate that at a depth range of 520 to 660 km the amount of TTG material is 6-7 times higher than the total mass of the surface continental crust.

Various constraints on PTO and tectonic erosion come from the well-studied Japanese Islands. The traditional view is that the Japanese islands have evolved since 520 Ma through five Pacific-type orogenies, which grew oceanward, thus creating a continuous 400-500 km wide zone of accretionary complexes/orogens, with TTG growth at the continental side of each orogen. However, the subducting oceanic lithosphere has produced five times more TTG crust compared with the present TTG crust in the Japanese Islands. This can be explained by the fact that in certain periods tectonic erosion dominated compared to the growth of TTG arc crust. Accordingly, Japan has lost four TTG arcs because of tectonic erosion. TTG material, such as trench sediment, arc crust, and continental margin crust, was fragmented by tectonic erosion and transported into the bottom of the upper mantle to depths of 520-660 km. Worldwide data suggest that tectonic erosion destroyed and fragmented most of Pacific-type orogens.

The first supercontinent, Nuna/Columbia, formed at 1.9—1.8 Ga, before which there were just numerous island arcs and embryonic small continents scattered across the Earth's surface. Therefore, for the first half of its history the Earth was dominated by Pacific-type orogens, which predictably created big volumes of TTG crust. Collision-type orogens appeared after 1.8 Ga, following the breakup of Nuna, although the total volume of TTG crust on the Earth's surface was still less than 25%. Since the formation of the second supercontinent, Rodinia, at 1.0 Ga, continental collisions have become dominating to deform the shapes of continents, but they never increased the amount of TTG crust on the Earth. In the next 250 m.y., Eurasia and the western Pacific will evolve into a new supercontinent, Amasia, as a result of double-sided subduction from the east by the Pacific plate and from the south by the Indo-Australian plate.

Over geological time, the subducted TTG crusts amounted to more than 10 times the present total volume of those on the Earth, and this volume accumulated predominantly in the MTZ at depths of 520-660 km, as inferred by P-wave velocity anomalies compared to the Preliminary Reference Earth Model (PREM) of Dziewonski and Anderson (1981), which describes a 1-dimensional depth profile of average P-wave and S-wave velocities, density, and attenuation. The calculations suggest that the MTZ contains 6-7 times of the present TTG crust, and we infer that the remaining TTG material has not yet been scraped off the subducting slabs, and is somewhere in the uppermost lower mantle.

The density calculations by Kawai et al. (2013) of TTG material throughout the whole mantle suggest that the TTG crust has accumulated at the bottom of the upper mantle, and did not enter the lower mantle. The P-wave velocity whole mantle tomography by Zhao (2009) also indicates the presence of continental-type TTG material in the MTZ, which is concentrated under two regions - Asia and North to South America, but absent under the major Pacific, Indian, and Atlantic Oceans.

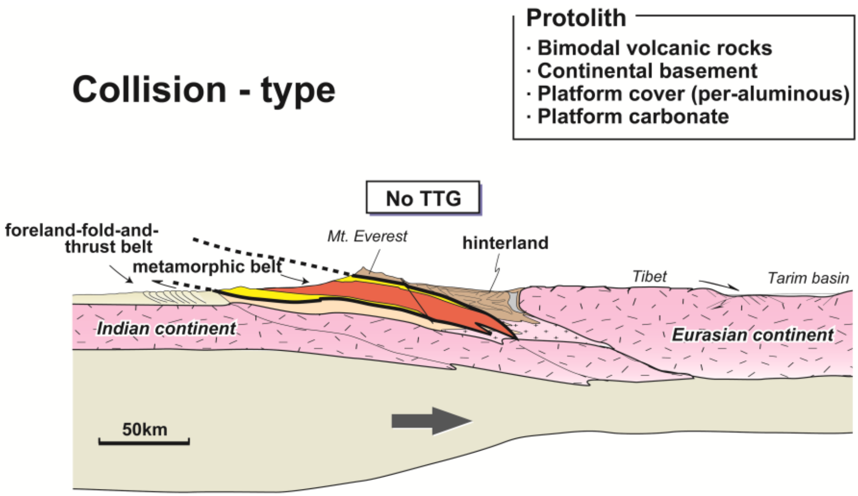
Stagnant slabs are well documented under East Asia by tomography, however, not in other areas. Interpretation of the composition of geophysical anomalies indicates that stagnant oceanic slabs must be accompanied by substantial volumes of TTG material. The triangular active subducting region in the western Pacific extending from Kamchatka and Japan, via New Zealand, to the western Himalayas could be due to the radioactive heating from that underlying substantial TTG layer, which generate hydrous plumes at a depth of 410 km, contributing to the formation of numerous intra-continental and marginal basins.

Mantle dynamics of the Earth over time must have been strongly affected or even controlled by the vast volume and distribution of TTG material in the MTZ, which in turn must have affected the development of superplumes, and the formation and disintegration of supercontinents.

Chapter 1. Introduction

What is orogeny? Concept of orogeny was established back to early 20 Century, through the research of geosyncline concept (no horizontal, usually vertical tectonics). It continued at early 1970. That concept of orogeny has been rather philosophical not based on a very detailed study of orogenic belts. Dewey and Bird (1970) were the first to speculate about orogenic belts based on the Theory of Plate Tectonics and to propose modern analogues of two types of orogeny or two end members. One is Cordilleran-type or Pacific-type or P-type and another is Continent-collision type or C-type (Fig. 1). The Andean mountain belts of South America were picked up as a world standard for Pacific type orogeny, whereas the Himalayan Mountains as Continent-collision type formed by the ongoing collision of India against Asia. The Japanese Islands were not considered as a Pacific type orogeny world standard because that time most geologists defined the importance of a geological phenomenon of orogeny as, first of all, mountain building. The Andes are over 4000 meters high mountains, but the Japanese Islands are typically less than 1 km high except supra-subduction zone volcanos, e.g. Mount Fuji (3776 m). In those days most geologists and geophysicists consider the definition of orogeny means mountain building as one of the most remarkable, impressive and important geological phenomenon.

Now term “orogeny” means not only mountain building, but also other complex geological processes, such as (i) the growth of continental crust by subduction-related magmatism; (ii) continental crust recrystallization and related formation of regional metamorphic belts; (iii) formation of a new geological structure through folding and mountain building; (iv) mountain building triggered erosion of continental mountain belts by rain and snow to delivery sediments into adjacent oceans. All these processes imply the phenomenon of orogeny as was later defined by Miyashiro and co-authors back in 1982 also based on the new paradigm of plate tectonics. In average, orogeny takes rather short periods of geological time, usually less than 100 Ma, and then finishes resulting in what we see in geological maps. As orogeny includes several complex geological phenomena, it represents an excellent and convenient term to describe the geological history of this or that particular continent and/or its composing orogenic belts. Nowadays we understand that orogeny is a kind of super complicated geological phenomenon which study requires extensive and intensive field studies including geological mapping and structural geology, laboratory studies of metamorphic rocks, geochronological investigations, igneous petrology and geochemistry combined with geophysical observations of the oceanic



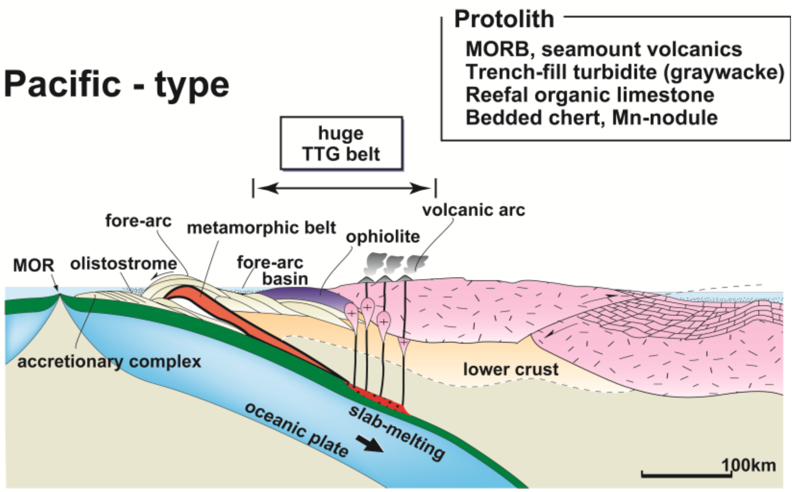


Fig. 1. Pacific-type (P-type) and collision-type (C-type) orogenic belts. The former create huge amounts of TTG crust, whereas the latter create no new crust, but modify strongly the pre-existing orogenic architecture. The two types of orogenic belts exhibit similar large-scale structure including orogenic core, which is regional high-grade metamorphic belt sandwiched between upper and lower unmetamorphosed units. Collisional belt is characterized by oceanward tectonic vergence and foreland fold-and-thrust belt, and Pacific-type belt by oceanward-thrust olistostrome (Maruyama, 1997). The protoliths of regional metamorphic rocks in are different in P-type and C-type belts (Maruyama et al., 1996).

floor and mountain belts and their underlying upper mantle. Those new data were combined together with the formerly proposed concept of orogeny including its two types: Pacific and Continent-collision.

These two types of orogeny must be more precisely defined. It took about twenty years of research to define more precisely what are P-type and C-type orogenies based on the results obtained from modern volcanoes and regional metamorphic belts (protoliths and age of metamorphic rocks and their PT-paths) and taking into consideration the exhumation tectonics.

All those data have drastically changed the understanding of the two types of orogeny. The focuses were made to (1) understanding petrogenesis of subduction zone magma, (2) origin and evolution of regional metamorphic belts with exhumation tectonics, and (3) formation of accretionary complex, (4) tectonic erosion. To support those understandings, geological mapping skill advanced, introduction of Ocean Plate Stratigraphy (OPS) concept with microfossil extraction skill and fossil species identification technology (SEM), age dating by LA-ICP MS for zircon, and P-T-time history of zoned garnet and other minerals combined with zoned zircons from surface to mantle depth and return back to the surface.

The simplified overall picture of these two orogenic belts is schematically shown in Fig. 1. A Pacific type orogen consists of (1) TTG-type batholiths dominated by felsic volcanic rocks, (2) accretionary complex, (3) high-P/T BS-EC regional metamorphic unit, and (4) fore-arc basin deposits. They form an orogen typically 400-500 km wide, 20-30 km thick, and 2000 km long. The core of the orogen is occupied by a regional metamorphic belt, which top and bottom borders are cut by the paired faults. A narrow fore-arc basin occurs between the volcano-plutonic sequences and the high-P/T regional metamorphic belt, which thickness is less than 2 km. The high-P/T metamorphic belt runs parallel to the batholith belt bounded by a low-P/T metamorphic unit at deeper levels. The general structure of a Continent-Collision type orogen is similar to that of Pacific-type orogens except the absence of both TTG belt and accretionary complex. So, it carries no TTG batholith belt but includes minor post-orogenic A-type granite plutons. The UHP-HP metamorphic belt with 4-5 km thick, twice larger compared to that of Pacific-type. The lateral continuation is variable, depending on the size of the colliding continents.

Orogeny takes geologically very short time, usually less than 100 Ma and then finishes. In other results a new orogen is typically formed under an older one and defined as a distinct unit in a geological map. If on orogenic unit is younger 100 Ma, i.e. formed at a modern consuming plate boundary, it is at least 1,000 km long, 30 km wide, and 3-10 km thick. If it is older in general > 100Ma, it tends to be fragmental in size and rock type, a few meters across, as commonly observed in the Central Asian Orogenic Belt. However, those fragmental assemblages in older orogens were originally as large as the modern examples (Fig. 2).

Pacific-type orogeny (PTO) has long been recognized as a contrasting accretionary alternative to continent-continent collisional orogeny. Since the original concept was proposed by Dewey and Bird (1970) and Miyashiro et al. (1982), there have appeared many new developments though, which make it timely to produce a new re-evaluated model based on new aspects. First, it is substantial growth of tonalite-trondhjemite-granodiorite crust (TTG-type crust). Second is the reductive effect of tectonic erosion. In more details, a modern Pacific-type orogen grows in six stages: 1) initial accretion and coeval supra-subduction TTG-type magmatism and HP metamorphism; 2) exhumation of metamorphic rocks to the mid-crustal level; 3) retrograde metamorphism; 4) initial exposure of a high-P/T regional metamorphic belt; 5) exposure of orogenic core; 6) post-orogenic processes including tectonic erosion. Accretionary complex is a fundamental constituent of a Pacific-type orogen. It includes limited ocean floor material, much terrigenous trench sediment, plus island arc, oceanic plateau, and intra-oceanic basaltic material from the ocean. The classic concept of a PTO stresses the importance of accreted oceanic rocks, new subduction-generated arcs and TTGs, which are added along continental margins.

Besides the above additional or positive aspects of a PTO, there are, as mentioned above, negative effects of tectonic erosion caused by subduction over time. The conventional geotectonic model for the evolving Pacific-type orogenic belts in Japan is as follows; after the breakup of the Proterozoic supercontinent Rodinia, Japan was born near the Yangtze (South China) craton. Proto-Japan changed from a passive margin to an active Pacific-type margin around 500 Ma, and has grown oceanward for ca. 500 km with the successive oceanic subduction from the Pacific side. Its oceanward growth was intermittent as punctuated by at least four orogenic peaks that produced four sets of regional high-P/T metamorphic belt and granitic batholith. Between these orogenic peaks, accumulations of large amounts of accretionary complex (AC) widened the arc-trench system. This scenario, however, is currently being challenged by new data and by new perspectives on the subduction-related process called tectonic erosion, which was originally proposed by marine geophysicists in the 1990s.

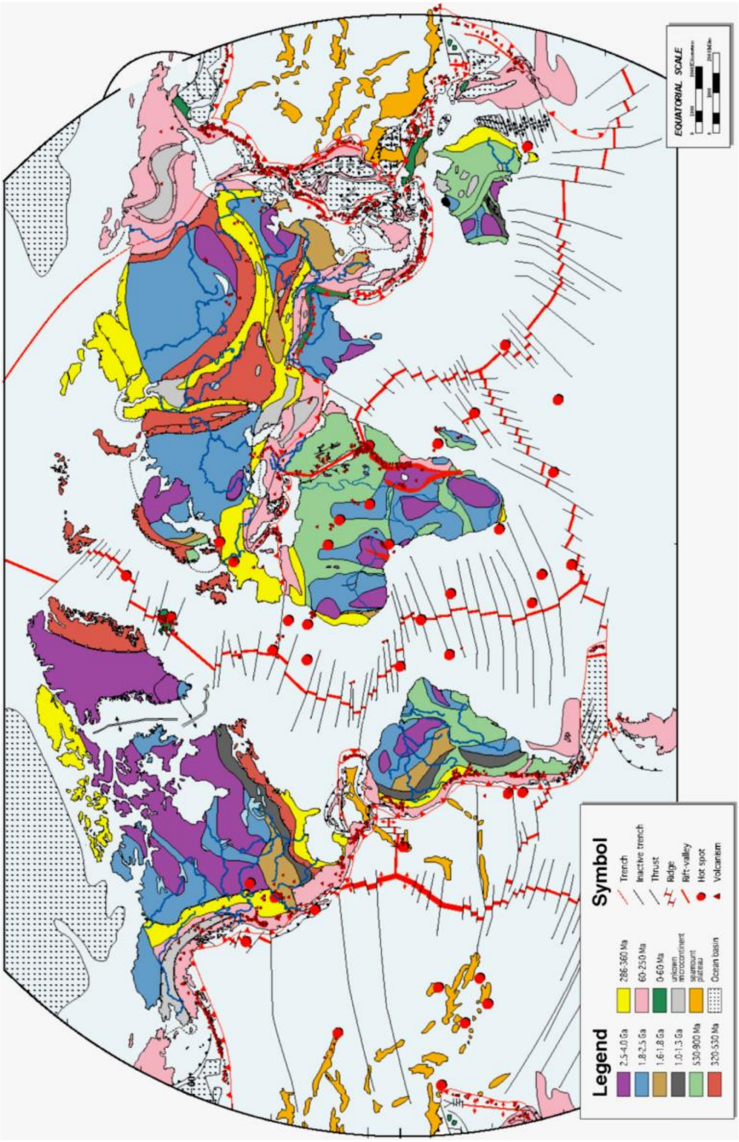


Fig. 2. Geotectonic map of the Earth. Continents are made of orogenic belts formed either by continent-collision or Pacific-type orogeny. To increase the TTG crust, Pacific-type orogeny has worked to cover 1/3 of the surface of our globe since 4.0 Ga until now. A geologic map of the Earth records the process of making continents and their dispersion over time.

Thus, the traditional view is that the Japanese islands evolved since 520 Ma through five Pacific-type orogenies, which grew oceanward, thus creating a continuous accretionary complex ca. 400-500 km wide, with TTG growth at the continental side of each orogeny (Fig. 3).However, the subducting oceanic lithosphere has produced five times more TTG crust compared with the present TTG crust in the Japan islands. This is explained by the fact that over time tectonic erosion has dominated the increasing arc-TTG crust. Accordingly, Japan has lost four arc-TTG crusts to tectonic erosion. TTG material, such as trench sediment, arc crust, and continental margin crust, was fragmented by tectonic erosion and transported into the bottom of the upper mantle at depths of 520-660 km. Worldwide data suggest that tectonic erosion destroyed and fragmented most of the Pacific-type orogens. The extensive tectonic erosion could be responsible for the presence of huge quantities of TTG material in the mantle transition zone (MTZ), where many subducted slabs are also ponded, as illustrated by mantle tomography. Evidence for the presence of TTG material deep in the mantle comes from the mantle density profiles of TTG, slab peridotite, lherzolite, and MORB, which show that TTGs are abundant only along the bottom of the upper mantle.

The C-type and P-type models of orogeny orogenic belts by Dewey and Bird (1970) reflect contrasting research histories. The first voluminous data sets came from the European Caledonian orogenic belts of early Paleozoic age, the late Paleozoic Variscan or Hercynian orogenic belts and the Cenozoic Alpine orogenic belt, all formed by continental-collision orogeny. Soon later extensive geological studies were performed in North America, particularly in its eastern part as the Appalachian foldbelt and other late Paleozoic Hercynian orogenic belts. As a result, the world geologists thought that continental collision a predominant orogeny and ignored Pacific type orogenic belts, because they did not discover Pacific type orogenic belts in East Europe and North America. Therefore, in spite of Dewey and Bird (1970) who introduced two types of orogeny (Cordilleran or Pacific type and Continent collision type), the Continent collision orogeny had been a prevailing process until the geologists started extensive research in the Japanese island and other foldbelts of the modern Circum-Pacific.

The definition of Pacific type orogeny has gained a strange direction. The eastern coast of the, North American continent includes the Appalachian orogenic belt formed by the collision of Baltica and West African cratons with North America, i.e. typical C-type. However the western coast of North America is quite different because no big continents are present in the Paleo-Pacific Ocean (in the Phanerozoic).

Therefore the tectonic development of the western coast of North

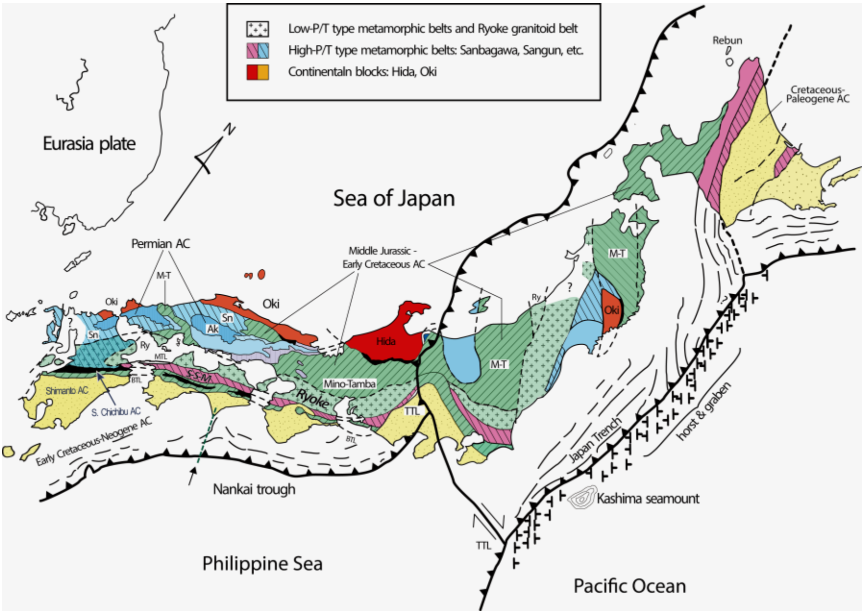


Fig. 3. Tectonic framework of central and southwestern Japan (modified from Isozaki and Maruyama, 1991; Isozaki et al., 2010). Ak – Akiyoshi AC; BTL – Bustuzo tectonic line; MT – Mino Tamba AC; MTL – Median tectonic line; S-S-M – Sambosan-Sambagawa-Mikabu belt, Sn – Sangun belt, TTL – Tanakura tectonic line.

America, which included the Cordilleran orogenic belt, faced collisional episodes. On the other hand, if we look carefully at the geology of the Pacific Ocean, we discover the Shatsky Rise and Ontong Java plateaus, which are recognized as aseismic ridges formed by rising mantle plumes. The Ontong Java plateau has about 50 km thick oceanic crust and is composed of basaltic materials, which are more buoyant than the underlying mantle. Therefore, if the Ontong Java plateau collides with a continental margin, it will induce similar to collision-type orogeny. In addition, although no thick continental crust is present in the Western Pacific, we see immature island arcs (Mariana, Izu-Bonin) and intra-plate volcanoes, such as the Emperor Hawaiian seamount chainPolynesia, etc. All of those topographic highs will collide in future and accrete along Pacific continental margins. Aseismic topographic rises, e.g., the Hawaiian volcanoes, will be first collide with island arcs or continental arcs to be entrapped into accretionary complex. Afterwards, the latter can be displaced along strike-slip faults, because subduction zones are typically not orthogonal but possess a strike-slip component. Therefore an accretionary complex can be displaced like the Franciscan complex in California (USE) migrating along the San-Andreas fault. In that case, fragmented accretionary complex can be removed and/or migrate, for example, from the southern tip of the Andes through Equator to the northern Alaska. The Alaska foldbelts include numbers of exotic terrains, which probably once formed much to the south and then navigated to the north to accumulate in Alaska. A typical example is the Wrangellia terrain, which is the largest oceanic terrain with continental crust underneath (Coney et al., 1980).

On the other hand, the history of research in Japan was completely different because of “a concept of accretionary complex”, which was introduced based on microfossils, conodonts and radiolarians. Japanese scientists made a revolutionary geological work in the Mino-Tamba Jurassic accretionary complex. Those days, the Jurassic accretionary complexes were believed to be late Paleozoic. T. Matsuda. Y. Isozaki, R. Hori, S. Kojima and other geologists separated microfossils and established a new concept of “Ocean Plate Stratigraphy” (OPS) (Fig. 4). They separates microfossils from each layer of deep-sea chert at less than 10 cm intervals following a detailed sketch map along the Inuyama river, Gifu prefecture (Fig. 5). They succeeded to demonstrate the presence of layer-parallel faults in a series of sediments from deep-sea bedded chert, through hemipelagic siliceous sediments up to trench turbidites, although the rate of sedimentation rate systematically changed depending on lithology. The sedimentation rate was 3-4 orders of magnitude slower in bedded chert compared to that of shallow marine hemipelagic sediments. They finally developed a breaking through model of “Ocean Plate Stratigraphy”, which can identify the time of sediment arrival and accretion from the ocean to the hanging wall of intra-oceanic arc or continental margin. The microfossil age from the lower turbidite tells the exact time of arrival to the trench. They started to apply the model of Ocean Plate Stratigraphy to other areas in Japan to understand the space-time distribution of accretionary complexes in Japan. Finally, they reconstructed the tectonic history of accretionary complexes in Japan through the anatomy of accretionary complex tracing the story from the youngest, ongoing subduction of the Pacific plate and Philippine plate to the structural top of a Cambrian accretionary complex.

However, their successful story disaccorded with the traditional geology not only in other countries, but even in Japan. That time, in the 1970-80-ties, most Japanese geologists still were using term “terrain” imported from the western coast of North America. In those days most important idea in Japan was the first to import the newly established concept from a foreign country. The older Japanese geologists, following the terrain concept, tried to rename the Japanese orogenic belts, such as Kurosegawa exotic terrain, instead of traditional Japanese geologic word, i.e., belt. One of the extreme interpretations was the meaning of the tectonic lines in Japan: before that the Median Tectonic Line defined in southwest Japan and Tanakura Tectonic Line defined in northeast Japan were considered identical because they were connected with each other before the opening of the Japan Sea. They considered those as strike-strip faults equivalent to the San-Andreas fault in the western coast of North America. The older geologists concluded that the Japanese belts all migrated from equatorial regions, like Indonesia, to Japan following the similar interpretation of the terrains emplaced along the western coast of North America.

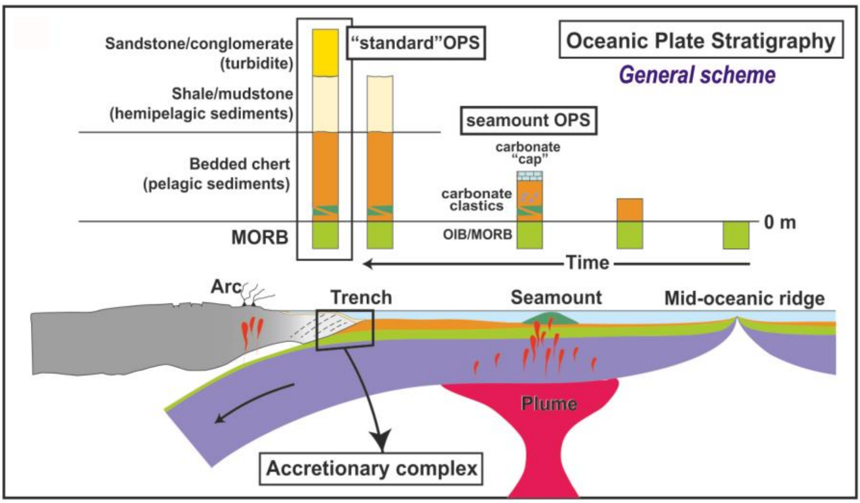


Fig. 4. A schematic illustration of the Ocean Plate Stratigraphy (OPS) model (modified from Maruyama et al., 2010).

But younger generations of Japanese geologists criticized such an approach. In the field of metamorphism a group led by Akiho Miyashiro performed a different unique tectonic synthesis as an extension of his concept of paired metamorphic belts (Miyashiro, 1960) based on the thermodynamics of metamorphism combined with plate tectonics.

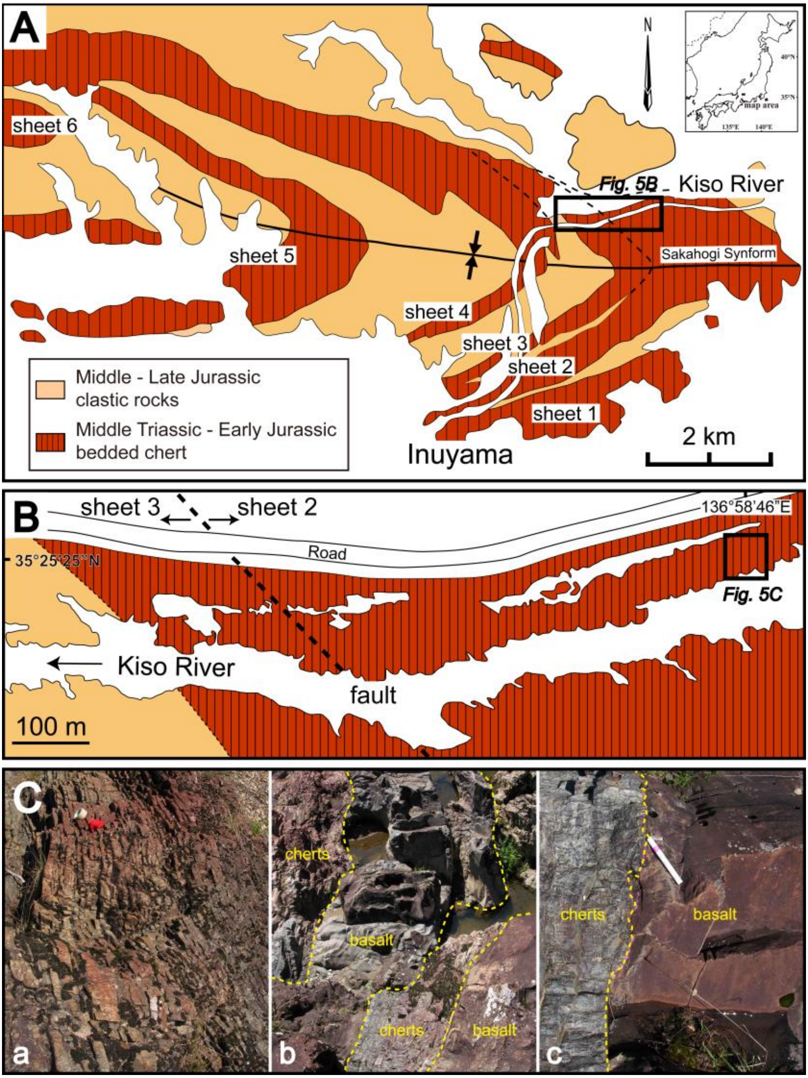


Fig. 5. The Inuyama area of the Mino-Tamba AC. A, Geological sketch map showing several bended sheets of Type 2 OPS (siliceous mudstone and turbiditic sandstone). B, Geological sketch map of the sampling area near Sakahogi station consisting of Jurassic chert hosting basaltic sills (modified from Matsuda and Isozaki, 1991; Hori, 1992). C, Chert-hosted basaltic sills (a, chert package; b, c, chert-basalt contacts).

A. Miyashiro focused on high-pressure, but low-temperature regional metamorphic belts (blueschist belts).Their geological, geochronological, petrological and geochemical studies and application of more advanced equipment produced abundant data sets to establish a new concept of orogeny. One of the major breaking through findings was the orogenic core occupied by highest grade metamorphic rocks with eclogites formed at 100-200 km depths and T = 600-1000℃. Before the discovery of UHP rocks, they thought it could form at ca. 10 kb, corresponding to a 30 km thick overlying rock unit. Therefore, three Himalayas must have been eroded to expose the orogenic core meaning cessation of both C-type and P-type orogenies. But in case of Japan a critical point is (1) the occurrence of UHP-HP metamorphic units sandwiched between the low-pressure units, and (2) absence or extremely small amounts of detrital sediments which are expected to form during and after mountain growth. The answer is simple: UHP-HP units were extruded as a thin slab into virtually non-metamorphosed units above and below, as a solid “sill”. Many remarkable examples of exposed orogenic cores can be found in the Himalayas, European Alps (C-type belts) and Indonesia (Timor non-volcanic arc).

A key point of P-type orogeny is the formation of TTG over subduction zone, i.e. production of new continental crust, which, however, can be destroyed and transported from the surface to the bottom of the MTZ, to create a new engine of mantle convection driving tectonic plates. Mechanisms of crust destruction and downward transportation, and the role of self-heating will be also discussed in this paper as these processes play a key role in the supercontinent cycle.

The hanging wall of P-type convergent margin can be destroyed by tectonic erosion or subduction erosion. Moreover, direct subduction of intra-oceanic arcs into the mantle is also common on the modern Earth, and was a dominating process in Archean time. Evidence for this comes from the OPS units discovered and investigated worldwide. The subducted TTG form a kind of “lost” or “second” continents growing through Earth’s history at the bottom of the upper mantle and uppermost lower mantle. They are gravitationally stable there, hence never return back to the surface. The process of transportation of continental crust material from the surface to the MTZ is unidirectional, therefore, the size of 2nd continents must have increased through Earth history, at least bigger than the total volume of 1st continents on the Earth (Kawai et al., 2013). Those TTG rocks are enriched in radiogenic elements such as U, Th and K, they would perform as a self-heater, generating heat by radiogenic decay providing temperature increase to 200 K per 100 Ma during the Archean and to 100 K during the Phanerozoic (Senshu et al., 2009). Thus, the 2nd Continents may work as an engine driving plate tectonics. These new facts and observations will open a new view of Earth through creating several new paradigms (Fig. 6).

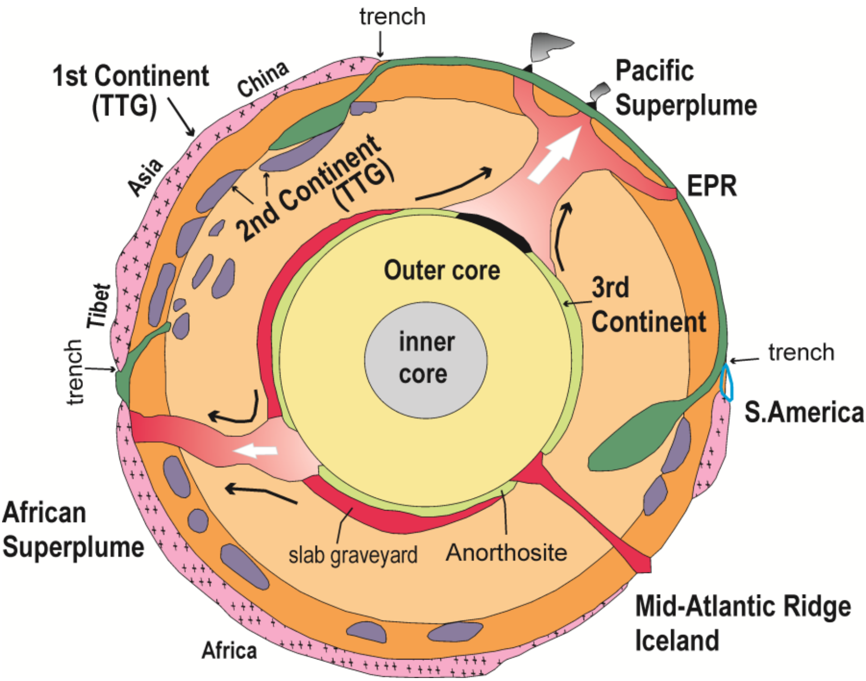


Fig. 6. Distribution of 1st, 2nd and 3rd continents of the Earth shown schematically along a cross-section of the Earth (after Maruyama et al., 2007a, b and Huang and Zhao, 2006) – to be discussed later. Both the 1st and 2nd continents are composed of TTG, but the 3rd is anorthosite in composition (Kawai et al., 2009). The second continents are largest under Asia, and not only in the mantle transition zone but also in topmost lower mantle. A cross-section of the Earth as a great circle passing Japan, Asia, Africa, Iceland and South America is shown. Three major mantle convections of Pacific superplume, African superplume and one super-downwelling under Asia are reproduced after Maruyama (1994) and Maruyama et al. (2007b).

Chapter 2. Two types of orogeny: Pacific and continent collision

2.1 Brief history of research

In the 1960-ties and 1970-ties the concept of orogeny was originally based on the geosyncline concept implying no horizontal, mostly vertical tectonic movements. That time the concept of orogeny was rather philosophical, not based on very detailed studies of orogenic belts. That approach was different from the modern concept of orogeny defining two major end-member types: Pacific (P-type) and Continent collision (C-type). Dewey and Bird (1970) were the first to speculate about orogenic belts in terms of plate tectonics and their first concept was very similar to modern analogues in case of C-type. The Andean mountain belts of South America were picked up as a world standard for Pacific type orogeny, whereas the Himalayan Mountains as Continent-collision type formed by the ongoing collision of India against Asia. The Japanese Islands were not considered as a Pacific type orogeny world standard because that time most geologists defined the importance of a geological phenomenon of orogeny as, first of all, mountain building. The Andes are over 4000 meters high mountains, but the Japanese Islands are typically less than 1 km high except supra-subduction zone volcanos, e.g. Mount Fuji. It took about twenty years to change the idea drastically due to new advances in studying modern volcanoes, regional metamorphic belts and exhumation tectonics. It was very important to understand geological, physical and chemical details by modern techniques, in particular, a tandem between dynamics of tectonic exhumation and isotope geochronology, e.g., the age of protoliths and the age of regional metamorphism correlated with changing pressure and temperature. The orogeny became a super multi-disciplinary subject, spanning field geology, structural geology, geochronology and metamorphic and igneous petrology combined with geophysical observations on the ocean floor and tracing the structure of mountain belts from the surface to the upper mantle. Finally, those advanced techniques, new approaches and new results were combined together with the new proposal of a concept of orogeny: Pacific-type and collision-type (Fig. 1).

2.2 Structure of Pacific-type orogen

The overall structures of both P-type and C-type orogens are relatively similar. P-type belt is typically 400-500 km wide, more than 2000 km long, and consists of several subhorizontal units. The major constituents of P-type belt are batholith belt, fore-arc basin, regional metamorphic belt and accretionary complex. The batholith belt (hereinafter TTG type or orogenic I-type) with the upper edifice of felsic volcanic rocks can be up to 15 km thick; it is emplaced on the continental side and extends over a distance of 2000 km parallel to the trench. The core of orogen is occupied by 2 km thick regional metamorphic belt (Fig. 1). The regional metamorphic belt and the batholith belt are separated by fore-arc deposits. The fore-arc deposits are those typically derived from adjacent magmatic arc. The accretionary complex consists of OPS units (the rocks scratched off the subducting oceanic plate) and terrigenous deposits derived from adjacent metamorphic and igneous belts (for details seethe relevant sections below).

In P-type belts the regional metamorphic belt occupies the central part of the orogeny and possesses a sub-horizontal structure (Fig. 1). In first sight it looks like intrusion, but it is not. The central part is a HP-LT metamorphic belt (blueschist belt) characterized by anticlockwise P-T (pressure-temperature) path through time. This HP belt is sandwiched between weekly metamorphosed HT-LP metamorphic sequences related to the batholith belt. The top and bottom surfaces of the regional metamorphic belt are cut by paired faults. The internal structure of the metamorphic belt, in terms of lithology, metamorphic fabric and P-T parameters, is symmetric. The exhumation of the metamorphic belt usually proceeds along the continental side of P-type belt, which hosts the highest pressure and temperature metamorphic rocks. The vergence angle declines oceanward (Fig. 7). The metamorphic rocks display crystalline fabric, deformation-related folding and penetrative fabrics. The source rocks (protolith) are those of the accretionary complex. Instead of accretionary complex, the upper unit of the metamorphic belt may include LP metamorphosed ophiolitic rocks. The P-T parameters of P-type belts are typically much lower (twice) than those of C-type belts (Maruyama et al., 1996).

Huge batholith belt is emplaced on the continental side of P-type belt lying parallel to the regional metamorphic belt. TTG intrusions may be exposed in its lower part. The igneous rocks may suffer LP-HT metamorphism and, coupled with the HP-LT belt, form so called paired metamorphic belts by Miyashiro (1960). The concept of paired metamorphism came from the geology of Japan, but now it seems to have an exceptional character. Now we know that the Ryoke granitoid belt was juxtaposed over the Sanbagawa HP-belt as result of the opening of the Sea of Japan Sea in Miocene time, at ca. 20 Ma. That juxtaposition exposed a part of the Ryoke belt north of the medium tectonic line (Fig. 3). The Sanbagawa belt is best exposed in southwestern Japan. In western Shikoku, it is clearly seen that the Ryoke belt rests on the top of the Sanbagawa belt (Aoki et al., 2010).

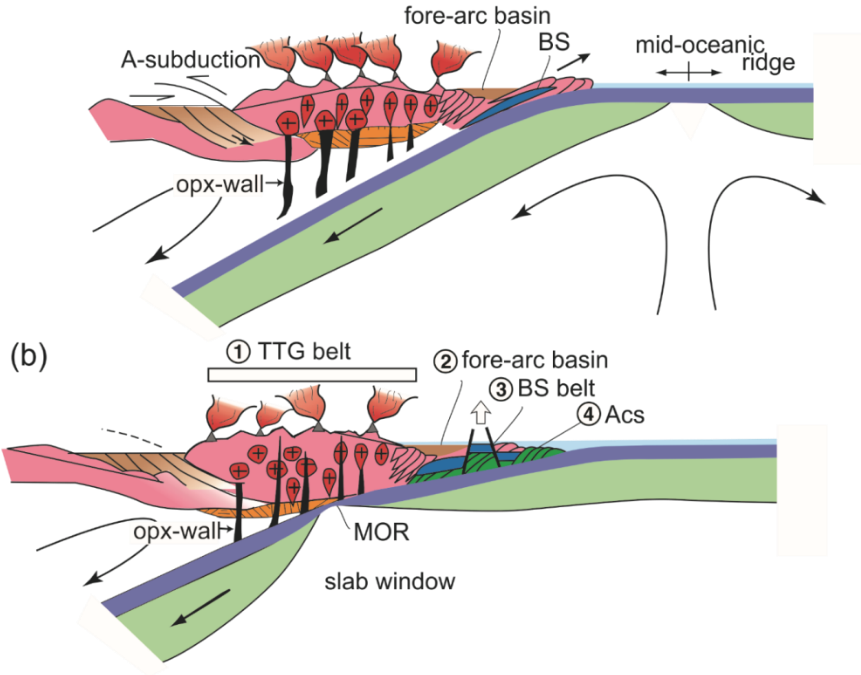


Fig. 7. Diagrams illustrating the formation of a Pacific-type orogenic belt (Maruyama, 1997); (a) formation of accretionary complex, its transportation and underplating at depths of 50—60 km and simultaneous regional HP metamorphism, followed by tectonic exhumation along the Benioff plane due to the approaching mid-oceanic ridge. This process is accompanied by TTG formation due by slab-melting. (b) tectonic juxtaposition of a high-P/T belt under a fore-arc followed by the doming caused by the underplating of accretionary complex and thick trench turbidites that accumulated after ridge subduction. This model does not account the effects of tectonic erosion, i.e. assuming continuous growth of accretionary complex oceanward (Maruyama, 1997).

Accretionary complexes over and below the regional metamorphic belts have different ages. In the case of the Sanbagawa belt of Japan, the overlying Jurrasic accretionary complex formed at 180 Ma ago (Isozaki et al., 2010). The Shimanto accretionary complex below the Sanbagawa belt formed at about 80 Ma. Hence the difference between the ages of the accretionary complexes is about 100 Ma. The Sanbagawa metamorphism picked at about 130-120 Ma (early Cretaceous), however the ascent of the metamorphic rocks to mid-crustal levels was only at 80 Ma ago suggesting a very slow rate of exhumation (Okamoto et al., 2004). To be discussed in details later the time span about 50 Ma. The protoliths of Sanbagawa metamorphics were the rocks of the Shimanto accretionary complex, which were tectonically eroded and transported by the subducting Pacific slab into the deep mantle, to the depths of 50-60 km, and then returned back to the surface. Regional metamorphic belt occupying a structural middle of P-type orogeny has a dome-like structure formed by its exhumation. The gently folded antiform is surrounded by synforms, both cut by normal faults. The structural bottom underneath the antiform is non-metamorphosed or weekly metamorphosed accretionary complex.

2.3 Structure of continent collision-type orogen

The structure of continent collision type (C-type) orogens is different from that of P-type orogeny mostly in terms of lithology it its constituents. Unlike P-type orogens which are characterized by huge batholith belts, granitic plutons are much less abundant in C-type orogens (Fig. 1). First, granitic plutons are smaller in size, typically several kilometers in diameter. Second, C-type belts include mainly A-type granitic plutons or leucogranites derived by the partial melting of HT metamorphic rocks, which protoliths were water-saturated siliceous mudstone or sandstone. Those intrusions crosscut the boundary faults separating the initial metamorphic units from top and bottom as has been shown for the Himalayas (Sakai, 2002).

Similar to P-type orogeny, the regional metamorphic belts of C-type orogens are also tectonically emplaced into nearly non-metamorphosed or weekly metamorphosed rock units bounded by horizontal paired faults from top and bottom. Those paired faults and the modern occurrence of regional metamorphic belts in C-type orogens clearly indicate the exhumation of regional metamorphic belts in solid state. The thermobaric structure of a C-type regional metamorphic belt shows high PT-values in its central part and near its continental side. The metamorphic fabrics are also characterized by symmetrical patterns. Those thermobaric and rock structural characteristics suggest emplacement of regional metamorphic belts be a process called “wedge extrusion” (Maruyama et al., 1996). The protoliths of C-type regional metamorphic rocks are passive continental margin deposits, i.e. remarkably different from the protoliths of P-type metamorphic belts, which are oceanic and terrigenous rocks of accretionary complexes (Kaneko, 1997) (Table 1). The pressure estimates for metamorphism of C-type orogeny are twice higher than those of P-type orogeny: the pressure can reach 70 kbar, e.g., in the

Table 1

Two major types of Orogenic Belts

Rock types/other aspects	Pacific-type (P-type):subduction-related orogens with accretionary complexes	Himalaya-type (C-type): collision-related orogens built over continental basement
	Key lithologies	
Shallow-marine sediments	Reefal limestone, siliceous shale and mudstone (OPS)	Thick platform carbonate and clastics
Deep-sea sediments	Bedded chert (OPS)	Absent
Turbidite	Trench-fill graywackes (flux of andesitic materials)	Platform cover (peraluminous with chloritoid, carpholite, etc.
Terrigenous fan	deep-marine flysch/forearc	continental molasse and/or foreland basin
Volcanic units	Mafic to andesitic and felsic lavas	Bimodal (basalt and dacite) lavas
Intrusive rocks	Huge subduction-related granitoid batoliths	Syn-collisional (I and S-type) and post-collisional (A-type) granitoids
	Related metamorphic units	
Key metamorphic assemblages	Blueschist, greenschist, HP-UHP	Blueschist, MP (kyanite-sillimanite), UHP
Blueschist protoliths	MORB, OIB	Peraluminous, bi-modal
Retrograde assemblages	Minor	Abundant
Regionally metamorphosed peridotite	Spinel-, plagioclase-peridotite Strongly serpentinized	HP and HP units after garnet- and spinel-peridotite
Paired metamorphic belts	May be present	Absent
	Other aspects	
Continental basement	Absent	Abundant granite-gneiss complexes
Ore	VMSD, black smokers, Fe-Mn and volcanogenic-sedimentary deposits	Petroleum (foredeep troughs), gold quartz veins; gems
Intraoceanic arcs with boninites	May present	Always absent
Structural features	Duplex OPS structures (older rocks may be above younger); dominantly asymmetric folds and small thrusts	Variable structures and deformation styles providing buckling, thrusting and contraction

Abbreviations: HP-UHP – high pressure-ultra high pressure; MORB – mid-oceanic ridge basalt; MP – medium pressure; OIB – oceanic island basalt; OPS - Ocean Plate Stratigraphy; VMSD – volcanic massive sulfide deposits.

Kokchetav Massif in Kazakhstan (Kaneko et al., 2000); the temperatures can be 900 or even 1000^{o}C, i.e. the Kokchetav massif hosts the world highest pressure and temperature metamorphic rocks (Maruyama et al., 2002).

C-type metamorphic belts exhibit a later episode of hydration, which may look similar to P-type subduction-related progressive regional metamorphism suggesting progressive high temperature dehydration and then exhumation. However, those progressively zoned metamorphic structures/minerals were extensively hydrated and recrystallized during tectonic juxtaposition at pressures of about 3-4 kbar. At that time the still ongoing subduction promoted extensive hydration and recrystallization, which obliterated progressively zoned isogrades and minerals. C-type belts are characterized by nearly 100% recrystallization. For example, eclogites – HP metamorphic rocks - may contain coesite pseudomorphs or coesite itself within small pond-like concentric metamorphic cores of amphibolite surrounding eclogites with chromite. This indicates high-pressure metamorphism accompanied by hydration, but at intermediate pressure kyanite-sillimanite facies of regional metamorphism. As a result, the metamorphic rocks formed by progressive UHTP metamorphism obviously recrystallized 20-30 Ma later, at intermediate or even lower pressures and temperatures. Evidence for this comes from the eclogites of the Alps and Himalaya and from the Kokchetav Massif of Kazakhstan (Katayama et al., 2000, 2003; Lui et al., 2002; Kaneko, 1997).

The structural bottom in C-type belts is present in the large fore-land fold-and-thrust belt. That belt growth in direction opposite to the direction of continent subduction. Olistostrome in the front of one thrust nappe can result from gravitational collapse of another nappe. In general, olistostrome of C-type orogen is rather similar to accretionary complex P-type orogen.

The orogenic core of C-type belt is also nearly horizontal to subhorizontal HP-MP metamorphic belt sandwiched between other geologic units. The sandwich structure is cut by normal faults to form several rectangular blocks, each typically 100 km across. The central part of the orogenic core is domed up as in the Dabieshan UHP belt of central China formed by the collision of North China and South China. The Cenozoic Himalayan collisional belt possesses similar structure.

2.4 Pacific-type orogeny as a process

Pacific-type orogenic belts have been extensively studied since the 1990-ties to produce huge amounts of geochronological, petrological geochemical and isotope data enabling us to define a new concept of Pacific-type orogeny. Key constituents are (from ocean to continent): accretionary complex, metamorphic belt, fore-arc basin and granitoid belt (or batholith belt) (Figs. 1, 7) The less than 2 km thick metamorphic belt makes a core of the orogenic belt and consist of HP-LT metamorphic rocks formed after accretionary complex, which can be subducted and metamorphosed at low temperatures and high pressures following the subduction geotherm. The structural bottom is non-metamorphosed accretionary complex. The geological structure in general gently dips toward the continental side, whereas the axes of folding axis show oceanward vergence. The world standards are Japanese Islands, California of North America, and Cretaceous orogenic belts of western Pacific. Figure 2 shows a process how to make a Pacific-type orogen (Maruyama, 1997). The material derived from accretionary complex decent into the mantle depth together with oceanic slab to a depth of 20 km. With increasing pressure and temperature, dehydration starts following the subduction geotherm, which is a clockwise PT-path. Assuming an average subduction speed as 10 cm/year and an angle of subduction as about 45 degrees, it may take about 0.4 million years material to travel from the trench to a depth of 60 km, which is surprisingly short in time. In general, trench sediments are removed coevally with the subduction of oceanic slab and can be transported probably further down, to the bottom of the upper mantle. However ay certain conditions, e.g., at ridge subduction, the subducting sedimentary package must stop at the 60 km depth, and then get underplated against the hanging wall of continental margin. The package/tectonic sheet detached from the descending oceanic plate start to move back to surface along the Benioff plane, in a direction opposite to that of the descending oceanic slab. The ascending flat-shaped geologic units experience regional metamorphism at temperatures of 700-800°C. The rocks in the center of such a platy body, which experience higher PT conditions, must be of lower viscosity, more ductile and soft compared to the colder top and bottom parts. Therefore, those soft parts can be easier removed to start moving upward following the PT-path of exhumation. As a result, a very thin platy unit of exhumed regional metamorphic rocks finally intrude early non-metamorphosed or weekly metamorphosed accretionary complex units at depths of 15-10 km, i.e. the brittle-ductile transition. The temperature at the transition is about 350 degrees. At that depth, the exhumation stops to form the mentioned above sandwich-like structure. At that stage, the continuously subducting oceanic slab carries abundant water from dehydrated rocks underneath the emplaced HT regional metamorphic belts. Those fluids react with the still high-temperature regional metamorphic belts to form secondary metamorphic minerals under the conditions corresponding to intermediate pressure facies (Maruyama et al., 2010) (Fig. 7).

The tectonic exhumation proceeds towards the highest point of the basement or trench slope break (TSB), which is topographic turning point from the bottom of the trench to the continent ward. Away from the trench slope break and closer to the volcanic front, the underplated segment gets more and more flat position. The TSB as a topographic turning point forms by a fault tracing the surface of deeply subducted slab and separating accretionary complex and the hanging wall on the surface. Along that fault, regionally metamorphous units follow the exhumation path giving a birth of a non-volcanic second arc. A series of sedimentary basins is present between the volcanic front and the non-volcanic arc. Such a double-arc structure is formed over a mature subduction zone, like in the Indonesian region (Banda volcanic arc and Timor non-volcanic arc) and similar in the Japanese island (Ryoke batholith and Sanbagawa metamorphic belt). The history of non-volcanic arc and the growth of TTG belt are a key to explain large amounts of materials derived from accretionary complex to the trench. If younger accretionary units are underplated beneath the already exhumed regional metamorphic belt at mid-crustal depths, they will push out the regional metamorphic belt at a structural region between the arc and the trench. Such a jack-up causes sandwich type geologic units to induce doming and related high-angle normal faulting.

The exhumation of a regional metamorphic belt is accompanied by the formation of granitic crust beneath the volcanic front. A 200-300 km wide batholith belt can be formed during 40 Ma. Such an anomalously high rate of the generation of felsic melts is possible only by the partial melting of subducted oceanic crust. Most petrologists think that the dehydration of subducting slab can induce partial melting of mantle wedge to give basaltic melts, which ascend to the Moho depth. But to make felsic magma by this process we must reach 20% partial melting of the lower mafic crust, which is hardly possible. Therefore, by this model we cannot explain huge amounts of TTG-type igneous rocks formed under Pacific-type orogenic belts during a relatively short period of time. Thus, Pacific-type orogeny includes the formation of accretionary complex, regional metamorphism and exhumation of metamorphic rocks to upper levels and related doming, and generation of TTG-type magmas. However, the two main unsolved questions remain: why P-type orogeny takes a geologically short time period and the regional metamorphic belt moving from the surface to the mantle and then back to the surface never reaching it. To discuss these questions we need one more constrains from the modern ongoing Pacific-type or Collision-type orogenies.

2.5 Modern analogs of Pacific-type orogeny

One cycle of orogeny takes about 100 Ma years including diverse geologic events and processes. Therefore, if we consider this or that modern analogue we must first identify which point of the 100 Ma interval a modern orogenic belt is now at. Based on the observed geologic phenomena, we divide orogeny into following six steps and show a modern analogue for each (Fig. 8).

<u>Stage 1. Initial stage of Pacific-type orogeny. Progressive metamorphism.</u> At Stage 1 the degree of metamorphism increases progressively to a depth of about 60 km in the mantle. Trench sediments or tectonically eroded crustal materials descend along the Benioff plane (Fig. 9a). Modern analogues are the South American Chile trench, from the middle to the south, northeastern Japan and Izu-Mariana arc (Fig. 8). These are Pacific-type subduction zones with volcanic fronts in the backward, e.g., there is an active volcano on the backside of the Chile trench. The Chile trench is characterized by ongoing extensive tectonic erosion supplying sediments to the trench, while the subducting slab is carrying those to the mantle (Vonheun and Yamamoto, 2010). The material transported down experiences progressive metamorphism. Evidence for this comes from the link between the distribution of the seismicity and the PT phase diagram down to 100 km (Omori et al., 2002, 2004, 2009).

<u>Stage 2. Middle stage of Pacific-type orogeny. Fore-arc basin. Non-volcanic arc.</u> The rocks transported and metamorphosed at the 60 km depth, stop there and start returning back to the middle crustal level (Fig. 9b). Modern analogue is the non-volcanic outer arc of Indonesia, in particular, Java Island. At that stage a non-volcanic outer arc becomes topographically above the sea level just off the Pacific type subduction zones. The Indonesian outer arc is characterized by the formation of EW-trending orogenic or Alpine-type serpentinite belt and amphibolite belt. This is a typical example of a non-volcanic arc where a mid-oceanic ridge separating much older plates subducted during the Eocene and later (Hamilton, 1979).

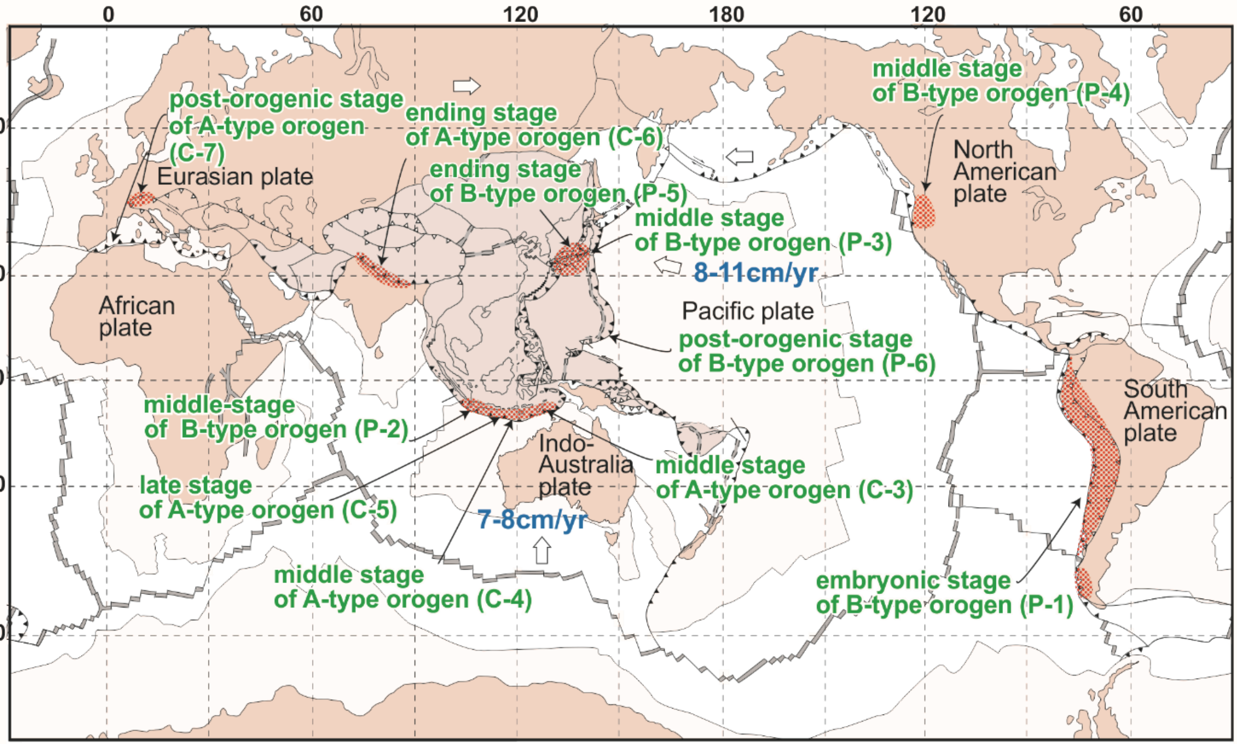


Fig. 8. World map showing sites of on-going orogeny at different stages along the major consuming plate boundaries (partly modified after Maruyama et al., 2010). Six stages of Pacific-type and collision-type orogenies are defined.

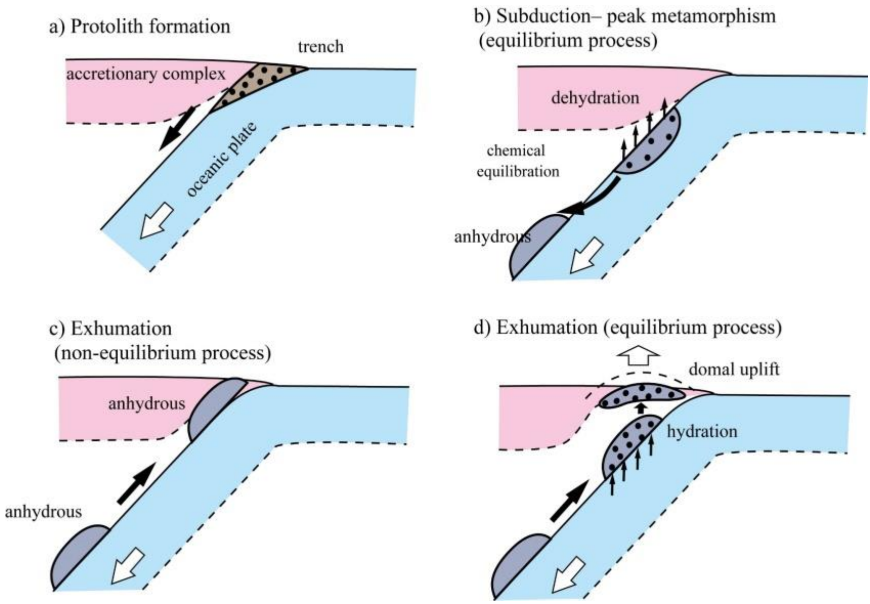


Fig. 9. A simple scheme of the first four stages of Pacific-type orogeny.

Stage 3. Late stage of Pacific-type orogeny. Formation of dome structures. Fore-arc basin sedimentary facies change from flysch to molasses-type. Borrowian-type metamorphism hydration and recrystallization (Fig. 9c). A modern analogue is the Kii peninsular and the southern Yamato basin in SW Japan. These geological structures all are characterized by clear dome-shaped topographic features. Recent exhumation and mountain building have been recorded in the Kii Peninsular by the sand waves described by a Japanese ancestor at 6000 years ago present along the beach, which are now elevated to 20 meters as can be traced in places. We know that six thousand years ago, when the global warming started/occurred, the sea level was 4 to 5 meters higher than the present-day sea level, as recorded in historical documents. Then we can calculate the meterage of doming by adding 4 to 5 meters that past sea-level point. In total, we estimate an exhumation of 15 to 16 meters during the last 6 thousand years. This corresponds to the 2-3 mm per year rate of exhumation or elevation. This value matches that of the Himalaya range, Taiwan and Indonesia non-volcanic fore-arc region (Timor Island). The Yamato area between the Kii Peninsular and the western Shikoku Island is a central portion of that region corresponding to the peak of uplifting. In between those and to the east of Kii Peninsular, there is a bay, named Kii straight. Also there is an Inland Sea between Kyushyu and Shikoku islands. All these 3 topographic features are formed by a NS-trending normal fault. Therefore, the outer region of SW Japan is a region of repeated subsidence and also uplifting, i.e. it has a kind of left-angular horst-and-graben structure, which is about 100 km across. To the south of the Kii Peninsular, i.e. between it and Shikoku, there is an outer fore-arc Shikoku basin, under which the youngest Philippine Sea plate has been subducting under Asia since the Miocene. The basin is as young as 8 Ma. Underneath the Kii Peninsular, the subducting Philippine Sea Plate shows that the subduction angle suddenly becomes deeper. This is contrasting to the east and to the west, where the subduction angle is rather gentle. Moreover, underneath the Kii Peninsular huge amounts of dehydrated fluids, coming up across the Moho plane, probably cause frequent tremor, which is ultraslow seismicity, also called silent earthquakes (Obara, 2009). Recently, this type of earthquakes has been traced by a better developed network of seismic stations. Those earthquakes suggest continuous fluid migration causing quick type subduction without any severe friction. Since the Miocene, the Kii peninsular has experienced a very fast uplifting to provide over 2000 meters high mountain building.

Stage 4. Exhumation and mountain building. After rapid exhumation, a part of regional metamorphic rocks is locally exposed to initiate mountain building (Fig. 9d). Modern analogue is the Olympic peninsula off Seattle at the western coast of North America. The region of the Olympic mountain is a piece of a ca. 10 Ma old Juan de Fuca plate, which is now subducting to form a cascade of calc-alcaline volcanic fronts or continental arc along the western coast of the United States. A non-volcanic outer arc is extended from Vancouver Island to the Olympic peninsula in the south (Fig. 10). That non-volcanic outer arc is 100 to 200 km wide with a 2428 m high peak and east dipping normal fault at its eastern margin (second fault). In the western part there is another fault, probably the first fault of the paired faults corresponding to an older Pacific-type orogenic belt. In between those faults there is a lawsonite-bearing weekly crystallized original metamorphic belt, which Miocene age was constrained by the fission-track analysis of zircon separated from lawsonite-bearing metamorphic schist (Brandon and Calderwood, 1990).

Modern analogue is the Shimanto Pacific-type orogenic belt in south-western Japan including accretionary complex, metamorphic belt and batholith belt. The Shimanto belt hosts the Sanbagawa metamorphic belt, which has long been investigated, mostly in terms of metamorphic petrology. Recently, it has been shown that it is a composite belt consisting of a high-grade eclogite part representing structural top, namely, the Sanbagawa belt, and a lower-grade epidote-amphibolite metamorphic belt. This is the youngest Pacific-type metamorphic belt in Japan eclogite facies (Aoki et al., 2008). Zircons separated from Sanbagawa quartz-bearing eclogite were analyzed by SHRIMP to yield an age of ca. 100 Ma (Okamoto et al., 2004). The lower grade metamorphic rocks contain magmatic zircons with oscillatory zoning, which yielded an age of ca. 80 Ma. Metamorphic white mica phengite was analyzed by K-Ar method to yield an age of 65 Ma, consistent with the 80 Ma age of magmatism (Aoki et al., 2007, 2008). Thus the history of the Sanbagawa metamorphism can be summarized as follows. (1) The pre-metamorphic sedimentary rocks became mixed and formed a thickened packet in the vicinity of an ancient trench through a variety of subduction-related tectono-sedimentary processes, probably in Early Cretaceous time (ca. 140-130 Ma). (2) The subducted protoliths underwent progressive metamorphism up to eclogite facies reaching a maximum depth of ca. 60 km in late Early Cretaceous time (ca. 120-110 Ma). (3) The high-P/T metamorphic rocks began to rise toward the surface (during the interval 110-50 Ma) with minimum estimates for the average cooling rate around 9-12°C/Ma and an average uplift rate around 0.6-0.5 mm/year. (4) Finally, at some stage after reaching the erosional surface, the high-PIT metamorphic rocks were covered unconformably by the middle Eocene clastic deposits (c. 50-4- Ma) (Isozaki and Itaya, 1990).

Stage 5. Exposure of metamorphic core. The highest grade eclogite facies rocks are exposed over a distance of more than 600 km: from the Kanto mountains near Tokyo to the Shimanto metamorphic belt. The SW Sanbagawa, Shikoku and Kii Peninsula metamorphic rocks are of much lower metamorphic grade, corresponding to glaucophane and pumpellite-actinolite facies are exposed very locally (Aoki et al., 2010).

The TTG granites and associated felsic volcanic rocks of about 65 Ma age are exposed more to the north, forming a 200 km wide and more than 1000 km long EW-trending granitic belt along the coast of the Sea of Japan. The Shimanto metamorphic belt and the granitic belt are separated by fore-arc deposits of 83.5 to 65.5 Ma age. The structural bottom of the Shimanto regional metamorphic belts is non-metamorphosed Shimanto accretionary complex. The boundary between the metamorphic belt and the accretionary complex is called Butuzo tectonic line.

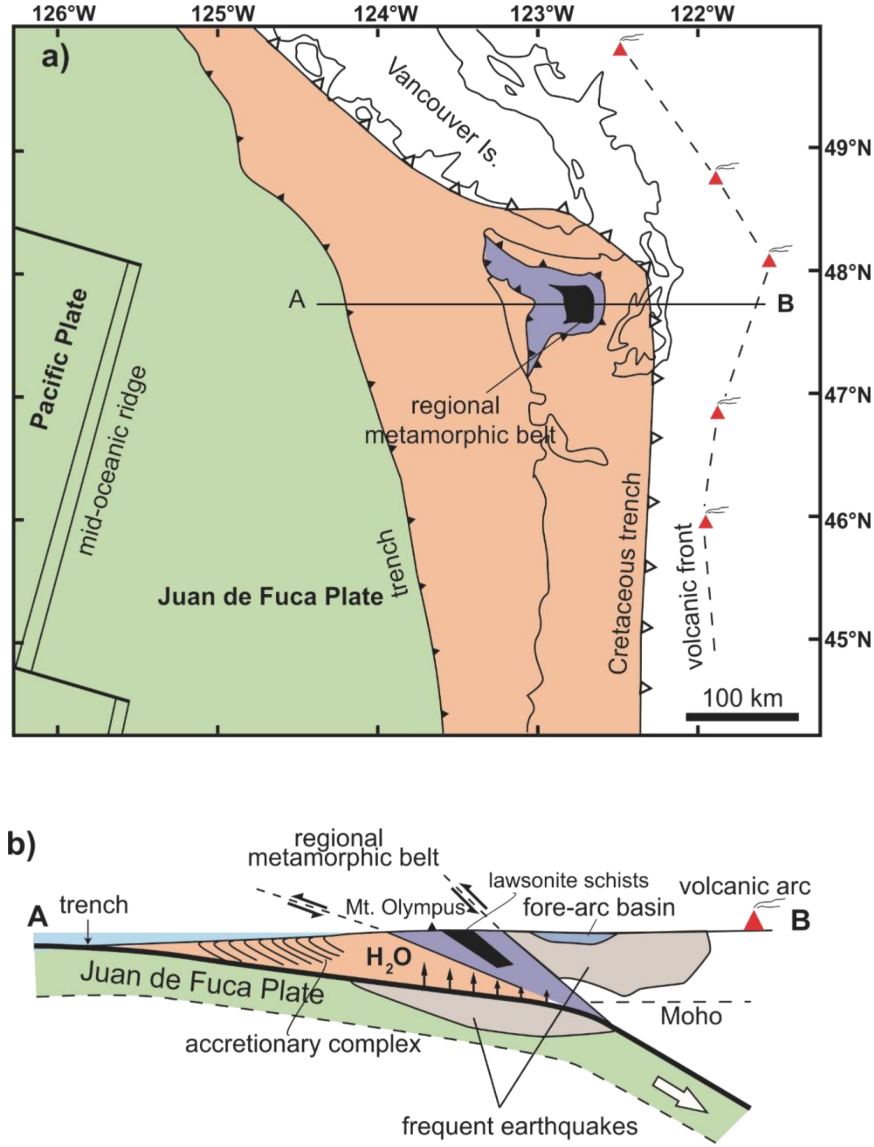


Fig. 10. The non-volcanic outer arc of the Olympic Peninsula, NW USA, which exposes the world's youngest high-pressure schist belt (Brandon and Calderwood, 1990).

Stage 6. Post-orogenic arc magmatism, retrograde metamorphism, deformation and sedimentation. Modern analogues are Mariana, Japan and Nankai trenches and related arcs. The final stage of Pacific-type orogeny includes the surface exposure of regional metamorphic belts, related tectonic movements, culminating arc magmatism and tectonic erosion, a geological phenomenon typical of P-type belts (Fig. 7b). Tectonic erosion destroys an already formed Pacific-type orogeny. Therefore, it is hardly possible to find/explore an older P-type belt with fully preserved structure. The major constituents of older Pacific-type orogenic belts are much stronger deformed and fragmented. The modern Mariana, Japan and Nankai trenches are sites of ongoing extensive tectonic erosion. This corresponds to the post-orogenic movement of the orogen (Hirata et al., 2010).

2.6 Collision-type orogeny and its modern analogs

In the 1960-ties, as world standards of Collision-type orogeny Dewey (1969) proposed Dalradian orogenic belts in Scotland and Ireland formed by the collision of North America and Baltica. A bit later, Dewey and Bird (1970) changed the idea and focused on the Himalayas, i.e. a much younger C-type orogenic belt. In the case of the Himalayan orogenic belt the continent collision started at about 50 Ma and still continuing (Kaneko et al., 2003). Below we consider eight stages of C-type orogeny. There are less examples of C-type orogeny compared to P-type orogeny. The eight stages of C-type orogeny include not only continent-continent collision, but also continent-arc collision. The present-day Himalayan mountain range is Stage 4.

Stage 1. Initial stage of continent subduction. Progressive regional metamorphism. During collision one big continent may subduct under another big continent, e.g., the subduction of the African plate beneath the Eurasian plate will soon close the Mediterranean Sea and form a new collision zone. The Arabian Plate has been subducting beneath also the Eurasian Plate since 20 Ma and the Indian Plate has been subducting under Eurasia since 50 Ma. There was a big ocean, oceanic subduction and related Pacific-type orogeny before the India-Eurasia continent collision. North of the Himalaya Mountains, there are Cretaceous Pacific-type blueschist belt parallel to the 200 km wide Indus-Tsampo suture and adjacent huge batholith belt.

The progressive regional metamorphism related to the India-Eurasia collision formed the 300-400 km wide Himalayan UHP-HP metamorphic belt running parallel to the northern continental margin of the Indian continent over a distance of more than 3,000 km. This UHP-HP belt is roughly equivalent to Pacific-type belts. The belt had long been considered as formed by intermediate-pressure progressive metamorphism at 10 kb and T=700-800℃. However, Kaneko et al. (2003) discovered coesite-bearing eclogite and associated UHP metamorphic rocks from the highest metamorphic grade overprinted by retrogressive hydration metamorphism along the kyanite-sillimanite facies series, and constrained its timing by dating zircon inclusion in coesite which yielded 48 Ma. The geology of UHP mineral-bearing high-grade metamorphic rocks indicate that their protoliths are continental shelf deposits formed during the rifting stage before the separation of India from Antarctica at 100 Ma. Therefore, those metasedimentary rocks are different from the protoliths of Pacific-type. The Himalayan metamorphic rocks formed after impure carbonate which probably subducted to a depth of at least 100 km at 48 Ma. It took the carbonates about 1.4 Ma at a speed of 10 cm/year to travel from the trench to the 100 km depth at a subduction angle of 45 degrees. However, based on the magnetic stripes in the Indian Ocean, the rate of subduction was two times faster than after the India-Eurasia collision, therefore it may take only 0.7 Ma to transport the sediments from the trench to the depth of coesite stability (100 km). Another young example of C-type collision is the ca. 20 Ma collision of South Arabia with Eurasia to form the Zagros Mountain Range (Safonova and Maruyama, 2014). There are also examples of continent subduction under island arc like the subduction of Australia under the Banda arc since the Miocene (Fig. 8).

Stage 2. Ascent of UHP-HP metamorphic belts to mid-crust levels. The exhumation of UHP-HP rocks was followed by extensive Barrowian-type retrograde metamorphism including rock hydration, recrystallization, and migmatization. Fore-arc basin and foreland fold-and-thrust belt also formed at that stage. Modern analogue is the eastern part of Banda non-volcanic outer arc of Indonesia (Figs. 8 and 11). There we can trace the subduction of the Australian continent beneath the arc. The northern leading edge of the subducting Australian continent can be detected to depths of 200-300 km at a distance of 1200 km north of the Java trench (Fichtner et al, 2010; Fig. 11). However, on the surface we see no signs of such collision and indentation. The regional metamorphic belt has not get exposed above the sea level yet, however, the supra-subduction magmatism stopped in Miocene time, and the area between the trench and the former volcanic front is present as regionally elevated non-volcanic outer arc. Between the volcanic and non-volcanic arc there is a topographically well-defined fore-arc basin. To the east, there is a so-called Weber Deep, a 5-6 km deep and 100 km wide oceanic basin. This fore-arc basin is related to the Tanimbar Island of the non-volcanic outer arc (Figs. 8-13).

Stage 3. Early exhumation of regional metamorphic belt and formation of dome structure. Fore-arc basin sedimentary facies are typically coarse-grained. In the same area between Indonesia and Australia, fore-arc basin sedimentary facies turn from deep-sea fine-grained facies (Weber Deep) to coarse-grained facies close to the island arc. At the eastern end of the Indonesia Banda arc there is a small island, Laibobar, which is about 5 km across and elevated to a height of 390 m. The island consists of Quaternary coral reef deposits separated from the underlying crystalline schists by a nearly sub-horizontal fault (Ota and Kaneko, 2010). We cannot observe lower units, however, if we consider the geologic structure to the west of the Laibobar Island, we understand that this area is just a top of exposed regional metamorphic belt, which dips to the north under the Weber Deep (Fig. 12). Although very deep the Weber Deep represents a fore-arc basin, because it is located between the northern volcanic arc and the southern non-volcanic outer arc. When was that fore-arc basin formed? No details still available because no deep-sea drilling has been performed there yet. However, the Weber Deep must have oceanic crust simply because this basin tends to become narrower to the west and disappears in the area north of Sermata. The distance between the arc and the trench decreases from north to south, and the structural top of the outer arc represents fore-arc ophiolite (Ishikawa et al, 2007; Ishikawa, 2011). There is a small island Dai at the south-western end of the Weber Deep. The Dai Island is underlain by the gabbroic rocks only of the lower fore-arc oceanic crust. Further to the west of the Weber Deep, the fore-arc basin disappears by a NS-compression, suggesting that that deep basin have oceanic crust plus underlying mantle. Typically a fore-arc basin formed by ocean floor spreading and consisting of oceanic crust and underlying mantle represents island-arc ophiolite, like, for example, the Cretaceous Coast Range Ophiolite in California. Therefore, the Banda Arc of Indonesia is a modern analogue of fore-arc ophiolite in general and records its emplacement in particular. Probably, the oceanic crust of the Weber Deep formed in the Miocene.

The ascending regional metamorphic rocks push the fore-arc ophiolites up. The protoliths of the metamorphic rocks are continental shelf deposits formed during the Permian-Triassic boundary rifting. Those protoliths must be subducted to a depth of at least 30 km and

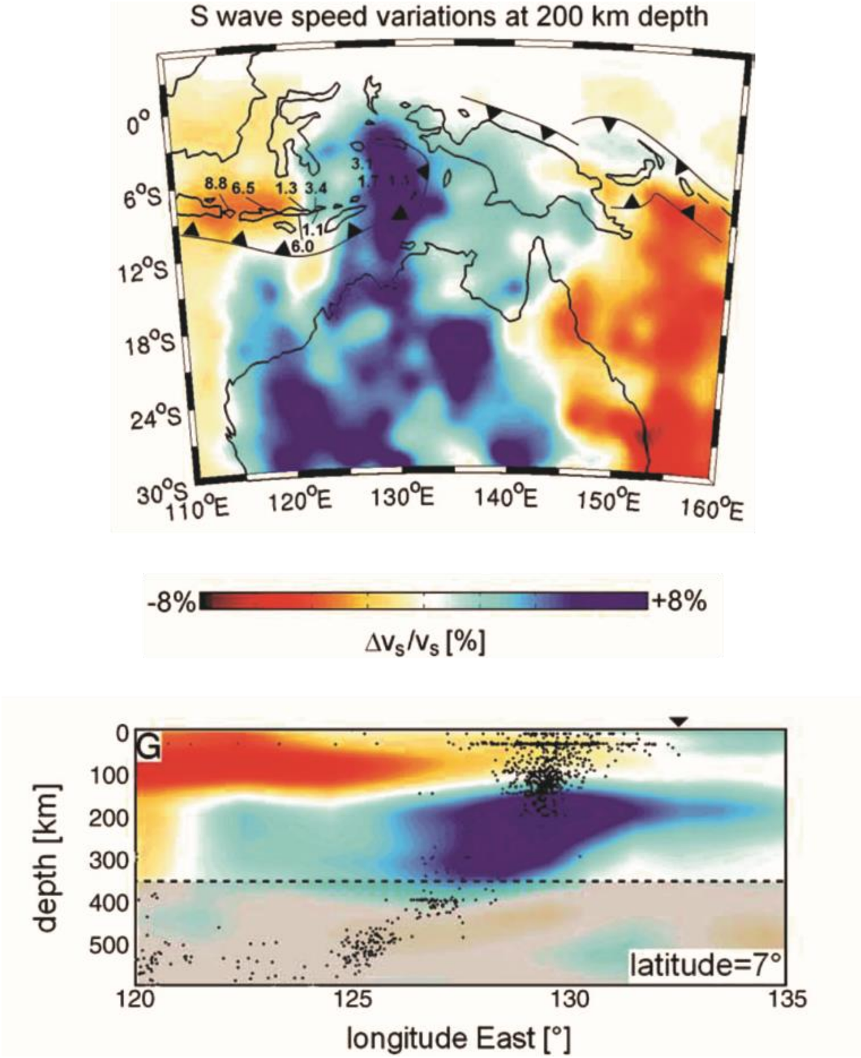


Fig. 11. Non-volcanic outer arc of the Banda arc, eastern Indonesia. Subduction of the Australian continent under the Miocene arc has already stopped (Kaneko et al., 2007). S-wave mantle tomography identifies the subducted Australian tectosphere down to a depth of 200 km to the north of the Eastern Arm of Celebes Island from the trench at 12°S for over 1200 km beyond the Weber Deep (Fichtner et al., 2010). The eastern parts of the two arcs further rotated anti-clockwise along the westward vector of the Pacific plate motion (Hamilton, 1979).

subjected to low-temperature, high-pressure regional metamorphism to yield glaucophane-bearing mineralogy and then returned back to the surface. However, a full dome-like structure has not formed yet. Therefore, the secondary normal fault has not fully formed either. However, west of Timor Island, there is a well-developed dome-like structure. Timor is a big island and the regional metamorphic belt is already elevated to a height of 3000 m. Timor back thrusts are well developed along the northern margin of the volcanic arc to form a transitional region from the point of the disappearance of Weber Deep in the north-east and farther to the south-west (Figs. 12, 13).
In Indonesia, the fore-arc ophiolite rests on the Timor regional metamorphic belt, over Leti, Moa, Sermata and Dai islands (see star marks) along the non-volcanic outer arc. The Weber Deep, a 150 km wide and 5-6 km deep fore-arc basin with oceanic crust, is present east of Dai (Fig. 12a).

North of the arc chain, at the northern foothill of the Java-Banda volcanic arc, there is an active back-thrust zone, although the volcanism stopped after the beginning of the Australia-arc collision. Fore-arc spreading formed the Weber Deep and its western extension, which are now exposed as ophiolites on the non-volcanic outer arc from Timor to Dai islands (Fig. 12b). At ca. 20 Ma Australia was located about1600 km south of the present-day position, and the Timor-Banda trench was located more to the south and had an EW strike. The collision of the arc with Australia bended the trench to the norths and closed the Weber Deep fore-arc basin from the west to emplace ophiolites on the non-volcanic outer arc. The eastern parts of these two arcs then rotated anti-clockwise by a westward directed vector of the movement of the Pacific plate (Hamilton, 1979).

<u>Stage 4. Slab break-off under non-volcanic outer arc.</u> Seismologically we can detect the subducted Australian continent, probably containing Archean continental crust, farther to the north of the volcanic arc. The Australian continent connection with the oceanic slab can be roughly estimated, because of unusual seismicity in this segment, where the seismicity plane parallel to the Benioff seismic plane is cut by a perpendicular segment suggesting oceanic slab break off (Fig. 13).

The anomalous distribution of seismicity suggests that the slab break-off is now ongoing as was originally suggested by Osada and Abe (1981). The slab break-off means that the force of slab-pull is now decreasing or fully disappeared, at least at the eastern half of the Indonesian non-volcanic outer arc. Therefore, we can suggest the

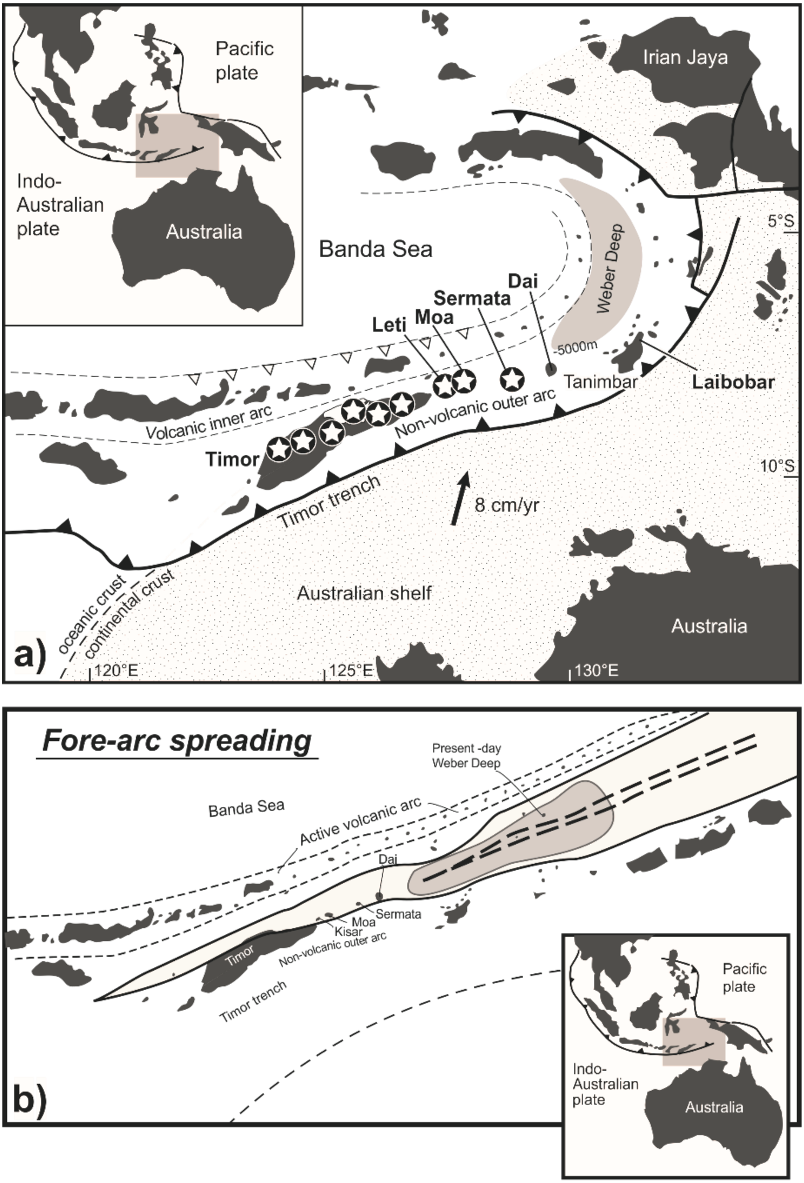


Fig. 12. Scheme of the Indonesian outer arc showing world's main type locality of island arc ophiolites formed by fore-arc spreading in the Miocene.

exhumation of high-pressure regional metamorphic rocks by the buoyant continental crust. The model of “wedge extrusion” can explain this processes of exhumation (Maruyama, 1996), which was also applied to the Himalayan Belt by (Kaneko et al., 2007).

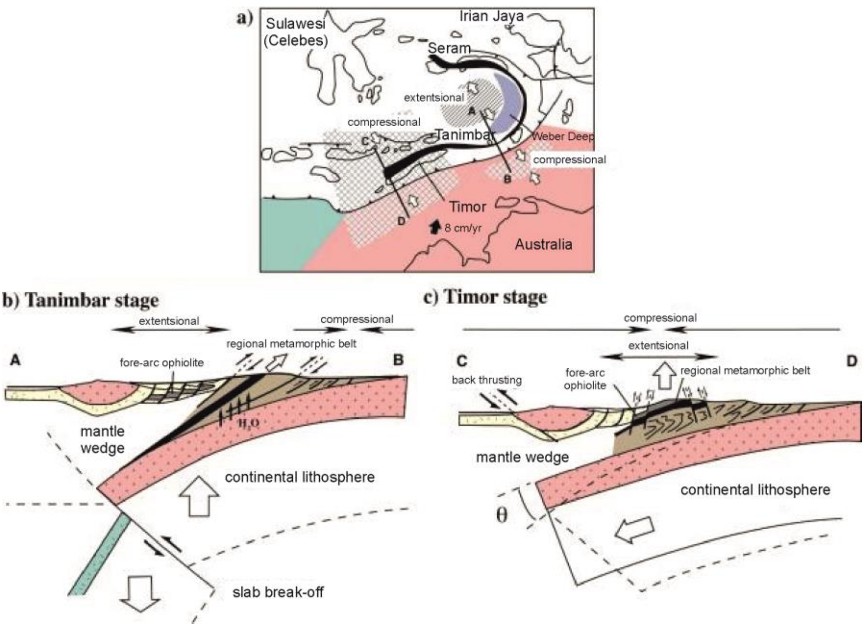


Fig. 13. An example of continent-arc collision in the Australia-Indonesia region accompanied by slab break-off, exhumation of high-pressure regional metamorphic rocks by “wedge extrusion” and finally by doming.

Stage 5. Second stage exhumation. Final emplacement of regional metamorphic rocks to mid-crustal levels. Modern example: the eastern part of the Indonesian outer arc including a small chain of Kisar, Leti, Moa, Sermata and Dai islands. The Indonesian outer arc includes larger Sumba and Timor islands (elevated to 3000 meters) in the west and a chain of smaller non-volcanic islands in the east. The smaller islands, Kisar, Leti, Sermata and Dai, are elevated to less than 200 m. The first three islands have similar stratigraphic sequences (top to bottom): island-arc ophiolite, glaucophane-bearing metamorphic rocks, non-metamorphosed or weekly metamorphosed continental shelf deposits. The shelf deposits originally formed at the northern margin of the Australian continent. The Dai island consists of island-arc ophiolite on the top, no exposed peridotites though, and metamorphic rocks underneath.

The Leti Island is underlain by progressively metamorphosed regional metamorphic rocks with glaucophane-bearing high-pressure, low-temperature minerals rocks. Figure 14 shows the results of petrological and chronological studies including metamorphic zoning in the non-volcanic second arc made from low- to high-grade rocks through mineral isograds and mineral reactions inferred from microprobe analysis of rock-forming minerals (Kadarusman et al., 2010). The north-south cross section reveals clearly delineated internal thermobaric structure. Ophiolite rocks form the structural top in the north. Underneath the fore-arc ophiolite there is a normal fault gently dipping to the north. To the south there is a regional metamorphic belt showing upgrading transition from blueschist-greenschist to amphibolite facies toward the structural intermediate. The structural intermediate of the regional metamorphic belt marks the peak of metamorphism, i.e., the highest pressures and temperatures, which and gradually decrease to the north and south. The structural bottom is a sequence of non-metamorphosed platform sediments, deposited on the rifted margin of the Australian continent. Glaucophane was first reported by (de Roever, 1940), but later Abe Kadarusman and co-authors (Kadarusman et al., 2010 reported that glaucophane-bearing schists formed after basalt and the presence of jadeite in addition to glaucophane. They demonstrated several metamorphic facies: pumpellyite-actinolite, blueschist, greenschist and epidote-amphibolite. This low-temperature high-pressure facies series, i.e. blueschist assemblage, is a characteristic feature of subduction related metamorphism. The K-Ar hornblende age of the Leti metamorphism is 16.5 Ma (Fig. 14). However, the progressively metamorphosed minerals suffered retrograde metamorphism later (hydration recrystallization) resulting in the 5.4 to 6.0 Ma K-Ar age obtained from white mica and biotite. The retrograde metamorphism of the Barrowian type intermediate-pressure metamorphic facies probably occurred at ca. 10 Ma, soon after the peak of progressive glaucophane subduction-related metamorphism (Kadarusman et al, 2010).

Thus, the progressive dehydration metamorphism occurred at about 20-16 Ma near the Moho depth. The metamorphosed sediments deposited in a trench now located more 1200 km to the south. Later, the trench was removed by the bending due to the continent collision. Meantime, the amphibolite facies regional metamorphic rocks formed after trench sediments ascended to the mid-crustal level, i.e. to ca. 10-15 km and then stopped. The rate of exhumation was about 1.5-2 mm/year only because of the continuing subduction of the Australian continent.

The continuous reactions of dehydration provided the migration of fluids, which caused extensive hydration of the high-P/T metamorphic assemblage. The trinity succession of the Leti Island, ophiolite - regional metamorphic rocks - non-metamorphosed platform sediments, has a typical sandwich structure, which is cut by secondary normal faults. Therefore, the island itself has been emplaced over the surrounding regions by the secondary normal fault. The Leti Island is an example of the second stage exhumation by the doming, which made the whole island chain. This event can be demonstrated by the Quaternary reef limestone, which has been already domed up to several tens of meters. North of the volcanic arc there is a back thrust found on the ocean floor by (Hamilton, 1979).

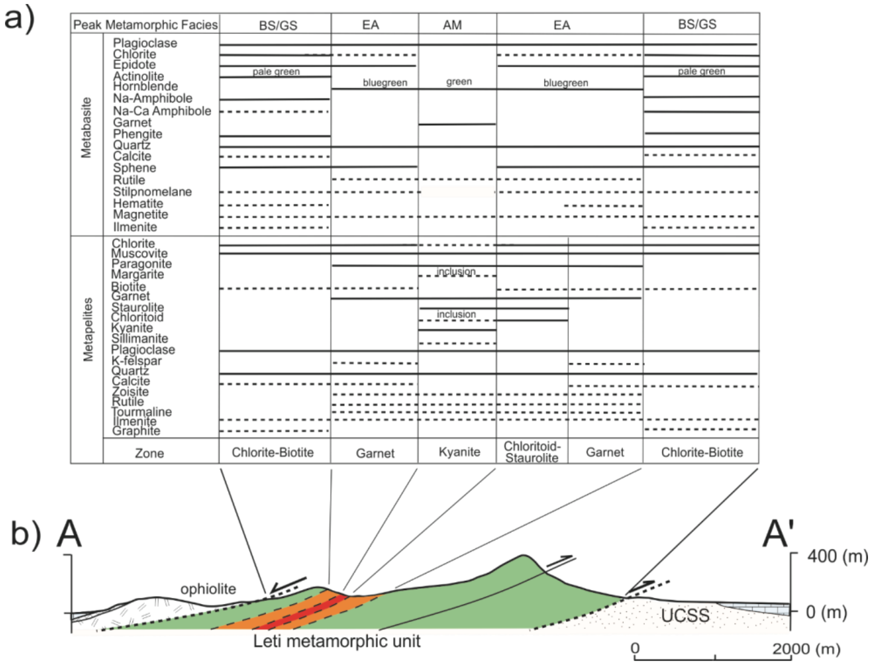


Fig. 14. A NS section across the Leti Island in Indonesia showing the sandwiched sub-horizontal structure of the orogenic core that was metamorphosed along the high-P/T metamorphic facies series in the Miocene (Kadarusman et al., 2010). The structural top and bottom are Miocene fore-arc ophiolites and non-metamorphosed sedimentary units of UCSS (upper crust shelf sediments). A thin 1-2 km metamorphic unit exhibits a symmetric thermobaric structure with a core of amphibolite facies surrounded by eclogite-amphibolite and blueschist/greenschist facies. The upper figure shows the appearance and disappearance of metamorphic minerals in both pelitic and basic schists in relation to the increasing metamorphic grade.

Stage 6. Surface exposure of the orogenic core of regional metamorphic belt. Start of mountain building. Schematically this stage of collision-type orogeny is constrained by two major processes: metamorphic wedge extrusion (Fig. 15a) and doming, i.e. orogeny sensu stricto (Fig. 15b).

In Indonesia, the most advanced state of continent-arc collision stage is the western Timor, where the mountain range reaches ca. 3,000 meters above the sea-level. The rock succession is the same as east of Timor, i.e. at the chain of smaller islands of the non-volcanic outer arc, i.e., fore-arc ophiolite, regional metamorphic belt, and non-metamorphosed platform sediments (Figs. 12, 13). The only difference between the small islands and the western Timor is the development of secondary high-angle normal faults cutting the sub-horizontal trinity of the rock units and the high-rate of mountain building. In Timor, the Quaternary reef carbonate has been uplifted to 1260 meters above the sea-level during the last 1.2 Ma suggesting a rate of 2-3 mm/year. This value corresponds to one of the world highest uplifting in the Himalayas (Kaneko et al., 2007).

The Himalaya Mountain Range is special for the presence late-orogenic and post-orogenic sediments caused by the India-Eurasia continent collision (LeFort, 1975). The Himalayan mountain building started at 9 Ma (Fig. 16). Evidence for this comes from the record of sedimentary environment estimated from drilled core information obtained at the margin of Bengal Fan of deposits in the Indian Ocean (Derry and France-Lanord, 1990). The Himalaya Mountains are still growing as shown by GPS measurements on the top of the Himalayas.

Before the 2000-ties, the concept of Himalayan orogeny included five assumptions. 1) The Indian continent collided with Asia at 25 Ma. 2) Granitoid crust beneath India was thrust under Asia to reach a thickness of 60 km accompanied by coeval regional medium P/T metamorphism. 3) The duplicated granitoid crust gradually uplifted by its buoyancy to expose the tectonic boundary of the lower unit of the subducted Indian continent. 4) The surface cover sediments above the basement rocks of the Indian continent experienced high-grade metamorphism overburden by the Asian continental crust (England and Thompson, 1984a, b). 5) Mountain building started at the same time as the collision, i.e. at 25 Ma. However, recent results have shown that the subduction of the oceanic lithosphere under southern Eurasia before the collision caused the formation of a Pacific-type orogenic belt and, therefore, we can highlight five important constraints on the Himalayan orogeny (Fig.16). (1) Subduction of the Indian continent reached depths of ca. 100 km and the metamorphosed units returned back to the surface. (2) The progressive regional metamorphism was not of medium-pressure, but high-P/T type. (3) The age of progressive metamorphism appeared not 25 Ma, but 48Ma. (4) The rocks ascended to the mid-crust depths at 25 Ma. (5) The Himalaya major mountain-building is not linked with the exhumation of the regional metamorphic belt. The mountain building started at 9 Ma and is still on-going (Fig.16d). The timing of 9 Ma was the period of formation of high-angle normal faults (Kaneko et al., 2003).

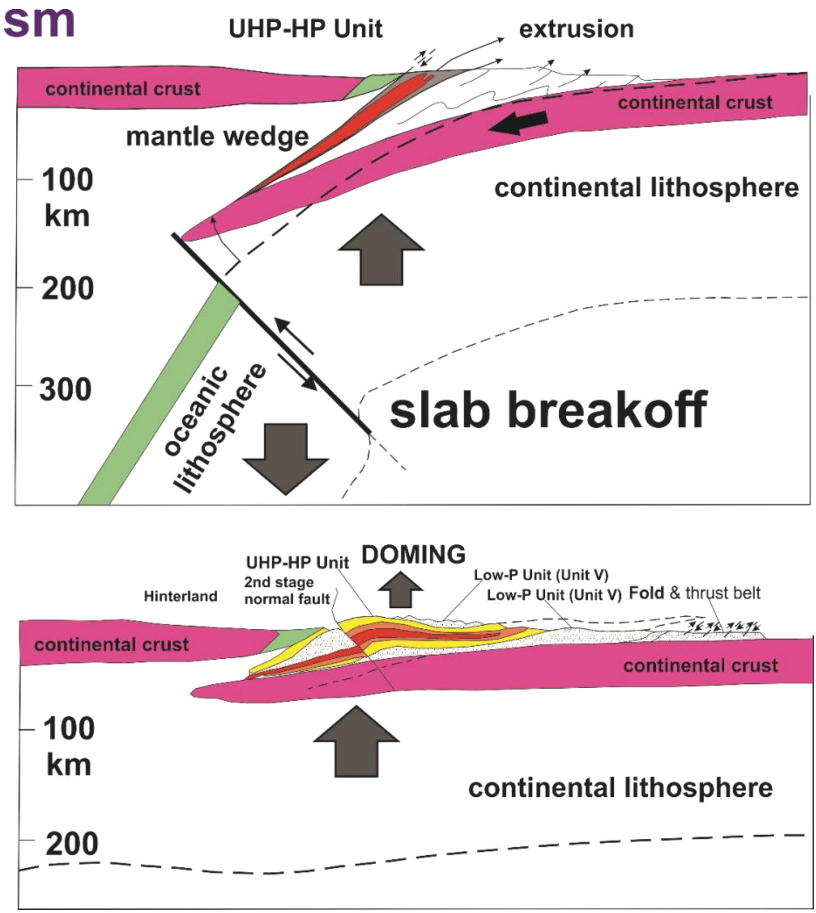


Fig. 15. Extrusion of the metamorphic core and doming to initial collision-type orogeny.

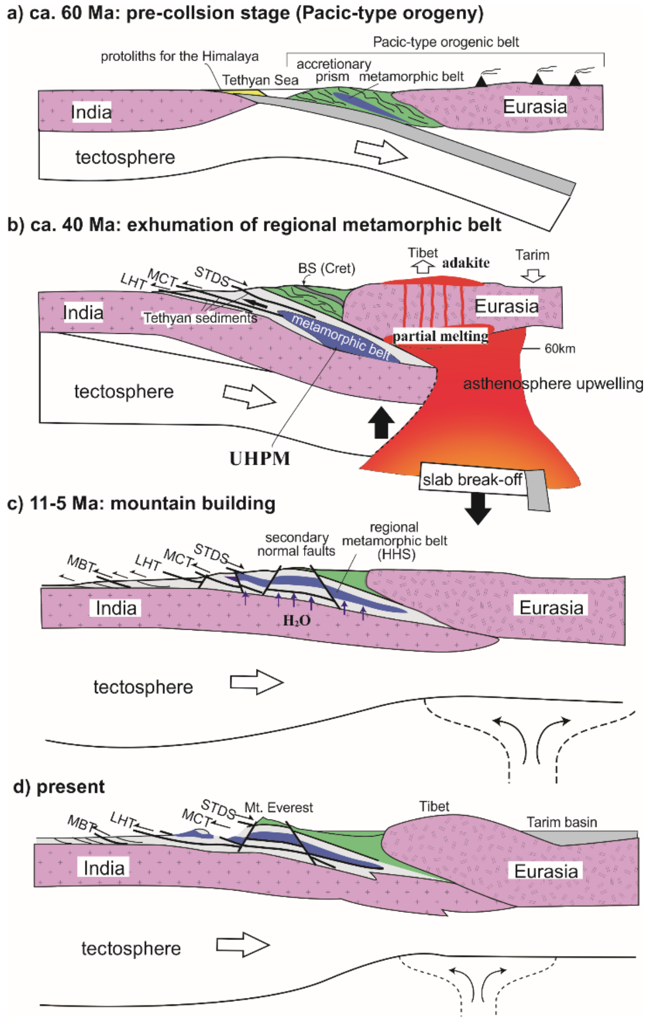


Fig. 16. Tectonic evolution of the Himalayan orogenic belt (Kaneko, 1997; Maruyama et al., 2010). (a) Subduction and collision of the Indian continent since 50 Ma (Kaneko et al., 2003) caused UHP-HP metamorphism (48 Ma) in the Himalayan orogenic belt, and the subsequent slab break-off triggered the asthenospheric injection under Tibet to form adakitic magmas. (b) The tectonic transportation of the UHP-HP belt as a thin slab up to mid-crustal levels by 25 Ma was followed by retrograde hydration recrystallization along the Barrovian-type intermediate geotherm did not affect the mountain building of the Himalayas. (c) The underplating of huge amounts of sediment formed in a foreland-fold-and-thrust belt jacked up the sandwiched tectonic units of the Himalayan orogenic belt at 11—5 Ma. (d) The Himalayan mountain building started in the Late Neogene and is still continuing. LHT – lower Himalaya thrust; MBT – main boundary thrust; MCT – main central thrust; STDS – south Tibet detachment system.

Another example is Taiwan Island, which is as large as Kyushu Island in Japan (Fig. 17). Taiwan is a continental fragment of the South China (Yangtze) Craton, which was rifted by the South China Sea during Eocene-Miocene time. The basement rock and its platform cover sequence have been subducting under the Luzon arc, i.e. in the eastern direction, since 15 Ma (Huang et al., 2006). The metamorphosed sediments were extruded NW-ward into the platform cover sequence followed by the still on-going mountain building. Nevertheless, the mountain-building is restricted to the NE corner and has not yet exposed the orogenic core of high-grade rocks metamorphosed at 700-800℃ and P>20 kbar (Fig. 17).

The timing of the collision is 6.5 Ma (Lin et al., 2003). The structural top of the Luzon arc is ophiolite exposed in the lowland east of the high mountain range. The lowland is separated from the range by a NE-SW trending strike-slip fault. The island-arc ophiolite is exposed in the Coastal Range.

<u>Stage 7. Post-orogenic intrusion of A-type granites.</u> This stage is characterized by intrusion of small typically A-type granitic bodies. The India-Eurasia collision caused the injection of astenospheric material underneath and related heating of the lower crust, which produced A-type granitic magmas. The granites intruded the lower reverse fault of the UHP-HP belt. A similar event was recorded east of Papua-New Guninea, on a small island called D'Entrecasteaux, where eclogite is trapped by post-orogenic intrusions of granites, suggesting the presence of a UHP belt underneath. In case of Pacific-type orogeny, the opening of the Sea of Japan Sea in the Miocene resulted in the asthenosphere upwelling under the eastern margin of Asia and opening of a back-arc basin to form the Japanese islands in Pacific Ocean. However, that process of extensional orogeny could be applied to continental rifting, e.g., the East African Rift system, which was initiated in Miocene time or even much older back to the Cretaceous.

In case of continent collision, the eruption of andesite-dacite-adakite magmas in the Tibetan plateau could be related to the partial melting of the lower mafic crust at a depth of ca. 70 km depth. That produced adakite in the thickened continental crust at high-pressure conditions induced by strong lateral compression resulted in the uplifting of Tibet (Chung et al., 2003). The formation of adakite could be a result of oceanic slab break-off and collapse into the deep mantle and related astenospheric injection (Molner and Tapponnier, 1975) (Fig. 15a, 16b).

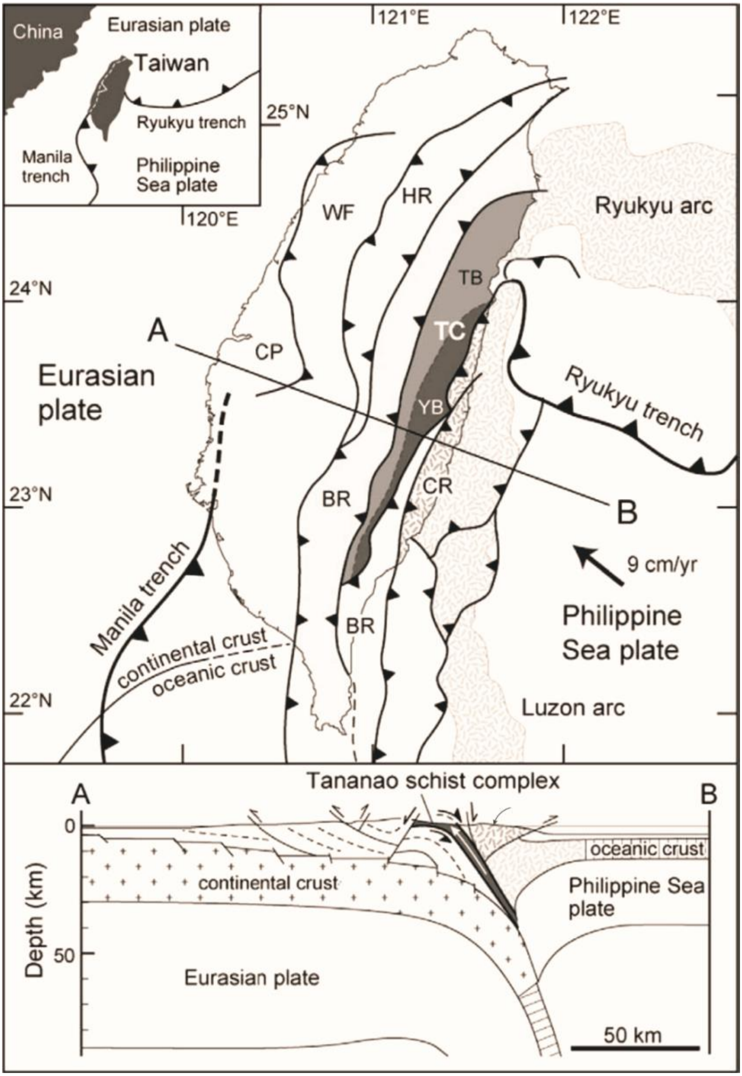


Fig. 17. The Miocene island arc ophiolites and glaucophane high-pressure schists in Taiwan (Beyssac et al., 2008; Ota and Kaneko, 2010) and the underlying collisional margin of Asia (Lin et al., 2003). The passive margin platform-type sedimentary units subducted eastward under the intra-oceanic island arc of the Philippine-East Taiwan, and collided to yield the Miocene Tananao schist belt along the collision zone. The present-day high elevation of the central highland (HR) composed of passive margin deposits of the Eurasian continent could be due to the mantle upwelling, which is a southward extension of the active back arc basin of the Okinawa trough since 5 Ma.

Those episodes asthenosphere upwelling in the Himalayas could be rather post-orogenic, probably occurred at a post-orogenic stage as the distance from the Himalaya mountain range is too far for that magmatism to be regarded as a phenomenon of orogeny. In general, C-type orogenic belts may host those very small granite intrusions, but C-type orogeny by itself can hardly produce new continental crust. The deformation related to C-type orogeny may occupy vast regions in front of collision, i.e. in the counterpart colliding continent, Eurasia, although no deformation in the indenting Indian continent (Fig. 16c,d).

<u>Stage 8. Post-orogenic extension and back-arc basin opening</u>. In some regions, C-type orogeny finally may result in back-arc basin opening and/or formation of extensional basin in a post-orogenic. The recent examples are the European Alps and the spreading of the Mediterranean Sea (Fig. 8). The Alpine C-type orogeny caused by the collision of Europe and Africa continent culminated in the opening of small ocean basins or rifted basins in the Mediterranean Sea. The southward subduction of the European continent under the African continent started and reached depths of 60 to 100 km. Evidence for this comes from the occurrence of blueschists with coesite and pyrope, which formed after continental shelf sediments (Chopin, 1984). Later, at 40 to 50 Ma, the metamorphosed continental shelf sediments returned back to mid-crustal levels. The extensive retrograde metamorphism accompanied by hydration recrystallization formed Barrowian-type mineral assemblages. The exhumation of metamorphic rocks induced the doming and growth of the Alpine mountains in Miocene time. The final episode of magmatism formed post-orogenic small intrusions of alkaline granitoids at 15 Ma. The succession of events was nearly identical to that of the Himalayan orogeny. More evidence for this comes from the major units of the Alpine orogenic belt, which includes the Penninic nappe of UHP-HP metamorphic rocks in a structurally intermediate position (Fig. 18). The overlying unit is the Austro-Alpine nappe, which was a part of the African continent. The structural bottom is the Helvetic nappe, a part of the European continent. These three structural units were tectonically juxtaposed by paired faults in Miocene time. Later, all of them were cut by high-angle secondary normal faults to dome-up and make the Alpine Mountain Range (Coward and Dietrich, 1989). This structure is not unique; it is typical of all orogenic belts. However, the scale of C-type belts is larger than that of P-type belts. In case of Alpine foldbelts, one of high-angle secondary normal faults appeared to be large a scale strike-slip fault as boundary of microplates. A typical example is the Innsbruck Line (Fig. 18).

The western part of Mediterranean Sea is underlain by less than 20 km thick continental crust. An example of a back-arc basin opened that time is the Tyrrhenian Sea west of the Italian Peninsula. Such a post-orogenic event of the Africa-Europe collision is different from the India-Eurasia collision. The latter formed no back-arc basins in Asia. In Europe the line of the back-arc opening in the Alpine region is transverse to the strike of the Alpine mountain belt and has a complex meander shape. The EW-trending western Mediterranean Sea was rifted and thinned for form minor back-arc basins with oceanic crust. The eastern part of the Mediterranean Sea was a remnant oceanic plate subducting to the north under the Aegean Sea (Fig. 18). Therefore, the asthenospheric mantle upwelling occurred only in the western Mediterranean Sea, across the Alpine mountain belt, suggesting a kind of interaction with the opening of the Gulf of Biscay in the Atlantic and the rotation of the Italian Peninsula.

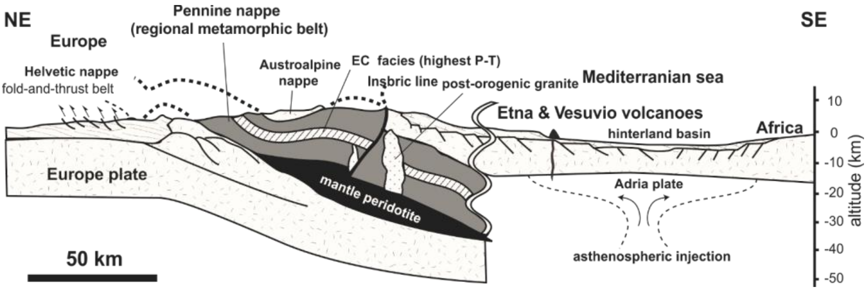


Fig. 18. Post-orogenic extension of the European hinterland of the Alpine orogen after the Africa-Europe collision (Coward and Dietrich, 1989; Maruyama et al., 1996).A NS cross-section of the Alpine orogen shows the structurally intermediate position of the Penninic nappe with the Alpine peridodite surrounded at the top and bottom by non-metamorphosed Austro-Alpine and Helvetic units. The thermobaric structure of the Penninic unit is symmetric with a structural core consisting of coesite-bearing eclogite facies rocks formed at a depth of ca.100 km and a temperature of 800—900℃. To the south, the western Mediterranean Sea subsided due to the post-Alpine extensional tectonics and by the adjacent asthenospheric upwelling presumably triggered by slab break-off. The active volcanoes of Etna and Vesuvius are surface manifestations of asthenospheric injections.

2.7. Models of exhumation of regional metamorphic belts

Among the major components of orogenic belt, regional metamorphic belt occupies the most important position as a core of the orogeny (Figs. 1, 7, 10, 13, 16). Moreover, the mountain building must have been solely controlled by the uplifting of crustal rocks from the mantle depth. Before the appearance and propagation of the concept of plate tectonics, Pacific-type orogeny had been discussed by few scientists, e.g., by A. Miyashiro (1961), who was the first to show a link between metamorphism and orogeny. He proposed the concept of paired metamorphic belts. We know that Pacific-type orogeny generates HP-LT blueschist belts and HP-HT eclogite belts, both may contain jadeite. Unlike collision-related orogeny, P-type belts also include huge granitoid belts/batholiths. Theoretically, those granitic massifs can initiate contact metamorphism at great depths to form mostly LP-HT andalusite-sillimanite facies metamorphic rocks. Those MT-MP metamorphic belts or zones should run parallel to the HP belts of jadeite-glaucophane type formed deep in the subduction zone, and both of those, if uplifted/exhumed, would be aligned parallel on the surface. In Japan, such a case was documented in the Ryoke granitoid belt, which is bounded by the Sanbagawa HP metamorphic belt on the oceanic side and by a lower grade metamorphic belt on the continental side (section 2.4; Figs. 3, 7). However, those paired metamorphic belts are extremely rare over the world, whereas pairs of batholith belts and blueschist belts are very common. This suggests that the simultaneous exhumation of HP-LT blueschist belts and LP-HT metamorphic belts is uncommon, although LP metamorphic belts are common beneath and around batholith belts.

In 1982 Akiho Miyashiro with co-authors released an excellent book "Orogeny" reviewing the concept of orogeny from Edward Suess and modern analogues in terms of plate tectonics (Miyashiro et al., 1982). Nevertheless, the time of publication was just before the discovery of UHP-HP rocks in C-type but also in P-type orogenic belts. Moreover, that discovery changed the understanding of intermediate-pressure metamorphism from typically C-type (as it had been considered before) to retrograde type related to the reactions of hydration as a secondary process. In relation to Pacific-type orogeny, for example, the formation of the intermediate type metamorphic rocks in the Sanbagawa high-pressure belt appeared not progressive, but a secondary hydration overprint.

Modern metamorphic petrology has gained precise pressure estimates from several collisional orogens, such as the Dalradian belt of Scotland, the Caledinian, Hercynian and Alpine orogenic belts of Europe, Hercynides and Caledonides, and the Himalayan belt in Asia. Before and still soon after the appearance of the plate tectonic concept geologists believed that a metamorphosed unit is a thick geologic body extended to the Moho depth. Formerly, the pressure estimates from those belts were at maximum 10 kbar, i.e. corresponding to intermediate-pressure type facies. This very value corresponds to the Moho depth, suggesting that the rifted basins or geosynclines in a "classic" sense first subsided deep down to 30 km, suffered high-grade metamorphism, and finally returned back to the compressional stage to make mountain building and expose Moho depth pressure rocks ay the surface. Such a scenario must imply fault-free isostatic rebound and related formation of huge amounts of detrital materials corresponding to a 30 km thick basin. Such a basin must be present nearby the Alpine mountains or at all other examples of C-type orogenic belts, because the present surface exposure of orogenic core exhibits 10 kbar and 700-800℃ metamorphosed rocks. To explain that hypothesis, England and Thompson (1984), based on the observation of the Gang delta in the Indian Ocean, proposed an elegant plate tectonics model for the Himalayan belt (Fig. 19). The geosyncline buoyancy-driven uplifting model suggests that, for example, the Indian continent must be thrust under the Asian continental crust to make 60 km thick continental crust. Then, to expose granulite facies metamorphic rocks, the upper 30 km of the Asian crust must be completely eroded away to the Indian Ocean. That model seemed to be an excellent solution of the puzzle, but the discovery of UHP rocks in the Alps and Himalaya destroyed the initial mid-pressure model. The presence of UHP-HP coesite-bearing rocks in glaucophane belts showed the maximal pressure estimated at about 26 kbar, i.e. much higher than the previous estimate of 10kb. Later, the pressure estimates increased even up to 60-70 kb, 900-1000℃, e.g., in the case of Kokchetav diamond-bearing regional metamorphic belt (Sobolev and Shatsky, 1990; Shatsky et al., 1995; Kaneko et al., 2000; Katayama et al., 2000). A very important issue is how the ascent/exhumation of metamorphic rocks proceeds and contributes to mountain growth.

De Roever (1957) was the first to notice the absence of large sedimentary basins in front of exposed HP metamorphic belts. He accordingly questioned the buoyancy-driven geosyncline concept, but suggested that there "must be a fault-contact". Recent detailed geologic mapping programs have shown that de Roevers was right. Regional metamorphic belts in both C-type or P-type orogens always occur as very thin slabs cut on the top and bottom by paired faults. A new model of wedge extrusion then appeared, which suggests high-temperature extrusion of high-grade metamorphic rocks into non-metamorphosed geologic units (Maruyama, 1990; Maruyama et. al., 1996) (Fig. 20).

In case of Pacific-type orogenic belts, J. Suppe and D. Cowen were the first to explain the exhumation of metamorphic rocks using the concept of plate tectonics (Suppe, 1972; Cowan, 1974). Geological survey works in the Franciscan complex of California allowed them to propose a two-way street model. According to that model a regional metamorphic belt was initially an accretionary complex, which was then gradually moved to deeper crustal levels by the down-going oceanic slab.

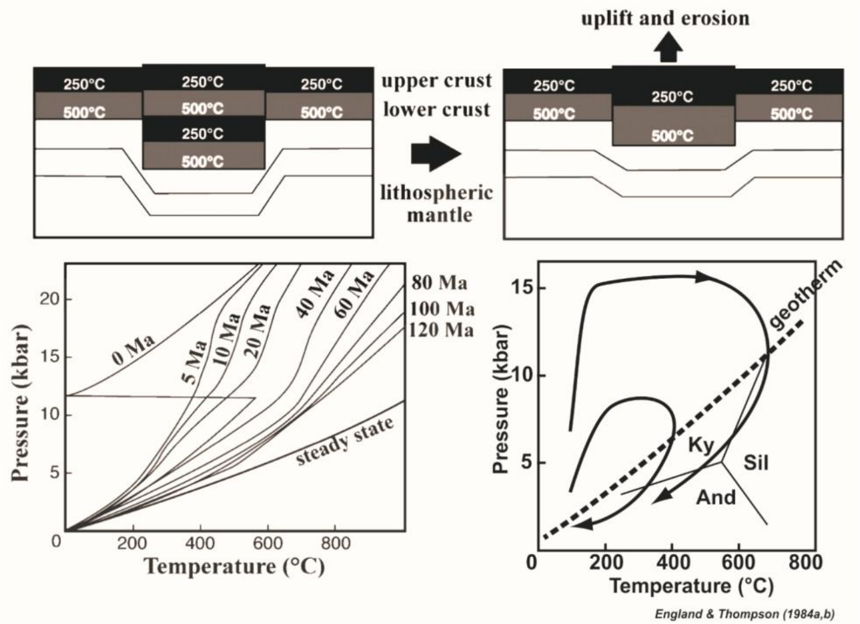


Fig. 19. A famous and often used model of the Himalayan collisional orogeny, which involves crustal duplication and buoyant exhumation of an orogenic core with a clockwise metamorphic P-T time path (England and Thompson, 1984a, b). If this model is correct, P-T must increase down to the Moho. The fact that the underlying unit of the Himalayan belt is non-metamorphosed below a subhorizontal reverse fault contradicts this model. Moreover, the discovery of coesite and/or diamond in the metamorphic belt does not support the idea of a clockwise P-T time path.

At a depth, those sedimentary units meet a strong backstop barrier at the Benioff plane. The subducted accretionary units ascend from the blueschist facies stability field to the surface along a high-angle hanging wall continuing to the surface. That was a new concept of HP regional metamorphism, in particular, in respect to Pacific-type orogeny. According to that model, the exhumation of HP metamorphic rocks (blueschist, eclogite) to the surface is discontinuous. That was a challenging idea for the traditional understanding of orogeny, which implied continuous exhumation/exposure. For example, in the Franciscan Complex of California D. Cowan (1974) tried to demonstrate that the regional metamorphism gradually changes from non-metamorphosed rocks through weakly metamorphosed to high-grade rocks without tectonic boundaries. But neither geological maps, nor so far obtained radiometric ages show a gradual transition from low-grade to high-grade rock or a continuous time span, respectively.

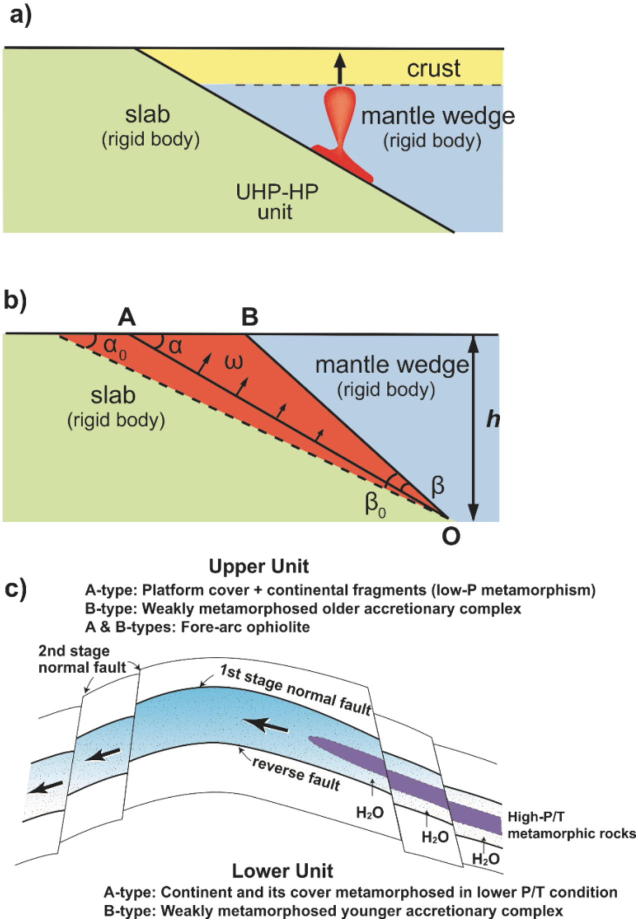


Fig. 20. A model of orogeny by wedge extrusion (Maruyama, 1990, 1997; Maruyama et al., 1996). Cjvvents see in the text.

The continuous exhumation model, which was originally developed by Suppe (1972) and Cowan (1974), later was revised by M. Cloos (1982). He was the first to interpret a serpentinite-bearing mélange belt, which is characterized by the presence of blocks of eclogite and garnet amphibolite high-grade metamorphic rocks in a matrix of non-metamorphosed or weekly metamorphosed rocks, to be derived by a process completely different from the traditionally idea of the formation of mélange by large-scale submarine land-sliding (Hsu, 1965). M. Cloos proposed a new model of “flow mélange” considering a very thin, less than 1 km, circulating fluid-enriched layer along with the subducting oceanic slab. That fluid-enriched steady state flow was assumed moving in the subduction channel, along the Benioff plane in two opposite directions, up and down. According to M. Cloos that water circulation should provide the ascent of high-grade garnet-amphibolite and eclogite from the depth of 30 km or even deeper up to the surface.

However, the field observations never supported his idea because the matrix material is very weakly metamorphosed: few, if any, lawsonite and/or lower glaucophane facies rocks may be present. The metamorphism of the prehnite-pumpellyite facies to the low-grade blueschist facies occurs at a depth of ca. 10 km and T<250℃. The matrix of mudstone definitely looks metamorphosed at lower PT conditions than that of tectonic blocks. M. Cloos explained that apparent contradiction by s shorter time, not sufficient to recrystallization. Nevertheless, the life time of the flow mélange is definitely longer than in laboratory experiments on metamorphism. The laboratory experiments usually last less than several months to one year under given high-P/T conditions, whereas the flowing of the mélange may take million years to travel from the surface to the depths of the BS-EC facies. Therefore, it is hardly reasonable to consider a several million years long geologic event always under equilibrium conditions. Therefore, the origin of the serpentine mélange belt of the Central Franciscan Complex looks more similar to that of olistostrome or submarine land-slide (Hsu, 1967).

Another model for the exhumation of HP rocks, which sensu stricto is a model of orogeny by wedge extrusion, was proposed by (Maruyama, 1990, 1997; Maruyama et al., 1996) (Fig. 20). The subducted sediments are metamorphosed at mantle depths, but they cannot rise vertically to penetrate the hanging wall of mantle peridotite because of rigidity. The regional metamorphic belt must exhume as a thin tectonic slab along the Benioff plane, i.e. by the mechanism of wedge extrusion due to the decreasing angle of subduction (Fig. 20a). Exhumation is not caused by buoyancy, but by the stress from approaching mid-oceanic ridge, which decreases the angle of subduction. The tectonic exhumation of the thin metamorphic slab stops at the depth of the brittle/ductile transition, i.e. ca. 10-15 km. At that depth, extensive hydration and recrystallization results in retrograde metamorphic transformation of the prograde mineralogy, in particular, along the margins of the exhuming slab. The subduction-exhumation path takes about 20-30 m.y. The orogenic doming (mountain building) is provided by a later underplated accretionary complex (Fig. 20b). High-angle normal faults crosscut the sub-horizontal paired faults on the top and bottom of the high-P/T regional metamorphic belt (Fig. 20c).

2.8 Summary on regional metamorphic belts in orogens

The discovery of UHP rocks revolutionary changed the concept of orogeny. The idea of progressive metamorphic isograds provided by the overlain rocks collapsed by the discovery of UHP-HP rocks in both Pacific-type and collision-type orogens, although in the case of collision-type the UHP-HP metamorphism appeared stronger (see Section 2.5 above).

The concept of regional metamorphism is based on the following three basic assumptions. (1) The progressive metamorphic recrystallization remains in exposed regional metamorphic rocks, as if knife-cut across the units subducted to 10-30 km depths; the exposed on rocks never get hydrated, simply because the progressive metamorphism occurs through dehydration reactions, i.e. the rocks once dehydrated cannot experience retrograde metamorphism through secondary hydration during their return to the surface. (2) The bottom must be continuous toward the underlying mantle because P-T increase continuously with depth. (3) The return path to the surface records isostatic rebound, hence huge sedimentary basin must be formed nearby the metamorphosed unit.

1) The first assumption collapsed by the discovery of UHP-HP rocks common in most metamorphic belts, in particular, in C-type orogens over the world. The UHP-HP rocks preserve in the core of extensively retrograded rocks formed through hydration reactions during the return path to the surface. This has been clearly demonstrated by geochronological evidence from the core (protolith), mantle (UHP stage), and rim (Barrowian stage). The age difference between the mantle and the core is ca. 20-30 Ma. The index minerals of UHP-HP metamorphism within zircon are completely protected from the infiltration of fluids during the way to the surface. For example, eclogites are typically surrounded by retrogressively hydrated rocks (from core to rim): coesite-bearing eclogite, garnet amphibolite, amphibolite (plagioclase-bearing), and greenschist facies.

2) UHP-HP rocks occur as a core of a thin slab intruded into unmetamorphosed units. The bottom is cut by a reverse fault, not continuous to the mantle. This strongly refuses the classic concept of

orogeny, i.e., a 30 km thick continuous isostatic rebound of orogen to expose the bottom of continental crust on the surface. This is also consistent with the observed absence of huge sedimentary basins associated with orogenic belts over the world, originally pointed out by de Rover (1957).

3) The progressive stage of isograds and reactions has been nearly lost except for small domains shielded from hydration reactions by, e.g., zircons. The classic concept of facies series shown in the P-T space by Miyashiro (1961) must be improved from clock-wise to anti-clockwise for the progressive stage, as has been demonstrated in many examples, and must be consistent with the numerical simulation of P-T change along the Benioff plane (Peacock, 2001).

In summary we highlight the following issues.

(1) Regional metamorphic belt always represents a very thin flat high temperature intrusion.

(2) The upper and lower boundaries of the HP belt are subparallel paired faults separating it from the non-metamorphosed units. The deformation fabrics of the boundaries suggest that the top fault is normal and the bottom fault is reverse.

(3) The internal structure of a regional metamorphic belt is symmetric. The thermobaric structure shows the highest pressures and temperatures in the structural middle (core).

(4) The upper and lower geologic units are non-metamorphosed or weekly metamorphosed suggesting a large pressure and temperature gap the two faults.

(5) The mineralogy of the lower unit may indicate juxtaposed mid-pressure-temperature metamorphism as in the Kokchetav massif (Terabayashi et al., 2005) as a result of contact metamorphism during the high-T intrusion of metamorphic rocks in the andalusite-sillimanite facies of metamorphism. The sandwiched structure of the Kokchetav C-type UHP-HP metamorphic belt also includes those three units, which are cut by high-angle secondary normal faults forming a related dome structure.

(6) The regional metamorphic belt may record an anti-clockwise PT time path because of retrograde hydration and recrystallization from HP to intermediate pressure types. If we do not know the point of earliest UHP-HP progressive metamorphism, the Barrowian metamorphism shows a progressive clock-wise rotation P-T-time change as recorded in zone garnets. If we use a P-T-time path to reconstruct the progressive metamorphism of UHP-HP rocks, the result will be opposite (Masago et al., 2010).

(7) The chronology of whole history of regional metamorphism by using SHRIMP U-Pb ages of zoned zircon crystals with tiny mineral inclusions from core through mantle to the rim definitely indicates the age of progressive metamorphism followed by the Barrowian stage hydration. The time span takes 20-30 million years, first very quick subduction within a few m.y., followed by ultra-slow exhumation up to the mid-crustal 10-15 km depth, to stay there 20-30 m.y. The total time span is about 30 Ma, followed by rather long mountain building, e.g., the Himalaya and Alpine mountains have been growing since the Miocene.

(8) By an unknown reason the isostatic rebound by driven buoyancy has long been considered a driving force of the exhumation of regional metamorphic belt, i.e. de-facto of orogeny. The problems of such a model were explained in the previous sections. The mode of occurrence of metamorphic belts (thin HP plate sandwiched between LP rocks), thermobaric structure, progressive metamorphism, anti-clockwise P-T-time path, and geochronological constraints combined with the timing of exhumation all are well-correlated with the ridge subduction stage in the case of Pacific-type orogeny and slab break-off in the case of collision-type orogeny. Moreover, in spite of significant mountain building typical of both Pacific-type orogens elevated to 3000 meters and collision-type orogens elevated to 5000-8000 meters, the associated sedimentary basin never shows the thicknesses of associated basins as large as those of the crust (30 to 50 km). All these constraints strongly indicate that metamorphic rocks must be tectonically extruded to the surface from mantle depths as a thin flat unit along the two parallel faults. The two-step exhumation model (Maruyama et al., 1996) implies that such a sandwiched unit starts rising from a depth of ca. 100 km, the highly ductile core would extrude faster and farer upward, than the surrounding lower-T rocks. At the depth of the ductile-brittle transition boundary, the UHP-HP core stops by the increased viscosity at a temperature of ca. 350℃ by the starting retrograde reactions (Fig. 20). Therefore, the exhumation is driven not by buoyancy, but by wedge squeezing out or wedge extrusion provided either by a shallowing subduction angle or/and by an approaching mid-oceanic ridge in the case of Pacific-type orogeny, or by slab breakoff in the case of collision-type orogeny (Maruyama et al., 1996).

Chapter 3. Petrogenesis of continental crust

Pacific-type orogeny plays an extremely important role in the increase of the volume of the continental crust, which, unlike the mafic crust, is a source of nutrients critical for the origin and evolution of life. Collision-type orogeny cannot produce new continental crust. This makes a major striking difference between the two types of orogeny. In that sense, the petrogenesis of granitic or andesitic rocks is of special importance. In this paper we often use term "TTG" while describing igneous rocks of approximately andesitic composition. TTG mean a tonalite-trondjemite-granodiorite assemblage formed over subduction zones. The term originally was used for Archean rocks only, but then became common in discussing younger Pacific-type belt. So, TTG rocks in general indicate a bulk composition of andesite erupted on the surface along subduction zones. On the other hand, TTG magma can form intrusions in the middle crust. In this chapter we will first summarize the genesis of island arc magma and then the origin of granites.

3.1 Origin of island-arc magma

In Japan, the petrogenesis of island-arc magma has been investigated since the first synthesis by H. Kuno (1959) who studied volcanology in the field and introduced basics of igneous petrology. The experimental works on magma genesis were started by N.L. Bowen as early as in the 1920-ties, followed up by D.S. Yoder in the 1960-ties and later by I. Kushiro in the viewpoint of plate tectonics. In the time when N. Bowen started working it in the Geophysical Laboratory, Carnegie Institution, Washington D.C., the world knowledge focused on the origin of diverse volcanic rocks. Bowen and his co-workers established basic phase diagrams in the basalt-andesite-dacite system, and concluded about the presence of a single primary magma of basaltic composition derived from the partial melting of peridotite. That melting forms the whole variety of magmas through fractional crystallization. The next generation of petrologists used that viewpoint speculating about fractional crystallization and how many primary magmas may exist. The concept of Plate tectonics distinguished three compositionally different primary basaltic magmas: 1) MORB, 2) arc basalt, and 3) OIB (Hawaii and Iceland).

Kushiro (1972) used the observations of H. Kuno on Japan volcanology to understand the origin of island-arc andesite and basalt. A. Ringwood (1974) considered the derivation of island-arc magmas by partial melting of MORB, which form at mid-oceanic ridge and later get subducted to be partly melted, because arc magmas are produced at lower temperatures compared to the mantle melting under mid-oceanic ridges. Later, Kushiro (1974) doubted that simple speculation and proposed an alternative of high-Al basalt and/or boninite-high-Mg andesite as subduction zone primary magmas. Tatsumi and co-authors (1983) proposed a unique model based on the global compilation of subduction zone magmas (Fig. 21). The model includes five major postulates. (1) The temperature of the source mantle is about 1400℃.

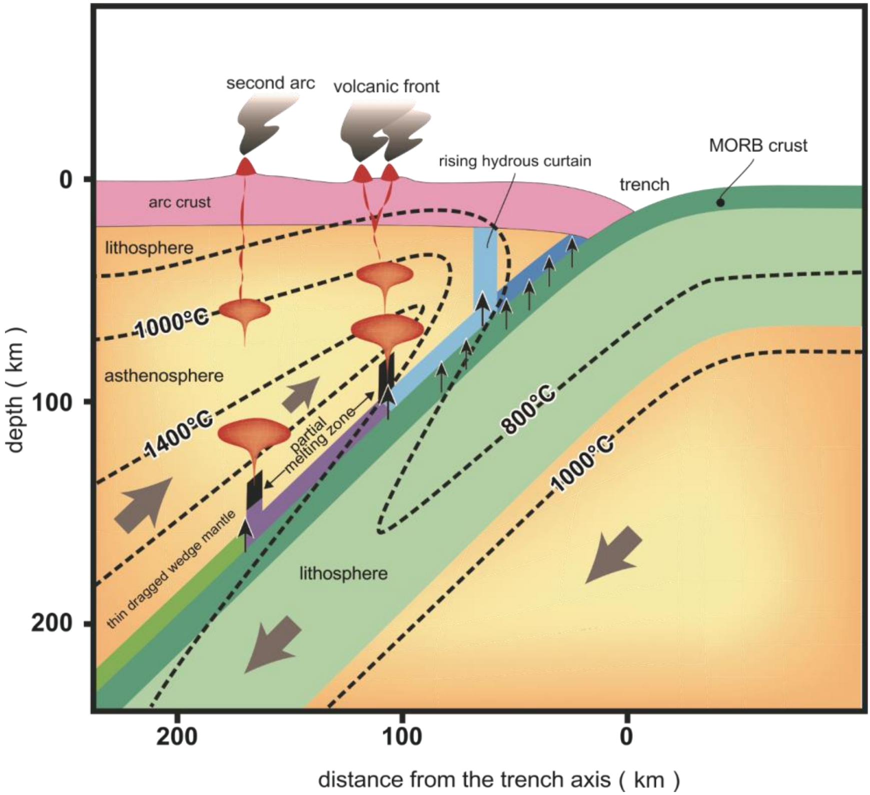


Fig. 21. A petrologic model of arc volcanism in NE Japan arc after Tatsumi et al. (1983). The descending, hydrous Pacific slab dehydrates at depths of 50-60 km to release aqueous fluids that hydrate the thin convecting wedge mantle, which subducts together with the Pacific slab. The dehydration occurs at two depths: 100 km and 180 km, which, respectively, correspond to the breakdown of hornblende beneath the volcanic front and breakdown of phlogopite beneath the second volcanic front. Note the presence of two curtain-like vertical columns of fluid-magma flow and the presence of ca. 1400℃ high-T mantle under the volcanic front, which is 50℃ higher than that of MOR.

(2) The source arc magma is nearly dry; water content of magma is one order of magnitude lower than arc primary magma, believed before. (3) At a depth of ca. 100 km, a curtain-like hydrous column forms over the subducting slab under the first volcanic front; the melting in triggered by the reactions of amphibole breakdown. (4) At a depth of ca. 200 km, another curtain-like hydrous column forms by the breakdown of phlogopite. The second hydrous column ascends to form a second volcanic front. Note that the first volcanic front produces more than 90% volumes of magma, but subordinate amounts of magma can be also generated by the second volcanic front. (5) To maintain such a system of material and energy circulation, continuous supply of fertile mantle underneath the volcanic front is necessary. A lot of new data, first of all from seismic tomography, have been obtained though since the appearance of Tatsumi’s model more than 30 years ago (Tatsumi et al., 1983). Now it is the time to improve it.

The data on seismic tomography (Zhao et al., 1994) clarified the velocity (P-wave) structure of the mantle wedge underneath the Tohoku arc in NE Japan down to a depth of 200 km. The seismic tomography images show low-velocity regions of magmatism beneath the volcanoes on the surface. Their research showed the following three important facts, which help to understand the petrogenesis of arc magma. Seismical data combined with MORB+water and peridotite+water phase diagrams make possible estimation of the thermobaric structure under NE Japan. If we adopt the water effect on the peridotite+water solidus, the H2O content of 4-5 wt.% in volcanic front primary magmas measured by glass inclusions in olivine (Ohta et al., 2007) indicates a temperature of ca. 1200℃ in the source mantle (Iwamori, 1998). This value is 200K lower than that of Tatsumi et al. (1983) who assumed nearly dry primary magma (0.2 wt.% H2O). The tomographic images show no two hydrous peridotite columns rising from the depths of 110 km and 200 km directly above the descending Pacific (Tatsumi et al., 1983). Instead, an obliquely upwelling curtain flow is observed at a point directly above the convergence of double-seismic zones at a depth of 230 km. The enlarged figure (modified after Hasegawa et al., 2009, 2010) shows trench-fill turbidite at depths of 40-80 km subducted together with the descending Pacific slab and already attached metamorphosed trench sediments on the hanging wall including serpentinized bodies traditionally called Alpine-type peridotites. A small triangular region named metasomatic-metamorphic factory (MMF) represents a mechanically separated material flow independent of the subduction zone magma factory (SZMF). MMF accepts the abundant fluids derived from the dehydrated descending Pacific slab to cause serpentinization and its solid intrusion into the attached metamorphic unit. The sources of water in the Pacific plate are oceanic sediments and basalt/peridotites hydrated by the transform faults near mid-oceanic ridge and by the faults related to the bending of the slab near the trench before subduction.

(1) The oblique counter-flow (anti-subduction) convection in the mantle wedge driven by the descending Pacific slab underneath the volcanic front started at a depth of about 200 km. The convection seems to start at a depth of ca. 160 km and attenuates towards the volcanic front at the Moho depth of ca. 35 km (Fig. 22). The obliquely rising flow moving toward the volcanic front was first detected by seismic tomographic images (Zhao et al., 1994, 2009). It was quite different from the classic numerical simulation of mantle convection, which shows horizontal mantle flow starting from the continental side at a depth of about 50 km depth, i.e. from the eastern margin of Asia, then moving across the Sea of Japan and finally meeting the descending Pacific slab and then going down together. That numerically inferred mantle convection pattern is quite different from that seen in seismic tomographic images.

(2) Seismic tomography images do not confirm the two subparallel vertically rising water columns proposed by Tatsumi's group (Tatsumi et al., 1983). Since the Zhao's pioneering work, tomographic images have been improved (Hasegawa et al., 1994, 2009) to include the data on P-waves first done by Zhao's group (Zhao et al., 1994), S-waves and Poisson ratios sensitive to bulk compositions including water. Using those new tomographic images, the pressure-temperature-composition (water, TTG) distribution patterns in the mantle wedge and their relationships with the distribution of volcanoes were constrained. The research areas were extended from NE Japan to SW Japan, and whole Asia. Those results declined the presence of vertical hydrous columns.

(3). Oblique upwelling counter flow and generation of small individual plumes. The 3D seismic tomography images by Zhao et al. (1994, 2009) show that the convection can be traced further to a level of ca. 50 km. There, the flow splits into several smaller flows like finger. Each finger may exhibit a plume generating magma and connecting the region of convection with the volcanoes on the surface. Each volcano stands isolated to around 100 km, therefore the convecting flow finally diverges into smaller branches. Each rising plume generates magma by decompression creating a volcanic front parallel to the trench, and volcanoes stand above the line going to the 100 km depth on the Benioff plane (Tamura et al., 2002).

Thus, by combining the tomographic images by Zhao and Hasegawa groups we evaluated the classic Tatsumi model of arc magmatism

(Tatsumi et al., 1985) proposing two hydrous rising columns as shown in Fig. 21. However, the seismic high-resolution tomographic images never show those two vertical hydrous columns.

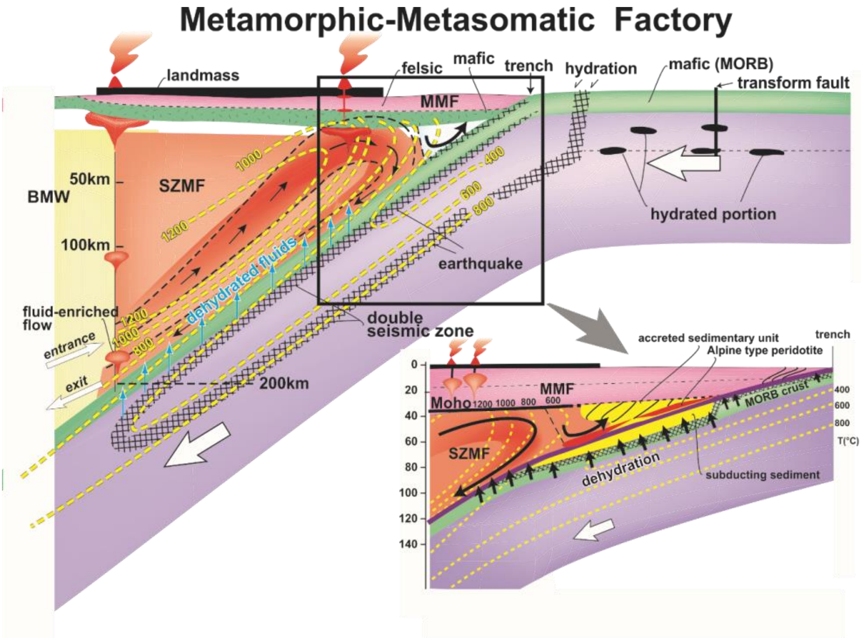


Fig. 22. The thermal structure and fluid migration pattern created by the descending Pacific slab under NE Japan modified by Omori et al. (2002, 2004, 2009) and Maruyama et al. (2009), based on the seismicity distribution patterns by (Hasegawa et al., 2009,2010).

Even at the depth of 200 km, the subducted slab can contribute relatively large amounts of water on its surface. In general, the subducting slab is heated by the surrounding mantle, from the top and from the bottom. Therefore, the center of the descending slab may remain relatively cold. The hydration of the slab originally results from the hydrothermal circulation and/or the transform faults near the mid-oceanic ridge. Then, near the trench, the bending of the slab before the subduction can form normal faults on the ocean side, along which surface water directly penetrate into the slab and hydrate slab peridotite. Those hydrated portions of peridotite also subduct and the hydrated slab gets heated by the underlying mantle and overlying mantle (Fig. 21). If so, the dehydration should occur not only on the surface of MORB crust plus overlying oceanic/trench sediments, but also deeper in the slab, where peridotite can be also dehydrated (Umino and Hasegawa, 1975; Hasegawa et al, 1987). The seismic images discovered a double seismic plane within >50 Ma old, i.e. rather thick, descending slabs. The earthquakes with epicenters deeper than 60 km are triggered by the reactions of embrittlement related to the dehydration of water-carrying peridotite (Hasegawa et al., 1994, 2009). Below we will discuss fluid flow patterns, petrological reactions and localization of dehydration, and their relation to seismicity.

(4) MMF, SZMF and BMW domains, fluid flow column and earthquakes. The fluids derived from the upper seismic plane of the double seismic zone must penetrate into mantle wedge through the already upper dry eclogitized MORB crust. However, the fluids derived from the lower seismic plane cannot penetrate into the upper seismic plane through the lower temperature peridotite in between. There the reaction of hydration is expected, because the temperature must be lower in the center than beneath the double seismic zone. With increasing depth, the lower seismic plane gradually shifts upward and finally the upper and lower seismic planes of the double seismic zone meet at a depth of 200 km. At that depth, fluids must generate a plume or several small plumes rising above the Benioff plane, although those plumes are small and cannot be detected by tomography (Fig. 21). The rising hydrous column starting at the 200 km depth points may cross the oblique convective flow at about 150 km and cause upward convection of the low velocity layer to reach the volcanic front (Fig. 22).

In a number of subduction zones, earthquakes at 50-200 km depth define two dipping planes, separated by 20-40 km, that appear to merge downdip (Fig. 22). Upper plane earthquakes are inferred to occur within the subducting oceanic crust, whereas lower plane earthquakes occur in the subducting oceanic mantle. A recent finite-element heat-transfer model for the region beneath NE Japanshows that the lower seismic plane cuts across isotherms at a shallow angle (Peacock, 2001). Lower plane earthquakes occur at ~550-800 °C at 100 km depth and at ~350-600 °C at 160 km depth. These conditions coincide with the dehydration reaction antigorite (serpentine) → forsterite + enstatite + H2O, which suggests that lower plane earthquakes may be triggered by dehydration embrittlement, which in turn suggests that the subducting oceanic mantle is partially hydrated. Serpentinization may occur in the trench-outer rise region, where faulting may promote infiltration of seawater several tens of kilometers into the oceanic lithosphere. If this hypothesis is correct, current subduction-zone H2O budgets may significantly underestimate the amount of bound H2O entering the "subduction factory.

The hydrous column, which forms at the depth of 200 km, i.e. at the meeting point of two seismic layers, can be very short, probably only 50 km high. This hydrous column may represent a bridge between the zone of supra-subduction magmatism and big mantle wedge (Zhao et al., 2007). If we look at the tomographic images of the upper mantle beneath the area between Beijing, China and the Japan trench about 2000 km apart (Fig. 22, 23), we can see three tectonic domains: (1) “Metasomatic-Metamorphic Factory (MMF)”, a small triangular zone of the mantle wedge having an independent convection system over the subducting slab to promote the exhumation of UHP-HP metamorphic rocks; the MMF triangle zone extends to the depth of 50-60 km at the Benioff plane; (2) “Subduction Zone Magma Factory (SZMF)”, located between MMF and the 200 km depth at the Benioff plane, which is a merging point of double-seismic zone planes. (3) “Big Mantle Wedge” – BMW - is the rest part of the mantle wedge, defined by the bottom of the stagnant slab at the mantle transition zone (Zhao et al., 2007).

The dynamics of MMF is based on tomographic studies (Hasegawa et al., 2009; Fig. 22) showing an E–W cross-section along NE Japan. The regionally metamorphosed sedimentary units are located immediately on top of the subducting Pacific plate and older regionally metamorphosed units are accreted above. The former is differentiated by P-wave, S-wave and Poisson ratio. Particularly 3.5–4.5% slow Vs km/s is critical. Two separate Alpine-type peridotite horizons are shown sandwiched by low velocity anomaly regions. Most active seismicity is present not only in the upper half of subducting Pacific but also right above the corner of mantle wedge. MMF is separated from SZMF on the left. High temperature mantle portion heats up subducting Pacific from above after the release of arc magma and turn toward deep mantle on the top of subducting plate (Kogiso et al., 2009). Above the boundary, anti-clockwise convection cell could be present to remove the regional metamorphic unit upward bringing together pieces of hydrated mantle wedge. The past examples of wrapped peridotites in the Pacific-type orogenic belts can be compared to infer the mechanism of emplacement. For example, Cretaceous Sanbagawa high-P/T belt in SW Japan includes Higashi–Akaishi peridotite (Kunugiza et al., 1986; Banno, 2004) and Horoman peridotite in central Hokkaido (Niida, 1984). Both have been called as an Alpine-type peridotite which could have been derived from mantle wedge. Top portion of Pacific slab could be metamorphosed at shallow depths from blueschist through lawsonite eclogite and eclogite facies down to 60 km depth. Below 60 km depth, MORB crust can be heated up effectively by the counter flow of mantle convection in the magma factory (Omori et al., 2009). The right bottom inset diagram in

Fig. 4 shows the frequency of seismicity in the MORB crust. The highest peak around 60–65 km depth suggests the dehydration from blueschist to amphibole eclogite (Fig. 22).

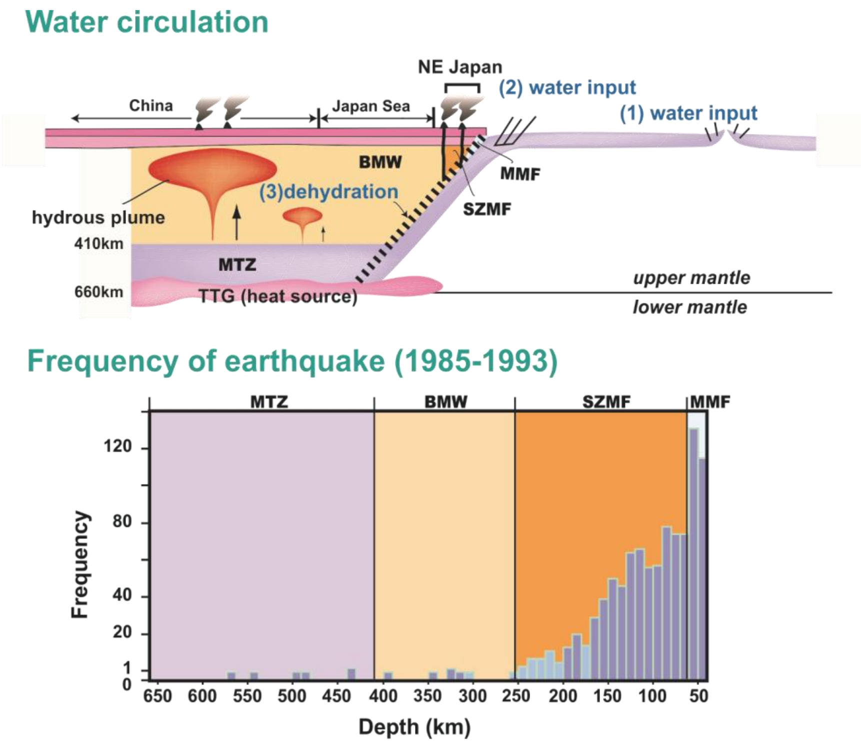


Fig. 23. Upper mantle from the Japan trench through the Sea of Japan to China. Under the NE Japan arc, the upper mantle is subdivided into MMF, SZMF, and big mantle wedge (BMW). BMW covers a wide range of East Asia, and suffered from hydrous plumes originating at a depth of 410 km caused by the TTG self-heating system underneath the hydrous mantle transition zone. Water in the mantle transition zone was supplied by the descending Pacific slab. The Pacific plate was hydrated along the transform faults near the mid-oceanic ridge and again at off trench by the bending slab along the normal faults and dehydrated during subduction caused by heating from the surrounding mantle. The lower figure shows the frequency distribution of seismicity versus depth. The most extensive dehydration occurs in the MMF, followed by the SZMF. Rare occurrence of seismicity in BMW and mantle transition zone (MTZ) is shown (Maruyama et al., 2010).

The SZMF is best characterized by the presence of double seismic planes and the lower seismic plane starts immediately below the trench axis. Large earthquakes by the bending of Pacific slab may cut the whole slab, or at least crack the middle depth leading to hydration even in the central portion of the slab. This causes the lower seismic plane earthquakes. The top portion of the Pacific slab is hydrated at mid-oceanic ridge with the hydration triggered by trench outer-wall normal faults by bending Pacific slab (Fig. 23). The convergence of the double seismic plane leads to the generation of large-scale hydrous plume which moves upward and define the boundary of the SZMF with the BMW.

Frequent occurrence of seismicity below the magma factory would supply fluid underneath leading to a decrease in the viscosity of the mantle wedge. As a result, the magma factory domain is characterized by lower viscosity and lower melting points to generate magma. Obliquely oriented rising mantle flow towards the volcanic front has been well defined by seismic tomography and supplies fertile high temperature mantle underneath the volcanic front. Twenty to 30% melting of those would generate arc magma of andesitic to basaltic andesite composition, at presumably temperatures of 1250–1300 °C at 10–15 kb with 0.3–0.5 wt.% H2O (Kogiso et al., 2009). However, if the water content of primary magma is higher than the value mentioned above 0.9–1.5 wt.% (30% partial melting), then melting temperature must be lower than 1250–1300 °C. Water content of melt inclusions in primitive olivine shows a range of 3–4 wt.% in the case of Mt Fuji volcano (Sato et al., 1992) or even more (4–5 wt.%) with high sulfur contents at several volcanoes standing on the volcanic front (Yamaguchi et al., 2006). If so, the melting temperature would be much lower ca. 1100–1200 °C or even down to 1050°C (Hirose and Kawamoto, 1995). After the extraction of magma, the mantle flow turns around and moves to the deep mantle along the subducting Pacific slab. The entrance of the fertile magma from BMW could occur at a depth of 160 km, and the exit of the depleted mantle occurs below at round 200 km depth. The magma factory is thus driven by extensive fluid infiltration by the underlying subducting oceanic slab. Zhao et al. (2004, 2007) proposed a big mantle wedge (BMW) model to emphasize the role of the stagnant Pacific slab and the BMW above the slab in the formation of the intraplate volcanism and mantle dynamics in East Asia. Similar to the regular (small) mantle wedge under island arcs, complex dynamic processes might have taken place in the BMW, such as corner flow and deep slab dehydration, which may cause upwelling of hot asthenospheric materials and lead to lithospheric fractures and thinning under East Asia (Menzies et al., 2007). The extensional rift systems and faults widely existing in East Asia (Tatsumi et al., 1990) may be the surface manifestation of these dynamic processes in the BMW and the mantle transition zone. Recent petrologic and geochemical studies also support this BMW model (e.g., Chen et al., 2007; Zou et al., 2008).

Stagnant slabs occur underneath Japan to East Asia for over 2000 km resulting from the eastward subduction of the Cretaceous to Jurassic Pacific plate from the east. The region immediately above the stagnant slab is termed as big mantle wedge (Zhao, 2004; Zhao et al., 2004, 2007). The bottom depth of BMW is estimated to be 410 km at which depth hydrous wadsleyite (beta phase olivine) transforms to olivine+fluid (e.g., Ando et al., 2006). Below this zone, dense hydrous silicates of wadsleyite and hydrous ringwoodite (gamma phase olivine) are stable. This region is a huge water tank in that almost five times the volume of water in the present day oceans on the Earth's surface is estimated to be stored in the dense hydrous silicates in this zone (e.g., Murakami et al., 2002; Maruyama and Liou, 2005). Immediately above the 660 km boundary, granitic crust of probable TTG composition is present which functions as a self-heating source to generate hydrous plume above the mantle transition zone as will be proposed here-in afterward. Several hydrous plumes are confirmed by detailed P-wave tomographic images with apparent connection to hot spots and volcanoes (see tomographic images in Zhao, 2004 and Maruyama et al., 2009). The top right part of the BMW is the magma factory (Tatsumi, 1989) and is the zone of generation of arc magma.

The temperature increases from MMF through SZMF to BMW. The dominant process in the MMF is the mechanical driving force of plate subduction, and the hydration of the mantle wedge by fluids derived from dehydration with incorporation of large ion lithophile elements (LILE) such a K, Rb, Na as well as other incompatible elements. The extensive recrystallization of trench sediments dragged down leads to regional metamorphism and these are exhumed back to the surface if adequate conditions prevail. The magma factory is driven by hydrous plumes generated at 200 km depth through the convergence of double seismic zones. The 200 km depth marks the boundary between the magma factory and BMW. The SZMF is bound by BMW by this plume, although several hydrous plumes of much smaller scale can be generated at all depths along the Benioff plane in the SZMF factory. A critical hydrous column defines the boundary between BMW and SZMF. The boundary is mechanically controlled by the largest dehydration reaction derived from the convergence of double-seismic planes at 200 km depth. The magma factory is hydrous and viscous due to the presence of interstitial fluids among minerals and this factory can generate arc magma underneath the volcanic front. To compensate the convective flow by plate subduction, more rigid material from BMW is incorporated near the bottom of the boundary between BMW and SZMF. On the other hand, the high-temperature residual mass after the release of arc magma is dragged by the down flow immediately above subducting Pacific slab. To maintain constant volcanism, material input from MBW to SZMF is crucial.

BMW occupies the largest portion, probably more than 90%, in the western Pacific upper mantle. However, magma production is minor in this zone. The driving force of this domain is minor amounts of dehydrated fluid and probably heat generated from the bottom of the mantle boundary layer (mantle transition zone).

(5) Water content. Another remarkable aspect of our model is the absence of hydrous minerals in the mantle wedge just above the subducting slab at depths deeper than 110 km. Many previous models have proposed that hydrous minerals are stable in the bottom layer of the mantle wedge if subducting slab is as cold as the Pacific slab beneath NE Japan and that the layer is dragged into deep mantle by subducting slab (Tatsumi, 1989; Schmidt and Poli, 1998; Iwamori, 2004). These previous models suggested that hydrous minerals in the dragged layer are stable beneath volcanic front and deeper suggesting relatively dry magmas derived therein. In another model (Kogiso et al., 2009), however, hydrous minerals are stable only at shallow (<120 km) depths in the mantle wedge corner. Therefore, the thin low-velocity layer just above the slab around 70–90 km depths found by the double-difference tomography (Tsuji et al., 2008; Nakajima et al., 2009) corresponds to the hydrous mineral stability area (Fig. 22; Kogiso et al., 2009). However, the low-velocity layer found at the deeper portion (Kawakatsu and Watada, 2007; Tonegawa et al., 2008) does not indicate the presence of hydrous minerals there.

The deep low-velocity layer, which was found by the receiver function technique (Kawakatsu and Watada, 2007; Tonegawa et al., 2008), is not necessarily a thin “layer”, because the thickness of the low-velocity “layer” cannot be constrained solely by the receiver functions. Instead, the receiver function technique is sensitive to a sharp velocity change within a short length scale. The model of Kogiso et al. (2009) predicts that partial melting in the mantle wedge occurs just above the subducting slab and in the slab at deeper than $\sim$120 km. Below this partial melting zone, hydrous minerals and/or free aqueous fluids are present, and the boundary of these zones are nearly parallel to the slab surface (Fig. 24).

Although amounts of aqueous fluids and hydrous minerals in the slab are difficult to constrain, it is clear that the fraction of “liquid” (silicate melt or aqueous fluid) in a certain volume of peridotite/basalt abruptly increases around the solidus with increasing temperature in a

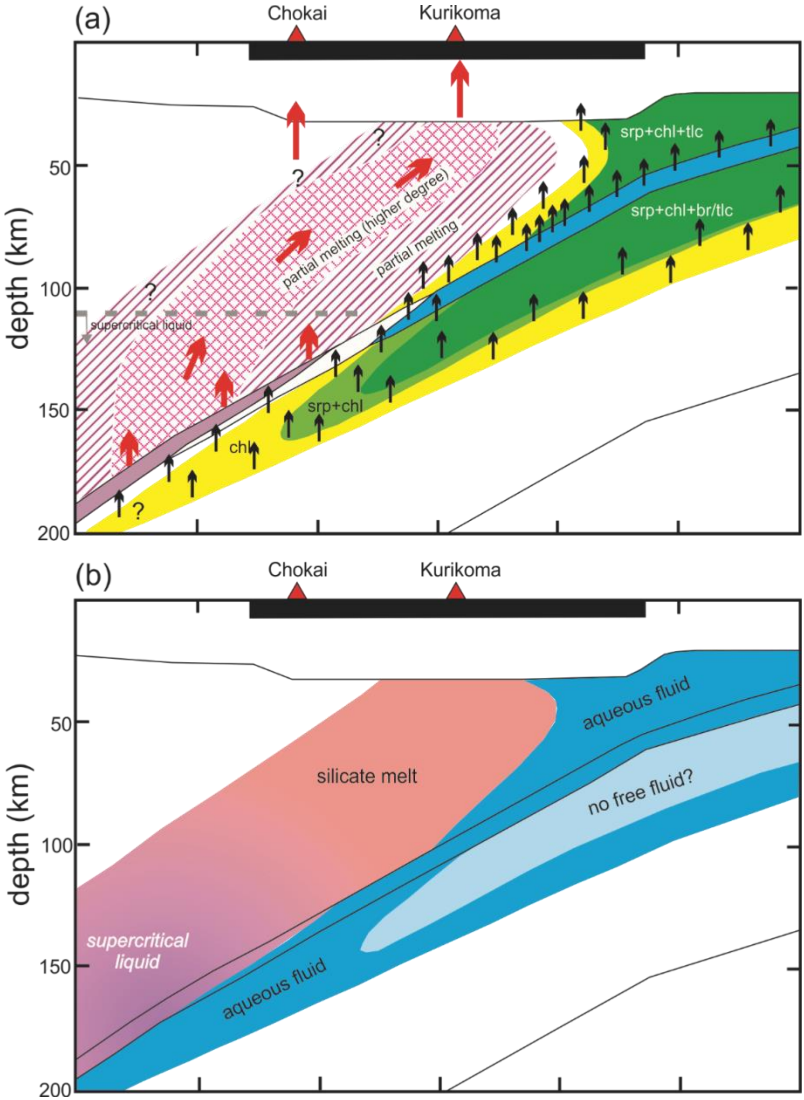


Fig. 24. (a) Stability fields of hydrous minerals and partial melt estimated from the temperature distribution (Fig. 22). Black and dark red arrows schematically show movement of aqueous fluids and silicate melts, respectively. A thick broken line shows the boundary of critical curve below which aqueous fluids and silicate melts exists as supercritical liquid. Abbreviations are: AmEc = amphibole eclogite, EpEc = epidote eclogite, BS = blueschist. (b) Estimated possible distribution of free aqueous fluid, silicate melt, and supercritical liquid.

hydrous peridotite and basalt systems (Fig. 25). Therefore, the fraction of partial melts in the partial melting zone is much higher than the fraction of aqueous fluids in the slab, at least along fluid migration paths. Thus, the bottom of the partial melting zone can be a sharp velocity-jump boundary (Fig. 25), which can be detected by the receiver function technique.

The high temperature of ca. 1,400℃ was estimated by (Tatsumi et al., 1985) for the primary magma of the mantle due to a very low water content of 0.2 wt %. This estimate came from the observation of phenocryst assemblages in andesite, dacite and rhyolites. Phenocryst assemblages appeared different across the arc reflecting different water contents in magma increasing to oceanward from volcanic front. Conversely, the water content decreases to the volcanic front, hence leading the water down to 0.2 wt % in the primary magma. Using the amount of 0.2 wt.% H2O, a hypothetical tholeiitic magma was calculated as a primary magma with olivine, orthopyroxene, clinopyroxene and spinel in equilibrium with melt to give a laboratory-based estimate of 1400℃ (Tatsumi et al., 1985). Therefore, the miss-understanding came from the water content estimate in the topmost magma reservoir, several kilometers deeper from the surface, where the magma is already phenocryst-rich and degassed, far away from the primary magma generated in the deep mantle (Fig. 25).

Much higher 4-5 wt.% water contents in the primitive melt appeared by studying melt inclusions in olivine (e.g., Oto et al, 2007). Those discoveries suggest a temperature of ca. 1100℃ in the source mantle, i.e. 300K lower than that of the previous estimate, using the H2O-dependence solidus curve (Iwamori, 1998). We do not know exact H2O content in the primary magma formed in the mantle wedge. The magma rises to the first magma reservoir at around the Moho depth, which is a density boundary between the mantle and the lower crust. Hence, the primary magma should stop at this depth to fractionate and lose a part of water. A liquidus phase must be olivine. Such a magma must contain at least 4-5 wt.% H2O coexisting with olivine. Then, the residual melt moves upward to stop at the second reservoir, at the boundary between the upper and lower crust, i.e. at the 15 km depth, where it forms plutons through fractional crystallization. A part of that magma would continue rising up to erupt on the surface. The third reservoir is underneath a volcano. At those several steps of ascent the magma loses volatiles. Therefore, the water content in the primary magma should be much higher (i.e., in the mantle wedge) and that in the third magma must be very low (Fig. 25).

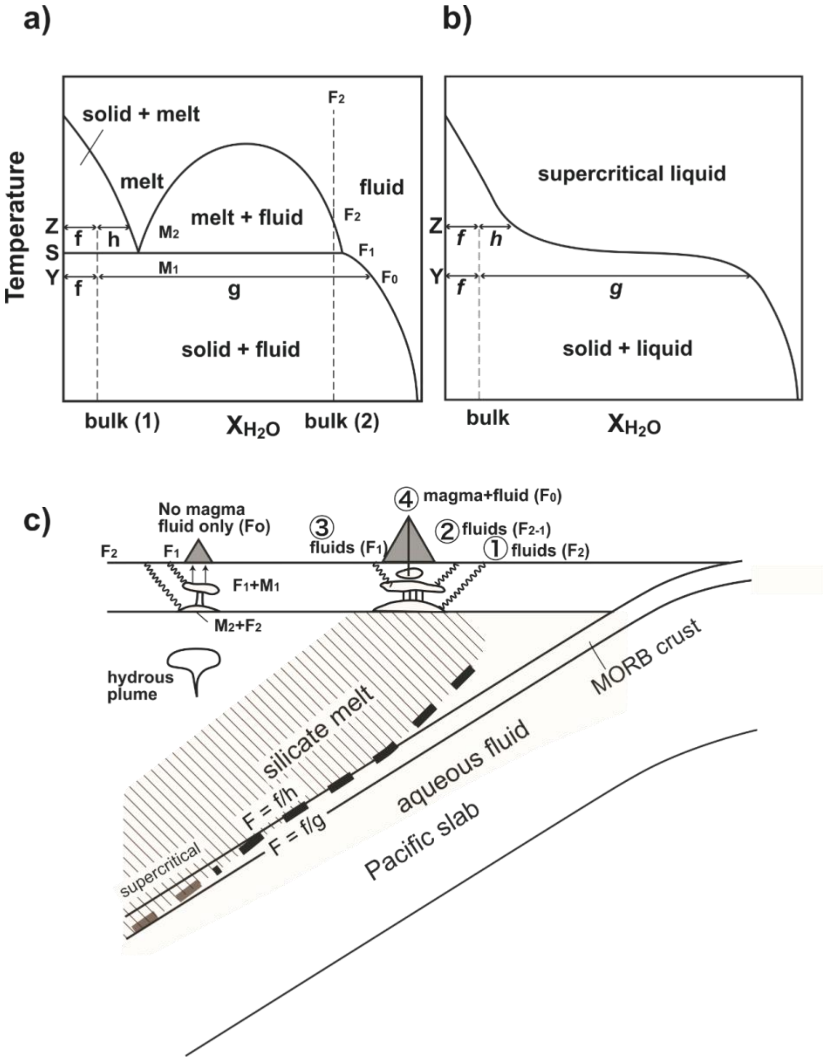


Fig. 25. (a) A schematic diagram showing phase relations in the binary silicate melt—H2O system. (b) Similar diagram in higher-pressure conditions. (c) Generation of arc magma, fluid migration, and compositional fractionation under NE Japan. A rising hydrous calc-alkaline (CA) magma generated in the wedge mantle stops at three different depths during its ascent: first at the Moho (30—35 km), second at a mid-crustal depth (15—10 km), and finally at a depth of several kilometers under the active volcano. The aqueous fluids released at three different depths, with the content of H2O originally higher than 5 wt.% in the mantle wedge, caused a drop of the liquidus line to promote crystallization of porphyritic magma in a supra-subduction setting (modified after Kogiso et al., 2009).

3.2 Petrogenesis of granitic magma

One possible mechanism is hydrous partial melting of oceanic basaltic crust in the subducting slab (e.g., Nicholls and Ringwood, 1972; Kay, 1980; Wyllie and Sekine, 1982). This mechanism had been proposed as the main cause for basaltic–andesitic magmas that dominate subduction zones. However, later melting experiments on hydrous basaltic systems (e.g., Beard and Lofgren, 1991; Rushmer, 1991; Wolf and Wyllie, 1994; Rapp and Watson, 1995) demonstrated that hydrous melting of basaltic crust dominantly produces silicic magmas with particular chemical characteristics like tonalite or adakite, which are limited in volcanic arcs beneath which relatively young (=hot) oceanic lithosphere is subducting. In addition, analytical and numerical models on the thermal structure of subduction zones suggested that subducting oceanic crust older than 20 Ma are unlikely to partially melt beneath volcanic arcs (e.g., Defant, and Drummond, 1990; Molnar and England, 1995; Bourdon et al., 2002). On the other hand, high geotherms of some arcs petrologically estimated from metamorphic rocks suggest that subducting oceanic crust as old as 80 Myr can partially melt beneath volcanic arcs (Kelemen et al., 2003). Partial melting of subducting oceanic sediments has also been proposed from basalt geochemistry for old subduction zones such as Izu-Mariana (e.g., Ishizuka et al., 2003).

Compared to the petrogenesis of arc basalt to andesite magma, the study of granitoids was more limited, probably because too simple mineralogy (quartz + feldspar) suggesting few petrological constraints. For example, the number of rock-forming minerals in the case of subsolidus phase relations for the regional metamorphism of basaltic protoliths is one order larger than in granites. As a result, electron probe micro-analysis, when invented and introduced to the geological community, was mostly applied to experimental petrology in general and petrogenesis of basalts and peridotites in particular.

Another problem was the texture of granites. Compared to volcanic rocks granite is completely coarse-grained. The most advanced progress for the petrogenesis of granite came from field-based and composition-based classifications. Ishihara (1978) classified granites into magnetite and ilmenite series, whereas Chapel and White (1974) introduced I-type and S-type granites. I-type corresponds to andesitic magma typically crystallized at mid-crustal levels, and S-type represents re-melted sandstone compositionally similar to the granitic crust. Except for S-type granites, the bulk compositions of granites are plotted onto several diagrams, such as AFM and major oxides contents versus fractionation index to argue for this or that magmatic rock series. These series assume that the whole-rock composition of granite corresponds to the parental liquid (Aramaki et al., 1970). Nowadays, the most well-known classification shows I, S, M, and A types of granites.

Bulk chemical composition of TTG. The average chemical composition of all crustal granitoids is andesitic. Therefore, their petrogenesis should be similar to that of arc magma. Many petrologists regard arc primary magma as basaltic though. If primary magma for granite is basalt, then their derived plutons, which are typically zoned, must consist of more than 50% of gabbro. However, most plutons do not carry gabbro or other kinds of cumulates such as peridotite at their bottoms. This strongly suggests that granitoids cannot be derived from basaltic primary melts.

The next idea came from more silicic primary magmas for arc andesite considering the origin of boninite to high-Mg andesite through the partial melting of hydrous peridotites. The occurrence of those magmas is restricted in time and space and is limited in volume suggesting unusual conditions, i.e., high temperature and higher amounts of water, i.e. continuous water supply parallel with continuous high-temperature melting of peridotite to maintain magmatism. Although boninite or high-magnesium andesite form in the mantle wedge, it is still not possible to generate huge amounts of granitic plutons from boninite or high-Mg andesite magmas. In addition, their bulk and trace element chemistry is different from crustal TTG-type granitoids.

Delamination of mafic lower crust and adakites. To overcome this problem, Kushiro (1990) and Tatsumi (2000) reconsidered their model to include amphibolite melting at the Moho depth to produce basalt in the underlying wedge mantle. The lowermost mafic crust typically consists of amphibolite, but at high temperatures and pressures and high FeO content granulite (garnet-plagioclase-pargasite-quartz-magnetite) may appear common as well. The bulk chemical composition corresponds to arc basalt. When this rock is partially melted, granitic magma can be formed under 20-30% meting remaining 70-80% restite of SiO_2-depleted ultramafic rock by the following simplified reaction: 1.0 high-pressure granulite (SiO_2 = 50 wt%) is equal to 0.3 granite melt (SiO_2 = 70 wt.%) + 0.7 restite (SiO_2 = 40 wt.%) under water-saturated condition. The generated felsic melt intrudes into the overlying upper crust and get settled at a density neutral point to be cooled down to form sills or dikes. On the other hand, the restite must be delaminated by an unknown reason, otherwise we cannot explain the granite-dominated continental crust. The surface of an active orogenic belt, i.e., mountain belt, is permanently eroded to remove the material into ocean. Therefore, the crust through time gets thinner. But the new granitic crust makes the crustal thickness balanced. The eroded material goes to the trench and then gets subducted by slab. The slab carries the sediments to form a new accretionary complex underneath of that semi-subducted material inevitably moving directly to the mantle. If primary magma is basalt, which is underplated at the Moho depth, then the mafic lower crust must get thicker through time, even if it is compositionally fractionate to produce more ultramafic restite. Can we see such mafic-ultramafic restite widely exposed in the orogenic core on the surface or in the lower mafic crust anywhere in the Earth? No. This fact definitely contradicts the observation making a new model necessary.

Kushiro (1981) and Tatsumi (2000) proposed that in the lower mafic crust conditions original basalt turns into amphibolite, which partial melting results in the formation of more mafic restite, similar to peridotite. The latter must be tectonically eroded away into deeper mantle through delamination. However, two important issues remains: (1) density and (2) size.

For example, the density of the residual lower mafic crust must be higher than that of the mantle to move down. The average density of the mantle immediately below the Moho seismic discontinuity (i.e. the base of the crust) is ca. 3.35 g/cm3, which is consistent with a mineralogy dominated by olivine, pyroxene, and a minor aluminous phase. Therefore to make such peridotite sinking down into the mantle its density must be at least 3.5 g/cm3, but it is usually 3.2 to 3.5 depending on FeO. The only way to increase its density is to increase the amount of magnetite (Fe_3O_4) but then a high oxygen fugacity is necessary, which may increase to some extent, hardly above 5 g/cm3 (Nakajima and Arima, 1998).

The delamination of continental crust seems to be impossible also because of too small size/amount. If we, following Tastumi and Kushiro, accept that the delamination occurs only under volcanic front, the amount of delaminated material must be extremely small (less than 1 km thick), because it occurs continuously. The next question is how often it occurs. If it occurs continuously, the size of restite must be very small. Therefore, delamination seems be impossible. For example, the subduction of arcs is common under SW Japan, where several intra-oceanic arcs easily subduct to the deep mantle without accretion (Yamamoto et al., 2009).

A key to resolve the problem of delamination could be the petrogenesis of adakites. Typical adakites are characterized by HREE differentiated patterns suggesting a garnet-bearing source rock, e.g., eclogite. If the angle of subduction is around 45 degrees, like in present-day NE Japan, the thickness of the lower mafic crust right below the volcanic front is only 30-35 km, i.e., not enough to get eclogite. The melting conditions are still within the stability field of plagioclase, therefore, the primary magma derived beneath NE Japan cannot be adakitic. If we consider that the composition of the granite crust is similar to adakite, its source cannot be in the lower crust, but much lower, to provide the melting of subducted MORB slab (Rapp and Watson, 1995). In that case, delamination of the lower mafic crust is possible. However, granitic rocks with highly depleted HREE are actually rare. Typical adakites form at the subduction of young slabs, around several million years old. The subduction zones related to those very young slab subduction zones are present in the Cascadia off Seattle, and in present-day SW Japan (Deviant and Durmont, 1990).

The composition of the volcanic and plutonic rocks observed at subduction zones may not match that of adakite though, in particular, in the case of young slab and flat subduction. For example, the composition of Late Miocene to Pliocene granitoids from the Taitao Peninsula near the Chile ridge subduction zone resemble those of TTGs and calc-alkaline series, but not adakites (which must be produced by melting of young and hot oceanic crust under eclogite to garnet amphibolite conditions). The Taitao granitoids were generated by partial melting of the subducted oceanic crust in garnet-free amphibolite conditions at depths shallower than 30 km. Therefore, slab-melting-related magmas do not necessarily show a HREE-depleted signature, which was used as evidence for slab-melting for granitic rocks of the TTG suites (Kon et al., 2013).

On the other hand, direct slab melting is possible if the continental crust is as thick as 50-60 km to form eclogite from amphibolite, because the stability field of eclogite expands to the high temperature part of PT-diagrams, even if the pressure is not very high. A typical example is the Tibetan plateau, where there is no possibility of slab melting, but the continental crust is thick enough to produce Eocene to Miocene dacite and granite of adakitic geochemical affinities (Chung et al., 2003). Thus, the model of adakite formation cannot always explain the petrogenesis of island arc granitoids.

Petrogenesis of TTG. The production rate of granite in Pacific-type orogenic belts and the origin of continental crust must be discussed from a different viewpoint, beyond petrology and geochemistry. First, during the Phanerozoic the continental was forming not permanently peaked in the middle Paleozoic and Cretaceous. The youngest peak of the formation of TTG was at 120-80 Ma (Larson, 1991), which well-studied examples are the Cretaceous batholiths of the Circum-Pacific and, to a lesser extent, Tethyan granitic batholiths. The 40 Ma of the Cretaceous granitoid magmatism formed a 200 km wide belt suggesting 30 km thick crust, i.e. the thickness of the upper felsic crust must be 15 km. Therefore, the average production rate of batholith formation is only 1 km3 per year and only 1 km wide trench. Then we must compare the volume of magma produced at mid-oceanic ridge (to make a slab) and the volume of new crust generated at island arcs (Reymer and Shubert, 1984; Engebretson et al., 1985). If we consider a 1 km wide belt of Cretaceous TTG along the circum-Pacific, then we must multiple 200 km (width) and to 15 km (thickness) and divide for a period of 40 Ma giving a value of production rate of 75 km^3 per million year (200 km× 15 km/40 m.y. = 75 km^3/m.y.). Since the Cretaceous, the average rate of mid-oceanic ridge spreading is about 5 cm/year (Engebretson et al., 1985). Then the rate of production of mid-oceanic ridge basalt is 5 cm/year to give a growth of 50 km/m.y. at the oceanic crust thickness of 6 km to give a volume of 300 km^3/m.y. The melting of 6 km thick MORB oceanic crust will generate 20-30% of TTG crust, i.e., 60-90 km3/m.y., i.e., comparable with the value of 75 km^3/m.y. These results suggest that 100% of the Cretaceous TTG can be formed by slab melting. On the other hand, island arc magma can be derived through fractional recrystallization of basalt produced by the melting of mantle wedge. In that case the production of island arc magma will be at 20% melting (Reiner and Shubert, 1984). The amount of 20% in respect to 300 km3/m.y. is equal to 60 km^3/m.y. This is a volume of primary arc basaltic magma, which then can be partially melted to 20-30% to yield 12-18 km3/m.y. of the 15 km thick upper felsic crust. The rate of production of continental crust is accordingly only 12-18 km3/m.y., which cannot explain the observed value of 75 km^3/m.y. Therefore, the Circum-Pacific Cretaceous granitic crust must have been formed nearly all by slab melting and the delamination process of lower mafic crust, to produce such amount of new crust, is not necessary, even in terms of petrogenesis.

If arc andesite forms by the 20—30% partial melting of arc basalt underplated at the Moho depth (Fig. 26a), the restite at Moho must be transported into the deep mantle by delamination (Fig. 26b), because of the thick lower crust over-accumulated in time (Nakajima and Arima, 1998; Tatsumi, 2000). In the next chapter we will show that the delamination of the mafic lower crust is not really necessary to explain the formation of arc andesitic/TTG-type magma because of the effect of tectonic erosion.

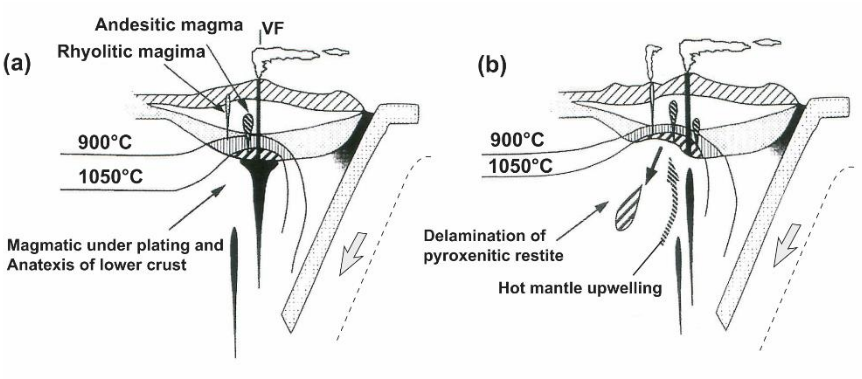


Fig. 26. The role of delamination in the formation of arc andesite.

Chapter 4. Tectonic erosion and Pacific-type orogeny

4.1 Tectonic erosion as a process

The first evidence for tectonic erosion at P-type active margins was obtained from seismic reflection profiles across the Tonga (Hilde and Fisher, 1979) and Japan trenches (von Huene and Uyeda, 1981; Hilde, 1983), showing prominent sediment-filled grabens and horsts with thinned sediments (Fig. 27a, b). The mechanism of tectonic erosion includes destruction of oceanic slab, island arcs, accretionary prism and fore-arc by thrusting, the prominent oceanic floor relief as horsts, grabens and oceanic rises scratching the hanging wall, and by (hydro)fracturing (von Huene et al., 2004; Yamamoto et al., 2009) (Fig. 27c, d). At subduction-related convergent margins, the TTG-type crust and material of accretionary wedges can be eroded and subducted with oceanic slab into the deep upper mantle. Evidence for this comes from the Cretaceous Shimanto accretionary complex of Shikoku Island in Japan, where accretionary units are spatially adjacent to the coeval granitoids of the Ryoke belt, suggesting that older accretionary complexes have been eroded (Fig. 3; Safonova et al., 2015, 2016), and from the present-day tectonic erosion and ongoing subduction of several oceanic arcs of the Philippine Plate under the Japanese Arc (Fig. 8).

Two contrast types of Pacific-type convergent margins – accreting or growing and eroding or narrowing - have been recognized so far (Scholl, von Huene, 2007). Eroding or non-accreting margins are characterized by the close approach of the margin's rock framework to the trench and small or lacking older prisms of accreted lower plate sediment (Fig. 27e, f). With time, eroding margins narrow with respect to a reference point on the margin; i.e., the trench advances landward. Accreting or growing margins are characterized by rocks deeply buried under thick older accreted units and frontal prism of actively deforming sediment scraped off the subducting plate. With time, accreting margins widen, i.e., the trench retreats seaward. The modern Pacific Ocean is surrounded by 75% of eroding (narrowing or non-accreting) convergent margins and 25% of growing or accreting margins. The global long-term rate of subduction-related erosion is much greater than that of crustal additions (Senshu et al., 2009; Stern, 2011). Tectonic erosion destroys orogenic components and structure, i.e. the basic framework of large-scale geological structures of Pacific-type orogenic belts from the trench continent-ward: accretionary complex high pressure regional metamorphic belt, fore-arc arc basin deposits and magmatic belt (Figs.

1, 28). The fourth often represents a granitoid batholith belt. The whole structure can be 2000-3000 km long and 300-400 km wide. All these four components of P-type belts can be tectonically eroded and thus transported to the deep mantle.

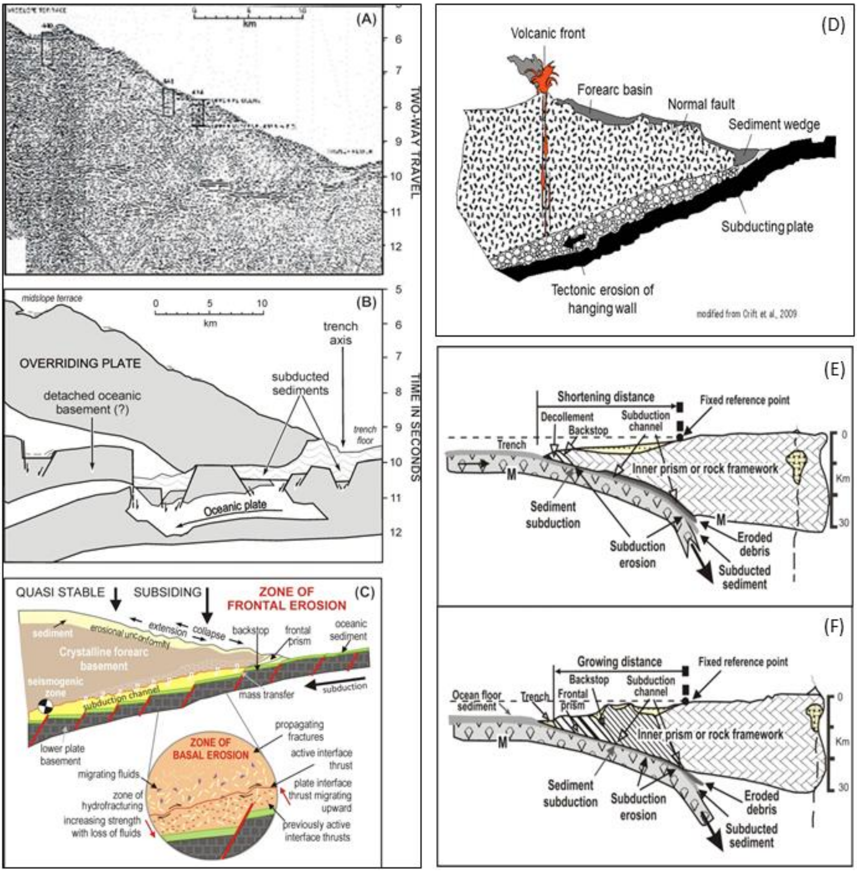


Fig. 27. Tectonic erosion at Pacific-type convergent margins. A and B, multi-channel seismic reflection profile plate (A) and its interpretation (B) across the axis and lower slope of the Japan trench along the N 35°45' latitude, showing the subducted oceanic slab (modified from Hilde, 1983). The data for the profile were collected by the Japan National Oil Corporation for the IPOD Japan Trench transect (Shipboard Scientific Party, 1980). C, subduction-erosion model indicating areas of most severe fracturing and dragging of dislodged fragments into subduction channel (from von Huene et al., 2004). D, erosion of arc hanging wall. E, F – eroding and accreting convergent margins after (Scholl and von Huene, 2007).

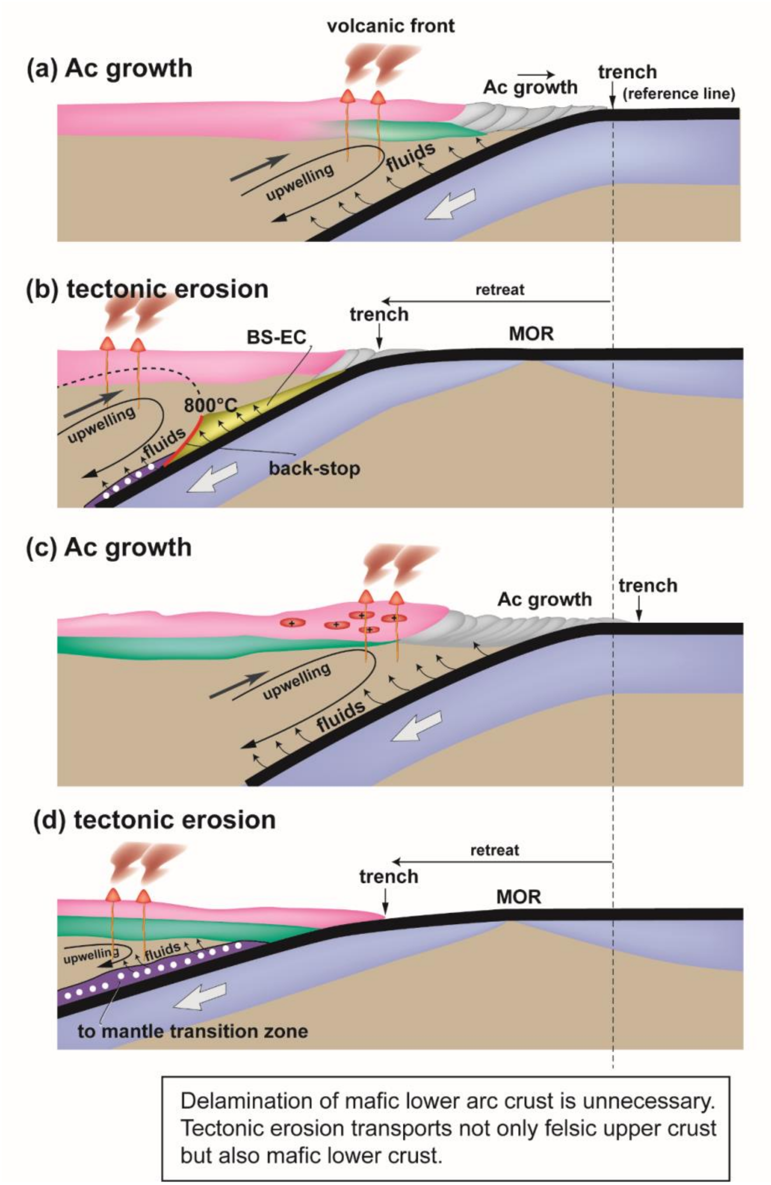


Fig. 28. A fundamental process of a Pacific-type orogeny is a rapid back-and-forth process caused by tectonic erosion and by continental growth during formation of an accretionary complex and batholith belt (Maruyama et al., 2011).

Recently the zones of subduction have received another focus as the only corridors on Earth through which the surface materials of both oceanic and continental crust can be delivered to the deep mantle (Maruyama et al., 2007; Yamamoto et al., 2009; Kawai et al., 2013; Safonova et al., 2015). During the last years more and more scientists have focused their research on the fate of subducted materials and their accumulation in mantle transition zone, MTZ (Kawai et al., 2013; Safonova et al., 2015).

So, the tectonic erosion at P-type convergent margins is also a trigger for supplying continental crust material into the deep mantle. As the tectonically eroded and subducted material can accumulate in the MTZ, the supply of large amounts of continental and oceanic crust material down the deep mantle may significantly affect the temperature of melting and the composition of their hosting mantle. This, in turn, can trigger shallow mantle plumes and their related within-plate magmatism like that widely manifested in Mongolia, eastern and NE China (for details see Chapters 5 and 6).

The normal stage of Pacific-type orogeny is defined by accretionary growth oceanward together with TTG (andesite in composition) growth at the volcanic front (Fig. 28a). When the mid-oceanic ridge approaches the trench, extensive tectonic erosion occurs, transporting TTG material at depths of 50—60 km and at T-max. 800℃ where they are recrystallized along the BS-EC facies series by the presence of a back-stop at the mantle depth. They return to the surface due to the wedge extrusion process (Maruyama, 1990, 1997) by decreasing the subduction angle over time. Simultaneously, huge amounts of TTG crust are accreted to the arc formed by slab-melting underneath the volcanic front (Fig. 28b). Note the parallel shifts of trench line and volcanic front landward during tectonic erosion and backward during accretionary growth (Fig. 28c). Tectonic erosion starts again when the mid-oceanic ridge approaches the trench. It occurs every 100 Ma and forms a set of four elements in a Pacific-type orogen cored by the regional metamorphic belt (Fig. 28d). Thus, tectonic erosion is recorded in the past geological history of the Japanese islands back to 520 Ma.

In the Japanese Islands (Fig. 3), these four components of P-type belts always form as a set of five Pacific-type orogens formed during the last 500 Ma (Fig. 29). However, we don’t know if those orogenic belts have been preserved well and their abundance is relative. Some of them have been well preserved, the best examples are the Sanbagawa HP-metamorphic belt and Ryoke LP- belt, however, their related fore-arc basin deposits are very limited. The Izumi fore-arc basin has been tectonically shortened and lost. Its north-to-south width is about 100 km;

therefore, paired metamorphic belts are almost in direct contact with it (Miyashiro, 1961; Aoki, 2009, 2010). There are also pre-Sanbagawa or pre-Cretaceous Pacific-type orogenic belts: Sangun blueschist-bearing regional metamorphic belt (180-170 Ma), Chizu glaucophane-schist belt (250-200 Ma), Renge blueschist-eclogite belt (360-330 Ma) and the oldest subduction-related metamorphic belts of 450 Ma. These belts are few kilometers or even smaller fragments of larger regional metamorphic belts. The rest of the four components also occur as fragments or absent. Among them, the most representative is the Kurosegawa belt hosting 450-340 Ma blueschist or other HP-LT regional metamorphic rocks and 250-200 Ma arc including ryolite-dacite-andesite volcanics, TTG-type granitoids, higher-grade amphibolite or even HP granulite faces rocks. All those lithologies occur as small geologic units, each rock type coexisting and indicating a mixture between trench and volcanic front. A vertical cross-section suggests that tectonic erosion is the most preferred process to mix all those units together, for example, to form serpentinite-dominated mélange belts (Fig. 29a).
Tectonic erosion is a new concept, which seems to be extremely important to discuss the history of the Earth, the growth of continental crust and the circular change of mantle convection in order to understand all geological phenomena and the evolution of Earth systems. In addition, tectonic erosion can explain the petrogenesis of granitic magma without delamination of the lower crust. But first of all, tectonic erosion plays a very important role of the evolution of Pacific-type orogens, because their components (TTG, fore-arc, metamorphic and accretionary units) all can be transported into the deep mantle. The observed constrains for the history of the Japanese Islands indicate that extensive tectonic erosion triggers the rapid growth of continental crust. For example, after the extensive tectonic erosion during the Cretaceous accretion in Shimanto was accompanied by the formation of the 50 km wide Sang-in granitic batholith belt. The accretionary complex expanded oceanward to approximately 200 km toward the trench, and the continental crust increased to about 100 km (Aoki et al., 2010). A key point is the age of slab (Fig. 28). Subduction of young slab is followed by strong tectonic erosion. The trench deposits were transported to Moho depths beneath the Japanese Islands together with a part of oceanic slab and contributed to a domed-up structure (see Chapter 2, sections 2.3, 2.4). During about 100 Ma after the large-scale tectonic erosion a new batholith belt probably formed oceanward paired with the growing oceanward accretionary complex to create a new Pacific-type orogen. The trench again moved oceanward to its original position. The volcanic front also shifted oceanward, parallel to the belt and the trench.

Therefore, the back-and-forth migrations of the trench and related volcanic front take place during each episode of large-scale tectonic erosion (Fig. 28). As the result, the position of the trench is more or less stable (Figs. 28, 30).

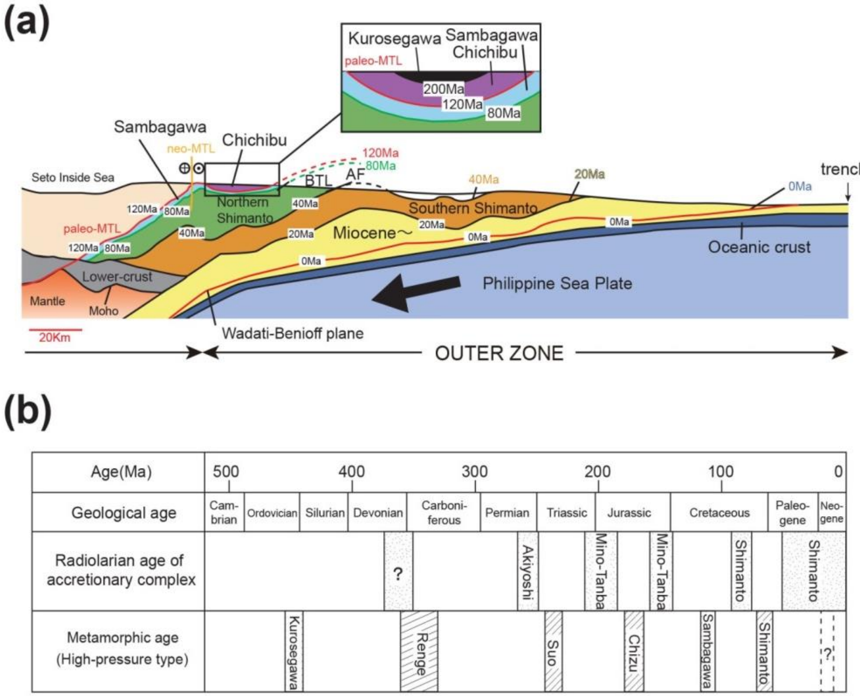


Fig. 29. Evidence for tectonic erosion on the Japanese Islands. A, a schematic illustration of NS cross-section of SW Japan. B, the ages of the geologic units. A and B show locations of extensive tectonic erosion, where the serpentinite mélange belt is present as a proxy of geologic body of tectonic erosion.

Figure 30 shows a summary of the geotectonic history of the Japanese islands. First, accretionary complexes expanded oceanward to a distance of ca. 200 km toward the trench (Fig. 30a). A normal arc stage defines accretionary growth oceanward and TTG growth at the volcanic front. Tectonic erosion occurs every 100 Ma to destroy a major part or the whole arc (Fig. 30b). During extensive tectonic erosion, the mafic lower crust is transported into the mantle, hence steady-state delamination of the mafic lower crust is not necessary (see Chapter 3).

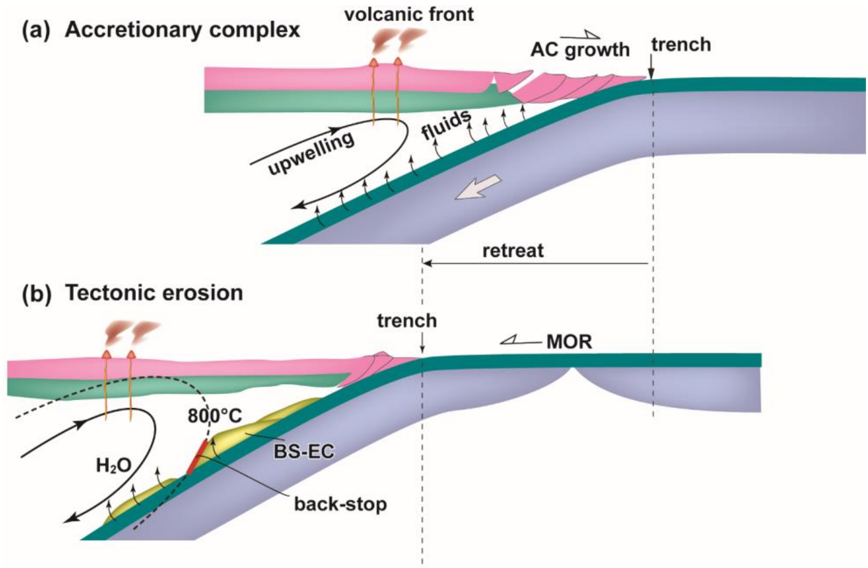
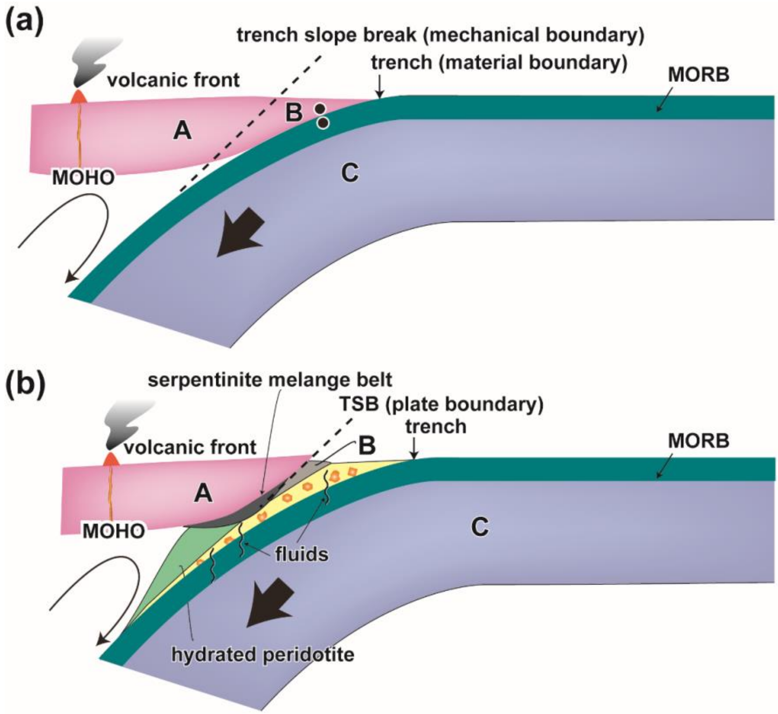


Fig. 30. Tectonic erosion is common around P-type convergent plate boundaries around the world.

The frontal part of island arc crust was extensively destroyed and transported into the mantle to suffer subduction zone metamorphism. When a mid-oceanic ridge approached the subduction zone, those blueschist-eclogite metamorphic rocks started ascending as a thin tectonic to mid-crust level along the branched Benioff thrust, which is the upper surface between arc and trench slope break, i.e. intermediate between the volcanic front and the trench (Fig. 31). A fore-arc basin formed between the newly formed non-volcanic second arc and the volcanic front. After the ridge subduction, another plate, which age was gradually increasing, started to subduct (Fig. 32). For more details see below sections 4.3 and 4.4.

Why the older orogenic belts of the Japanese islands are typically smaller in size and show incomplete sections? The answer is tectonic erosion along the Circum-Pacific subduction zones. Why the formation of accretionary complex is a rather exceptional event? A common idea since the 1970-ies is that when oceanic lithosphere subducts under



(c)

520Ma subduction initiation

Tectonic setting / Age	580Ma	450Ma	360Ma	250Ma	240-208Ma	Early Jurassic?
volcanic front	ophiolite	calc alkaline rock limestone volcanic rock clastic rock granite gneiss, amphibolite		granite		fore-arc sediment
subduction zone			eclogite amphibolite blueschist		blueschist (jadeite～glaucophane type)	

Fig. 31. Formation of serpentinite mélange belt allows transportation of tectonically eroded materials into the deep mantle. A small triangular region "B" mechanically belongs to a descending oceanic slab "C", although materially it is in the left side of the trench, i.e. belongs to the volcanic arc hanging wall (material boundary). The mechanical boundary corresponds to trench-slope break (TSB) shown by dotted line (a) (Nakamura and Shimazaki, 1981). When extensive tectonic erosion occurs, eroded blocks of various compositions and ages can move down to the mantle (b). The bit younger age of BS facies rocks than the CA volcano-plutonic sequence (c) (Suzuki et al., 2010).

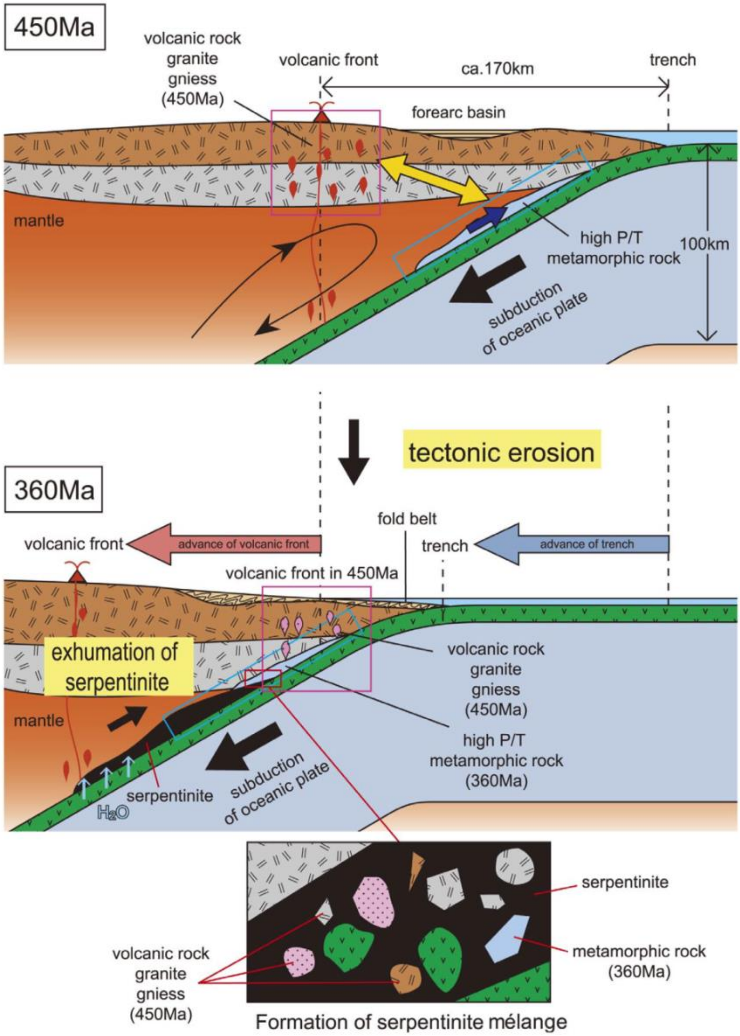


Fig. 32. Schematic cross-sections showing the mechanism that formed the serpentinite mélange due to extensive tectonic erosion. Above: 450 Ma, an era of continental growth. Subduction of an oceanic plate beneath a continent formed a volcanic front 170 km inside from the trench. A forearc basin formed between the trench and the volcanic front. Below: 360 Ma, destruction of the continent from the former trench to the 450 Ma volcanic front. New volcanic front and trench moved together to the west for a distance over 100 km. 360 Ma regional metamorphic rock and serpentinite exhumed along the slab surface, and captured 450 Ma magmatic components beneath the previous volcanic front at a shallower depth (Suzuki et al., 2010).

continental margin, the frontal portion of the hanging wall of continental margin must growth oceanward to form accretionary complex. Therefore, orogenic belt must also grow oceanward meaning trench retreat. However, the real geology contradicts that idea because the structurally upper but older orogen occurs always fragmentally and is smaller in size than a younger one.

New break-through came from marine geophysicists made seismic profiling across Pacific subduction zones, from trench and landward (Hilde and Fisher, 1979; von Huene and Uyeda, 1981; Hilde, 1983). The others summarized all the research of subduction zones from the whole Circum-Pacific and concluded that only 25% of convergent margins are accreting, but 75% are eroding (Clift et al., 2004; Scholl and von Huene, 2007).

4.2 Evidence of tectonic erosion

In Japan, the ongoing tectonic erosion is ubiquitous and appeared much stronger, than marine geophysicists show, even in the rest 25% of subduction zones. The toe of the Shimanto accretionary complex is in the Nankai trough, where the Philippine Sea plate is subducting under the southwestern Japan and extensive tectonic erosion is ongoing. The top of the Philippine Sea plate is covered by trench turbidites; part of them is now transported to the deep mantle and another part is attached underneath the lower crust (Park et al., 2002). In the western margin of Pacific region, in particular, in southwestern Japan, five intra-oceanic arcs are now being subducted into the mantle (Yamamoto et al., 2009). However, the deformation of the trench is negligibly small, indicating negligible accretion. The collision of the Honshu Island and Izu-Mariana arc induced oroclinal bending. The extensive erosion and accretion of island arc crust and overlying sediments must be continuous during at least the last 15 Ma, however, geological investigations show less than 1% of accreted material suggesting that 99% of the crust must have subducted and annihilated during the arc-arc collision and related orogeny (Zhao et al.,1994; Yamamoto et al., 2009; Zhao, 2009). Moreover, Honshu arc continental crust about 50% is tectonically destroyed and brought to the deep mantle, it is clear by Hirato et al, 2010. The 15 Ma paleogeography of the Japanese Islands shows that the position of the Ogasagawa and Nankai trenches were 150 km oceanward (Hirata et al, 2010). Thus, the degree of tectonic erosion appeared much higher compared to that evaluated by marine geophysicists.

The tectonic erosion of the last several dozens m.y. is ubiquitous in subduction zone trenches in a global scale, in particular, in the Circum-Pacific region. There, tectonic erosion can be studied by remained accretionary complexes formed during the last 500 Ma. We can explore older episodes tectonic erosion and understand mechanisms of tectonic erosion by examining what kind of geologic structure and what kind of rocks remained on the surface. All these can be understood by studying on-land geology, however, no systematical geological research of tectonic erosion in on-land orogenic belts has been done yet.

It is easy to identify the timing of tectonic erosion. Figure 29 shows the north-south cross-section of the geologic structure of Japan, including the structural bottom of the descending Philippine Sea plate. Fig. 30 is a summary of the formation of accretionary complex, which looks discontinuous. Geological data show that the subduction was continuous during the last 520 Ma, however, accretionary complexes have been reconstructed for only 30% of geological time. Of course, destruction of accretionary complex must be stronger with geological time as tectonic erosion affects the already existing hanging wall of an accretionary complex and orogenic belt. However, the degree of tectonic erosion is very high even at younger P-type convergent margins. A new accretionary complex probably formed above the Benioff plane, as marked by 0 Ma on the cross-section (Fig. 29). Similarly, 20 Ma, 40 Ma, 80 Ma, and 120 Ma are marked on the cross-section referring a geophysical exploration by Ito and Sato (2010). A sudden jump in numbers is seen in the enlarged inset figure on the structural top of SW Japan arc, where the Kurosegawa zone rests on top. It is a serpentinite mélange belt including a 450 Ma volcano-plutonic subduction-related sequence of amphibolite, high-pressure granulite and BS facies rocks formed at 360 Ma and 250 M (vertical sequence from Moho to the surface at 450 Ma). Hence, at present time a big portion is missing between the arc front and the trench over a distance of 150 km. Formation of accretionary complex is not restricted by shallow levels, but goes deeper. The position of 15 Ma line from the trench to a depth of 30 km corresponds to the Benioff plane at 15 Ma in SW Japan. The formation of the 15 Ma accretionary complex can be reconstructed by on-land geology and we can trace the accretionary complex to deeper levels at which it is folded. The south to west cross-section indicates the Benioff plane at 35 Ma and 80 Ma. Therefore, 50% of the continental crust of SW Japan is occupied by a very young Shimanto accretionary complex formed during the last 80 Ma. Moreover, the lower mafic crust is absent in SW Japan (Ito and Sato, 2010) providing more evidence for tectonic erosion (Fig. 29).

4.3 Mechanism of tectonic erosion

The tomographic and geophysical data from fore-arc regions allowed marine geophysicists to propose three different models or mechanisms of tectonic erosion. (1) The bucket model implies the presence of horsts and grabens in the structure of the subducting plate near the trench wall (Hilde and Fisher, 1979; von Huene and Uyeda, 1981; Hilde, 1983). (2) The destruction of the ceiling of the hanging wall by topographic rises such as seamounts and horsts of the horst-and-graben structures formed by normal faults cutting the subducting plate (Kodaira et al., 2000). (3) Pore pressure exceeds hydrostatic pressure to destroy large blocks into small pieces (Yamamoto, 2010). (4) Formation of serpentinite mélange (Suzuki et al, 2010). Mechanism no. 4 seems to be the most important and will be described below in more details.

In SW Japan (Fig. 3), the Jurassic and Cretaceous accretionary complexes are overlain by Middle Jurassic (170 Ma) serpentinites. Those serpentinites host blocks of lithogically and geochronologically different rocks: (a) Late Triassic jadeite-lawsonite-glaucophane schists (208 to 220 Ma); (b) a Permian accretionary complex; (c) early Carboniferous glaucophane schists (360 to 340 Ma); (d) late Ordovician granitoids (450 Ma) and (e) high-grade amphibolite and mafic granulite formed at 10 kbar and 800°C, kyanite-sillimanite and epidote-amphibolite metamorphic rocks; (f) Silurian-Devonian supracrustal rocks (calc-alkaline volcanics and platform sediments) and (g) a Carboniferous-Permian-Triassic coherent sequence of sediments formed in a fore-arc basin or on a continental platform. All those diverse rocks, both crust-derived and mantle-derived, are chaotically mixed (Maruyama, 1981; Maruyama et al., 1984). We can classify these rocks into several groups though and five the following interpretations.

1. 450 Ma HP granulite volcanic formed under an island arc at a Moho depth and about 100 Ma younger jadeite-glaucophane schists formed by subduction.

2. Contrasting rocks types: 250 Ma granite and surface supracrustal rocks and 50 Ma younger LT-HP glaucophane schists formed in one tectonic setting.

3. The 450-350 Ma and 250-200 Ma rock groups should have formed at a distance horizontally 150 km from each other in different tectonic settings: volcanic front versus deep-seated subduction zone.

4. To be mixed together within a 150 km wide intermediate region, these two groups first must be tectonically eroded and transported to the mantle. Then the rocks at the bottom of a section beneath an older

volcanic front can be juxtaposed with much younger subduction-related rocks. As a result the older volcanic front should move oceanward to become a new trench, and a younger volcanic front moves landward, 150 km inland the continent. The time gaps in the case of the older pair (450 vs. 350) and younger pair (250 vs. 200Ma) are about 100 Ma and 50 Ma, respectively.

The tectonically eroded fragmental blocks can be transported to the deep mantle along the new trench (Fig. 31a). During that process the upper overlying mantle wedge peridotites must be hydrated to become serpentinite along the Benioff thrust. Therefore, that tectonic boundary should become a new Benioff plane. Below that new Benioff plane the oceanic slab subducts together with deep-sea sediments, trench turbidites and tectonically eroded fragments set into matrix, presumably, trench turdidites. Those rocks can be transported into the deeper mantle by the descending slab and experience dehydration to release fluids and cause serpentinization of the hanging wall of the mantle wedge (Figs. 22, 32; Peacock, 2001; see chapter 3, section 3.1). During serpentinization the density of the ambient material shifts from 3.3-3.4 g/cm^3 for peridotites to 2.7-2.0 g/cm3 for serpentinites depending on the degree of serpentinization. As a result, the volume of serpentinite drastically increases, the viscosity decreases and a new decollement easily forms to make a new Benioff plane. Without such serpentinization, the tectonically eroded huge blocks of hanging wall hard rocks cannot be easily transported to the mantle. Tectonic blocks formed in front of a descending slab must be cemented by matrix sediments, presumably facilitated by dehydrated fluids underneath, and pasted tightly together with the underlying slab, otherwise the sediments could not be subducted. The formation of serpentinites directly above the descending slab creates a new decollement zone, which acts as a new Benioff plane, making it possible for tectonic blocks to subduct into the deep mantle. Note that the serpentinite zone develops along the mechanical boundary and not along the material boundary. When extensive tectonic erosion occurs, eroded blocks are cemented by trench turbidite and descend into the mantle to cross the mechanical boundary at mantle depth. If tectonic blocks are loosely cemented with the oceanic slab, mantle transportation does not occur. Instead, accretion to the hanging wall occurs. If a mechanically very weak rock such as a serpentinite is formed above the tectonic blocks, tectonic transportation of those blocks becomes possible (Fig. 31b). The most extensively developed serpentinite mélange belt includes two remarkably different types of rock such as a vertical sequence of volcanic front from the surface to the Moho depth and high-P/T schists formed above the Benioff plane, suggesting extensive shortening between trench and volcanic front over 150 km. The protrusion of serpentinite along the Benioff thrust tectonically captures recrystallized LT-HP metamorphic rocks of the same age (250 to 200 Ma) and the rocks of the 450 Ma basement together with supracrustal rocks already emplaced at mid-crustal levels. All rock blocks are hosted by serpentinite as xenoliths, and the protruding serpentinite finally reaches the surface (Fig. 31b). The Kurosegawa belt carries traces of two episodes of tectonic erosion, at 200 and 340 Ma, and thus represents a typical locality of strong multi-stage tectonic erosion. This mechanism of exclusively strong tectonic erosion is different from that proposed by marine geophysicists based on seismic profile data over the inner wall of trench.

4.4 Importance of serpentinite mélange

The serpentinite, which protrude along the Benioff plane and carry fragments of volcanic arc, sedimentary and subduction-related jadeite-glaucophane rocks represent an indicative geological evidence of past tectonic erosions. The formation of serpentinite mélange belts cannot be inferred from seismic profiles made across the inner wall of trench by marine geophysics though. Thus serpentinite mélange belts provide important constrains to unravel past tectonic erosions based on on-land geology of Pacific-type orogenic belts and, probably, of C-type belts as well. As was shown above the Kurosegawa serpentinite belts carry evidence of, two large-scale tectonic erosion episodes at 340 and 200 Ma. By these two cases of tectonic erosion we can roughly estimate the amount of eroded mass, which could be equivalent to the present-day Japanese Island. Therefore, such a huge amount of crust was probably removed to the deep mantle. However, why, in spite of such strong tectonic erosion, old accretionary complexes and other geological units and sedimentary rocks survive?

The most typical rock units of the Kurosegawa zone are in the Shikoku Island (Fig. 3), which rapidly uplifted in mid-Miocene time. In the Key Peninsula of Honsu, an island next to Shikoku, the Kurosegawa zone is absent. However, in Shikoku, the late Cretaceous Sanbagawa blueschist belt is domed up together with the secondary high-angle normal fault as seen in north-south cross-section (Fig. 29a). There is a pair of antiform and synform structures. Within the synform the Sanbagawa axis is above the axis of the Kurosegawa zone, which is structurally higher.

What is the mechanism to tectonically erode a very old slab during subduction, like in the case of the Mariana arc? First, tectonically eroded huge rock fragments, may be up to several kilometers across, slide down as a submarine landside to form olistostrome deposits at the trench together with the cementing matrix. Then, all those units started subducting together with the descending Pacific slab. The first tectonic plate boundary is extension of the branching fault running from deep levels of the Benioff plane to the surface of the arc-trench gap, which is the trench slope-break or TSB (Fig. 31). The small triangular region forms between the hanging wall of the stable plate and the subducting oceanic plate and is separated from the hanging wall by the TSB mechanical boundary. During tectonic erosion the eroded blocks are cemented by trench turbidite and descend into the mantle to cross the mechanical boundary at mantle depths (Fig. 31a). If tectonic blocks are loosely cemented with the oceanic slab, mantle transportation does not occur. Instead, accretion to the hanging wall occurs. If a mechanically very weak rock such as a serpentinite is formed above the tectonic blocks, tectonic transportation of those blocks becomes possible (Fig. 31b). Tectonic blocks formed in front of a descending slab must be cemented by matrix sediments, presumably facilitated by dehydrated fluids underneath, and pasted tightly together with the underlying slab, otherwise the sediments could not be subducted. The formation of serpentinites directly above the descending slab creates a new decollement zone, which acts as a new Benioff plane, making it possible for tectonic blocks to subduct into the deep mantle. Note that serpentinite zone develops along the mechanical boundary, not along the material boundary. The most extensively developed serpentinite mélange belt includes two remarkably different types of rock such as a vertical volcanic front from the surface to the Moho depth - and high-P/T schists formed above the Benioff plane, suggesting extensive shortening between the trench and the volcanic front over 150 km (Fig. 31).

Serpentinite forms by the fluid derived from the slab and hydrating the wedge mantle. Serpentinization drastically changes the elasticity of mantle peridotites, which become more ductile and their density drops from 3.5 to less than 2 g/cm3 because of significantly increased volume. Then, the expanded serpentinite ascends along the upper tectonic boundary and then gradually extrudes. Thus, the eroded blocks can be easily transported into the mantle together with the subducting slab. At the Mariana trench, the Pacific slab is also cut by normal faults oceanward by the extension of that very old plate as a result of a high angle of subduction. Therefore, the material of the oceanic plate can be also fragmented to cause landslides to fill the trench forming another type of olistostrome also transported to the deep mantle by the subducting slab.

In summary, tectonic erosion depending on the type of slab occurs not only at the frontal part of the overriding plate, but also at the subducted top of the Benioff thrust underneath. Depending on the angle of subduction, the surface of the overriding plate, in particular, the non-volcanic arc and the fore-arc basin, can escape tectonic erosion, because of extensional tectonics. Nevertheless, a very strong horizontal compressional stress causes significant horizontal shortening and/or isoclinal folding, and the reverse fault causes a very complex deformed structure (Fig. 32).

Serpentinite intrudes into complex geologic units, deform them and are also deformed together with diverse rocks trapped as blocks in the matrix of serpentinite (Maruyama, 1981) (Fig. 32). The presence of serpentinite belts in P-type orogeny is very important in terms of tectonic erosion, which destroys the landward hanging wall in front of the trench and the top of the slab by subducting topographic highs (Fig. 27). It is easy to imagine how the descending slab carries those fragmented blocks into the deep mantle. However, those blocks can tectonically return to the surface by moving in a direction opposite to the subduction. How can we explain serpentinite mélange belts, which contain a variety of rock types from the volcanic front and the arc underneath to the subduction-related HP metamorphic rocks? Figure 31 shows a possible scenario, which key point is a fault branching from the Benioff or decollement zone starting at a depth of about 50 km. The branching fault connects the deep-seated subduction zone with the fore-arc and second non-volcanic arc. This is a big fault dividing the descending Pacific slab together with the small corner of triangular regions (Fig. 31). The subducting slab dehydrates to generate a large amount of water at a depth of ca. 50 km to rise toward the branching fault. The released fluid causes serpentinization of fore-arc mantle peridotites. The expanded serpentinite move up along that fault to protrude and making a kind of serpentinite seamount (mount?) on the surface above the trench inner wall. The serpentinite protruded along the branching fault trap the rocks of the hanging wall and finally to expose (squeezed out) to the surface.

4.5 World examples of tectonic erosion: Japanese, Islands, Izu-Maria arc, British Isles

At the beginning of the Meiji Era the geotectonic subdivision of Japan was characterized by the along-arc zonal arrangement of narrow geologic units traditionally called belts, which were differentiated from each other by their rock type and megafossil age. The two most striking tectonic boundaries in Japan were Fossa magna in central Japan, which divides NE Japan from SW Japan, and the Median Tectonic Line (MTL) in SW Japan, which divides the continent side (Inner zone) from the ocean side (Outer zone) of SW Japan arc (Fig. 3). In addition to these two, the Tanakura Tectonic Line (TTL) in NE Japan, and assumed Western Kyushu Tectonic Line represent two remarkable tectonic features that define the eastern and western margin-bounding faults of the back-arc basin, the Japan Sea that opened during the Miocene. These apparently linear faults have been regarded to be the most important tectonic features of Japan for a long time; however, they are too young to constrain the overall orogenic framework of Japan, which can be traced back to 600 Ma (e.g., Yanai et al., 2010). Instead, these faults merely represent Miocene micro-plate boundaries; e.g., Fossa faults as a Miocene ridge-transform fault system, and MTL as a fore-arc contracting boundary within SW Japan during the opening of the Japan Sea. Although less conspicuous when identified in the field, more important and critical boundaries of neighboring orogens were later proposed by Isozaki and Itaya (1991), and thereafter by Isozaki and Maruyama (1991), with special reference to the upper and lower boundary faults between non-metamorphosed accretionary complexes (AC) and/or high-P/T metamorphosed AC units. The latest data and new perspectives obtained from the detrital zircon chronology suggest we need to pay more attention to serpentinite mélange belts that define large age-gaps within the multiple stacks of AC, high-P/T AC, or ophiolites belts (see section 4.4). Serpentinite mélange belts often contain tonalite-tronjemite-granodiorite (TTG) and related rocks, together with coeval AC and also jadeite/ glaucophane-bearing blueschists, although all in small amounts. The mixing of these rocks indicates severe shortening of a past fore-arc crust between deep-subduction zone and volcanic front, in other words, the largest-scale tectonic erosion to juxtapose rocks of trench and volcanic arc, extending over 150 km wide (Figs. 30-31). Serpentinite mélange belts occur as a subhorizontal klippe on the top in SW Japan, where ca. 580 Ma ophiolite occurs as the oldest unit in Japan. On the basis of the latest data, in particular, the detrital zircon chronology obtained by laser-ablation induced coupled-plasma mass spectrometer (LA-ICP-MS) rapid analysis (Nakama et al., 2010), the geotectonic subdivision has been thoroughly revised (Isozaki et al., 2010). When plate tectonics was introduced in regional metamorphic belts in Japan, Y. Isozaki and S. Maruyama (1991) first introduced the migration of the oceanic plates on the Pacific Ocean side, originally proposed by Maruyama and Seno (1986), which has been derived by the Stanford group of paleo-reconstruction of plates in the Pacific Ocean. The models have come from relative plate motion suggest episodic periodical ridge subduction should be related to the periodical Pacific-type orogeny occurred (Maruyama, 1997). By this plate reconstruction back to Jurassic, 180 Ma, a paleo-geographic reconstruction of Japan has been proposed combining obtained OPS from Japan with paleo-plate geometry (Maruyama et al, 1997). The basic idea for these researches shows the constant growth of accretionary complex oceanward. However, recent discovery tectonic erosion around the circum-Pacific subduction zones by (von Huene and Scholl, 1992) showed the importance of tectonic erosion by which hanging wall of continental crust including accretionary complex has been tectonically collapsed to remove it into the deep mantle. Therefore, such kind of surprising speculation by marine geophysicists astonished Japanese geologists. In addition to marine geophysicists, several Japanese geologists proposed a concept of tectonic erosion based on detailed on-land geology of serpentinite mélange belts (Suzuki et al., 2010, Isozaki et al., 2011; Figs. 33-35). As a result, they landmarked a remarkable change of the older ideas of the tectonic evolution of Japan.

Figures 33 to 35 illustrate the overall geotectonic history of the Japanese Islands and five episodes of strong tectonic erosion during the last 500-600 Ma. At about 600 Ma, Rodinia continent started to be rifted and its continental fragments were dispersed over the world. In the center of Rodinia, the Paleo-Pacific Ocean was born, and successively enlarged in space through time. Along the margin of the South China craton, passive continental shelf deposits formed. From continental slope to oceanward, a very thick sedimentary pile was underlain by oceanic crust, which was a part of the Paleo-Pacific oceanic plate. The Paleo-Pacific Ocean started to shrink at about 520 Ma, when the passive margin turned into an active margin with subduction zone (Fig. 33a). But the boundary was not an oceanward boundary of continental crust. The boundary was inside the ocean simply because the boundary of continent is not straight, rather irregular, like in the case of the Indian continent. Therefore, proto-Japan started by the initiation of the subduction of the Paleo-Pacific oceanic plate and initially evolved as an intra oceanic island arc (Fig. 33b-d). This model suggests a proto Japan sea-like ocean was present between the intra-oceanic proto-Japan arc and the South China continent. By 200 Ma That island arc crust disappeared completely (Isozaki et al., 2010) (Fig. 33e).

Fragments of 500-400 Ma (U-Pb) granitoids and 360-340 Ma (K-Ar) HP-LT metamorphic rocks were tectonically eroded into the deep mantle, remaining less than several kilometers across blocks (Fig. 33b). A new intra-oceanic arc was active at 300-200 Ma to induce Pacific-type

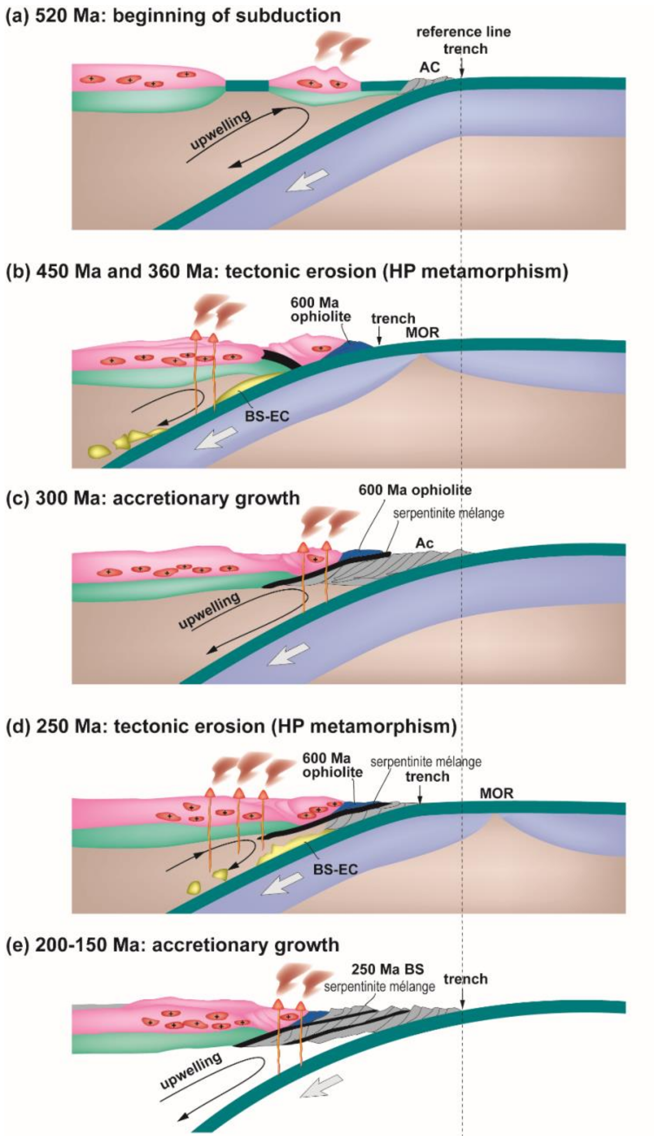


Fig. 33. Cambrian to Jurassic geotectonic evolution of the Japanese islands modified after Suzuki et al. (2010). (a) 520 Ma: beginning of subduction and TTG formation, (b) 450 Ma and 360 Ma: tectonic erosion, high-P/T regional metamorphism, and TTG formation, (c) 300 Ma: accretionary growth oceanward and CA volcano-plutonism, (d) 250 Ma: tectonic erosion, high-P/T metamorphism, and TTG formation, (e) 200—150 Ma: extensive accretionary growth and CA volcanism and plutonism.

orogeny with an accretionary complex growing oceanward (Fig. 33c). Another episode of tectonic erosion occurred at about 200 Ma and resulted in the shrink of proto-Japan (Fig. 33d).

The Triassic collision of the South China and North China cratons formed UHP to HP regional metamorphic belts, which protoliths were platform sediments. This event closed the proto-Japan Sea. A large Jurassic accretionary complex formed along the new plate boundary oceanward by the material supplied from the Triassic amalgamated continent of Asia (Fig. 33e). Formation of Jurassic accretionary complex was accompanied by the formation of extended volcano-plutonic belts on the Asian continental side and the oceanward migration of the trench, by which the Pacific-type orogenic belt expended oceanward to a distance of probably 100-200 km (in horizontal projection).

At about 130-120 Ma strong tectonic erosion started again and continued up to 80 Ma (Fig. 34a). During the Cretaceous, in a period from 120 to 80 Ma, a 200-300 km wide granitoid belt appeared parallel to the trench. Therefore, the continental crust expended far oceanward. The granitoids could be derived from oceanic slab MORB. However, the strong tectonic erosion on the trench side destroyed the older accretionary complex and the basement rocks, i.e., older granites, which all were transported to the mantle by the subducting slab. A part of them returned back to the surface after extensive recrystallization at mantle depths. Those events can be related to the subduction of the mid-oceanic ridge separating the Izanagi and Kula oceanic plates, which existed in the Paleo-Pacific Ocean that time (Fig. 8).

The age of accretionary complex at the bottom and top of a metamorphic belt is difference from that of the latter. For example, the accretionary complex overlying the Sanbagawa metamorphic belt in SW Japan formed at 180 Ma (Isozaki et al., 2010). The Shimanto accretionary complex below the Sanbagawa formed at ca. 80 Ma. Hence the age difference between these two accretionary complexes may reach 100 Ma. The Sanbagawa metamorphism peaked at 130-120 Ma (early Cretaceous), however the emplacement of subduction-related metamorphic rocks to the mid-crustal level was at 80 Ma, indicating a very slow exhumation rate (Okamoto et al., 2004).

In general, the time span between the peak of metamorphism and the exhumation to mid-crustal levels should be about 50 Ma (see Chapter, 2 sections 2.2, 2.4). The tectonic erosion of upper geologic units supplied the protoliths of the Shimanto accretionary complex to the trench and those were carried by the Pacific slab into the deep mantle. A part of the subducted rocks was metamorphosed at about 50-60 km and then exhumed (Figs. 34a, b, 35).

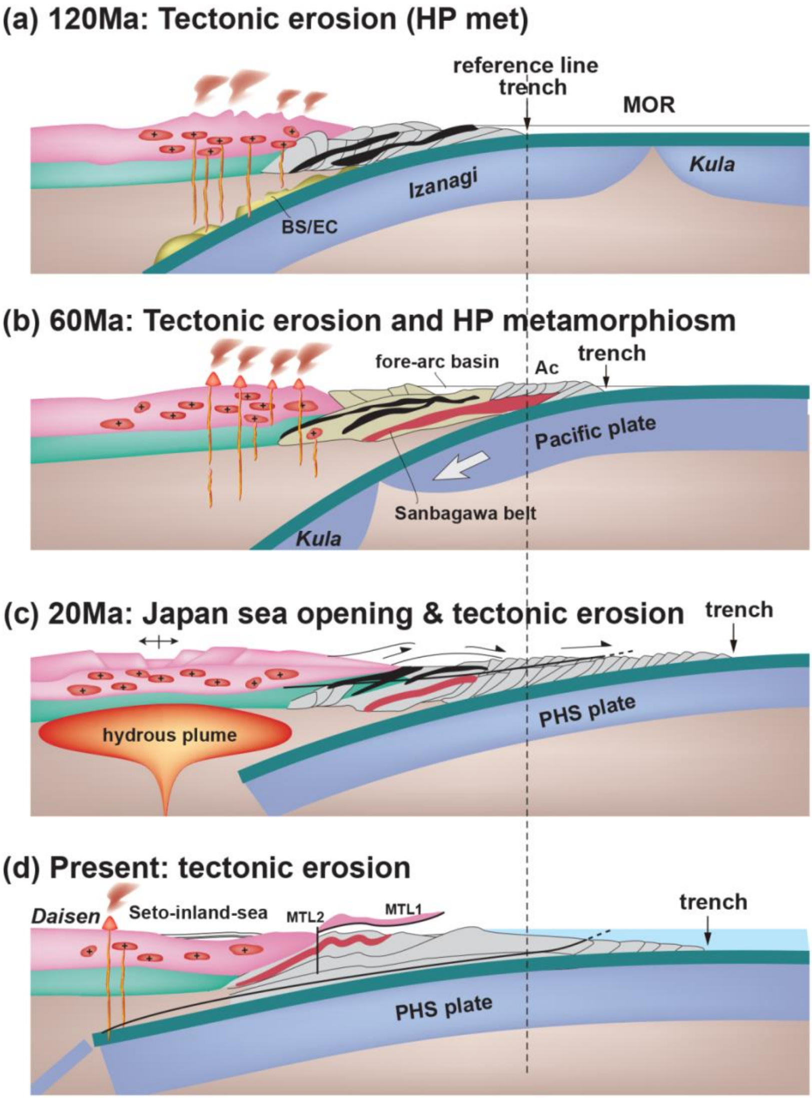


Fig. 34. Post-Jurassic geotectonic evolution of the Japanese islands modified after Suzuki et al. (2010) (continued after Fig. 32). (a) 120 Ma: tectonic erosion, Sanbagawa high-P/T metamorphism, and TTG formation with subsequent extensive growth of 80 Ma accretionary complex and simultaneous TTG growth, (b) 60 Ma: tectonic erosion, HP metamorphism, and TTG formation, (c) 20 Ma: Japan Sea opening, extensive tectonic erosion, and exhumation of high-P/T belt, and (d) present: tectonic erosion.

At the late stage of Pacific-type orogeny (Chapter 2, Figs. 7b, 9d, 10), which rather extensional, the regional metamorphic belt occupying the structural middle of a Pacific-type belt experience doming and accordingly acquire a dome-like radial structure consisting of gently folded antiforms and synforms formed around the dome. The secondary high-angle normal fault develops at the final stage; it cuts all three sandwiched units including the structural bottom of the non-metamorphosed or weekly metamorphosed accretionary complex.

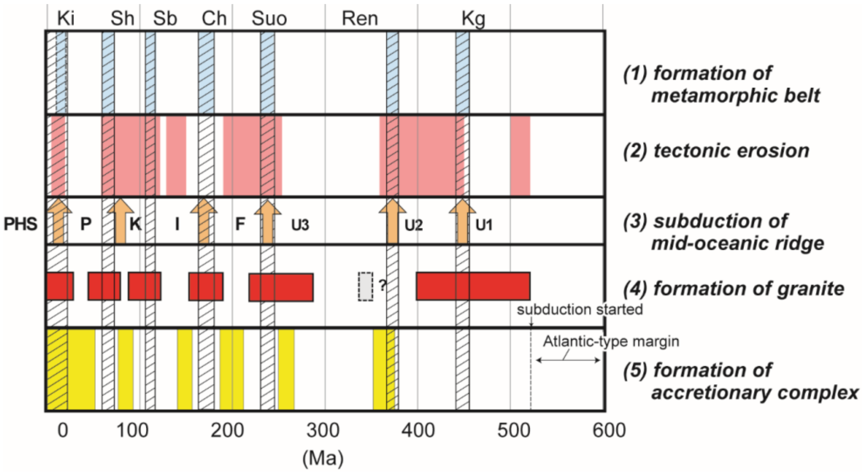


Fig. 35. Time-table of the last 600 Ma of the geotectonic evolution of the Japanese islands shows (1) formation of regional metamorphic belt, (2) tectonic erosion, (3) subduction of mid-oceanic ridge and names of plate,(4) formation of TTG, and (5) formation of accretionary complex. Metamorphic belts: Sb - Sambagawa, Ch – Chizu, Suo - Suo, Ren - Renge, Kg – Kurosegawa. Plates: PHS - Philippine Sea, P - Pacific, I - Izanagi, K - Kura, F: Farallon, U - unnamed.

In Miocene time, at about 25-15 Ma, the Sea Japan opened by the rifting at volcanic front, and the present-day Japan Islands formed (Fig. 34c). These events caused extensive tectonic erosion and destroyed the 150 km wide Cretaceous volcanic front in NE Japan. In SW Japan, beneath, the upper and lower boundaries of the Sanbagawa metamorphic belt at the brittle-ductile transition, i.e. at a depth of ca.15 km and 350oC, served as listric normal faults, by which the upper Cretaceous continental crust was thrust over the lower. The upper unit was eroded by weathering and the sedimentary material was supplied to the newly formed trench oceanward. That shortening exposed the Ryoke belt, which was a part of the Cretaceous Sanbagawa belt, along the MTL (median tectonic line; Fig. 3). The strong NW shortening uplifted the continental crust and triggered its erosion to form a very thick and wide Miocene accretionary complex of the Shimanto belt. At the same time, a part of the crust was probably removed together with the descending slab. The eastern Hokkaido arc collided with the western Hokkaido arc and they together collided with the Honshu arc. The collision consumed the intervening ocean at about 0.5 Ma. In Cretaceous time, strong tectonic erosion also occurred west of Hokkaido to form a serpentinite mélange belt (Ueda et al., 2000, 2010). In Shikoku (Fig. 3), the 100 km wide Sanin granitic batholith belt formed at about 60 Ma, coevally with the formation and exhumation of Shimanto blueschists. This could be related to the subduction of the mid-oceanic ridge separating the Kula and Pacific oceanic plates (Maruyama et al., 1997).

In the east, the proto Izu arc moved northward to collide with the Honshu arc at about 15 Ma to shorten the Honshu arc to about 50%. Therefore, the 50% of the Honshu arc was destroyed and that crustal material was transported into the mantle. Moreover, during the Eocene to Miocene a proto Izu-Mariana intra-oceanic island arc was also tectonically eroded and partly disappeared. Those tectonic events moved the position of the volcanic front and trench to about 150 km to the west.

Recent tomographic images by Akira Hasegawa from the Tohoku University show the Japan Trench at a depth of 15-10 km beneath Sendai (Hasegawa, Yoshida, 2015). The tomographic images suggest the presence of serpentinite protrusions moving to the fore arc or trench inner wall, but not yet on the surface of the ocean floor (Maruyama, 2009, 2010). South of the Boso Peninsula and east of Tokyo, a series of serpentinite conical seamounts is present on the landward slope of the Izu-Ogasawara-Mariana trench, about 50 km west of the trench axis, landing from north to east, parallel, in places, to the trench. The inner wall of the trench consists of Tertiary volcanic rocks of boninitic to normal calc-alkaline series, gabbro, serpentinite, and peridotite, plus glaucophane schists suggesting the presence of deep-seated regional metamorphic belts underneath. A LT-LP accretionary complex was drilled or dredged on the surface (Hirata et al., 2010). In short, those serpentinite belts host rocks formed at different PT-conditions. As we argued above, the modern occurrence of serpentinite and its trapped petrologically diverse rocks could be due to tectonic erosion. The Mariana fore-arc shelf is cut by the Japan Trench in the north, therefore, the Tertiary arc crust was tectonic eroded and probably removed into the deep mantle (Hirata et al., 2010).

More evidence for the multiple events of tectonic erosion in the Japanese Islands comes from the U-Pb age spectra of detrial zircons from mid-Paleozoic to Mesozoic sandstones and recent river sands in Japan, showing secular change in provenances that supplied terrigenous clastics (Fig. 36). Note the 3 distinct stages in the over-500 million-year history of the Japanese Islands in terms of terrigenous clastics from granitic sources; i.e. before the Late Triassic (ca. 200 Ma), Jurassic to mid-Cretaceous (ca. 200-90 Ma), and after the Late Cretaceous (ca. 90 Ma). This suggests that a major provenance change occurred twice in the vicinity around Proto-Japan and the Japanese Islands. During the Paleozoic to Triassic, major source was Paleozoic arc granite that developed in the juvenile Japan that probably formed an intra-oceanic arc (proto-Japan arc complexes). Despite the physiological proximity, the terrigenous flux from the two major continental blocks (South and North China) was highly limited. The major re-organization in provenance regime around 200 Ma occurred in relation to the collision between South China and North China that caused the abundant production of terrigenous clastics and delivery to the arc-trench system along the Circum-Pacific. The second change around 90 Ma was likely induced by the establishment of the huge Cretaceous batholith belt in the Japan arc that formed a great barrier wall in front of the two old continental cratons to block the terrigenous flux (Isozaki et al., 2010).

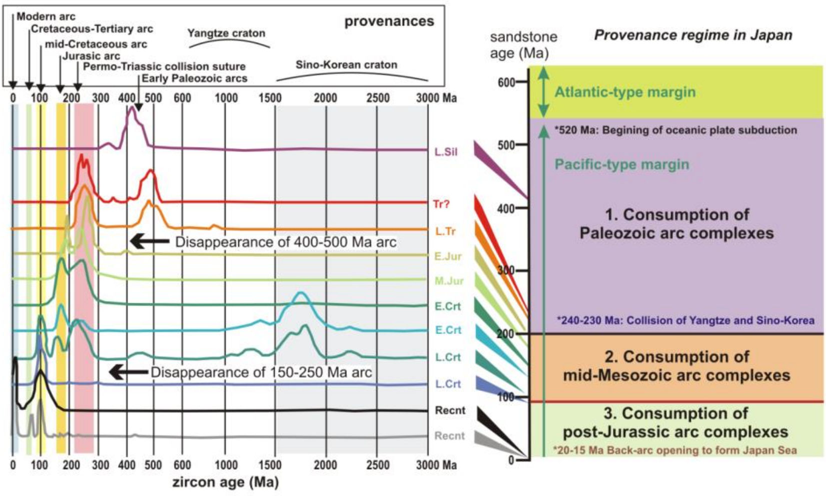


Fig. 36. The timing of tectonic erosion is estimated by the U-Pb age frequency diagram of detrital zircons in sandstones formed from 400 Ma to the present sands at a river mouth (Nakama et al., 2010b). Two periods at ca. 200 Ma and 100 Ma are the timings of extensive TTG crust erosion.

In addition to the more expected examples from the Circum-Pacific, let's consider older examples from both sides of the Atlantic Ocean, in particular, the British Isles (Fig. 37). The British Isles fundamentally comprise a northern Laurentian fragment in Scotland and a southern Avalonia fragment in England and Wales (Fig. 37a). England—Wales was formed through a Pacific-type orogeny during the Ediacaran—Cambrian, whereas Scotland was rifted away from Laurentia at ca. 800—500 Ma (Fig. 37b, c); the resulting continental margin was covered by platform sediments, subducted under an intra-oceanic arc, and returned to the surface at 470 Ma (Fig. 37d). Directly after 470 Ma, Scottish Dalradian was subjected to Pacific-type orogeny by the northward subduction of the oceanic plate from the Iapetus Ocean (Fig. 37e). Avalonia was separated from West Africa at 480 Ma. Before and after separation (back to 700 Ma), England—Wales was underlain by the Pacific-type orogeny and voluminous TTG magmas intruded. Avalonia migrated northwards to collide with Scotland by 440 Ma, consuming a few intra-oceanic island arcs on the way (Fig. 37d, e). The Iapetus Ocean was closed by the Avalonia collision, whereas the Rheic Ocean was at its maximum width when the Iapetus Ocean was closed at 440 Ma, but it closed again when Africa collided to form the Appalachian orogen at 400 Ma (Windley, 1995). Moreover, at the end of Paleozoic, formation of Gondwana supercontinent occurred through orogeny. First, Baltica-northern European continent collided with Laurentia-north American continent and to form Scandinavian orogenic belt in early-middle Paleozoic, then west Africa collided against North America to form Appalachian orogenic belt during the middle Paleozoic. As the result, apparent supercontinent Gondwana was formed. From Great Britain, to Newfoundland, the eastern part of Baltica is missing of collided continents, therefore, in these regions, North American continents subducted underneath intra-oceanic arc, namely arc-continent collision orogeny occurred, and there after Avalonian small continental blocks collided, originally derived from west African craton, collided against North American continent to form precursor of United Kingdom. It was about 420Ma (Fig. 37f).

4.6 A summary

The birth of the precursor of British Isles proceeded by the collision of intra-oceanic arcs, which delivered TTG material into the subduction zone at about 600-500 Ma. Evidence for this comes from the Appalachian foldbelt and northern Africa foldbelts aged at 700-500 Ma. The Pacific-type orogeny along the western margin of North America started at about 500 Ma. The mirror subduction on the opposite side of the Pacific Ocean, at the eastern margin of pro-Asia continental blocks, started also at 500 Ma to form similar Pacific-type orogenic belts, a precursor of the Japanese Islands. Along the western coast of North America, several episodes of Pacific-type orogeny caused extensive tectonic erosion to deliver TTG material into the underlying mantle. The orogenic belts extend from western Canada to Alaska, where the Triassic-Jurassic continental break-up occurred to create a tectonic situation similar to that in the present-day south-western Pacific region, which is characterized by several micro-plates with continental basements.

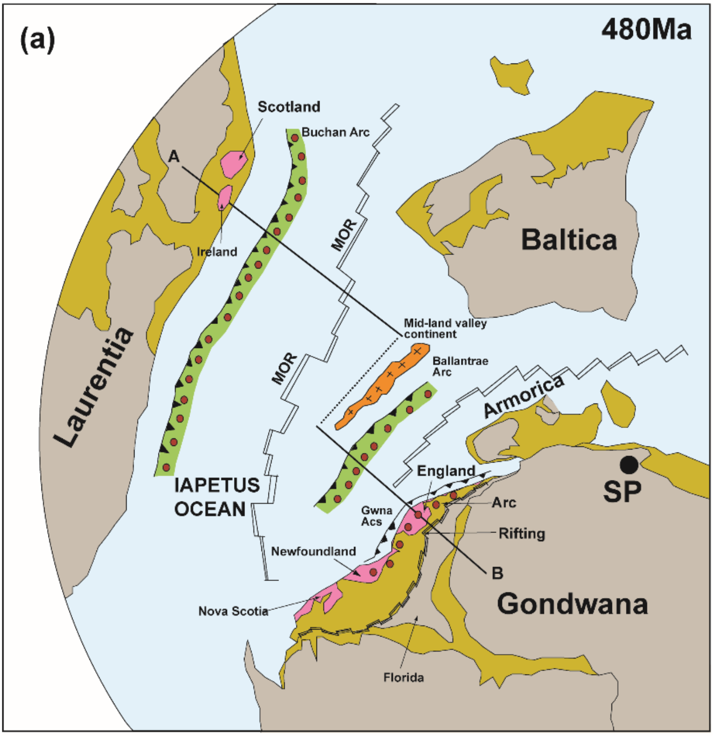


Fig. 37. Geotectonic development of the British Isles: a paleo-geographic reconstruction for the Early Ordovician (a) and time-series cross-section along line A-B (b-f). Comments for (b-f) see in the text.

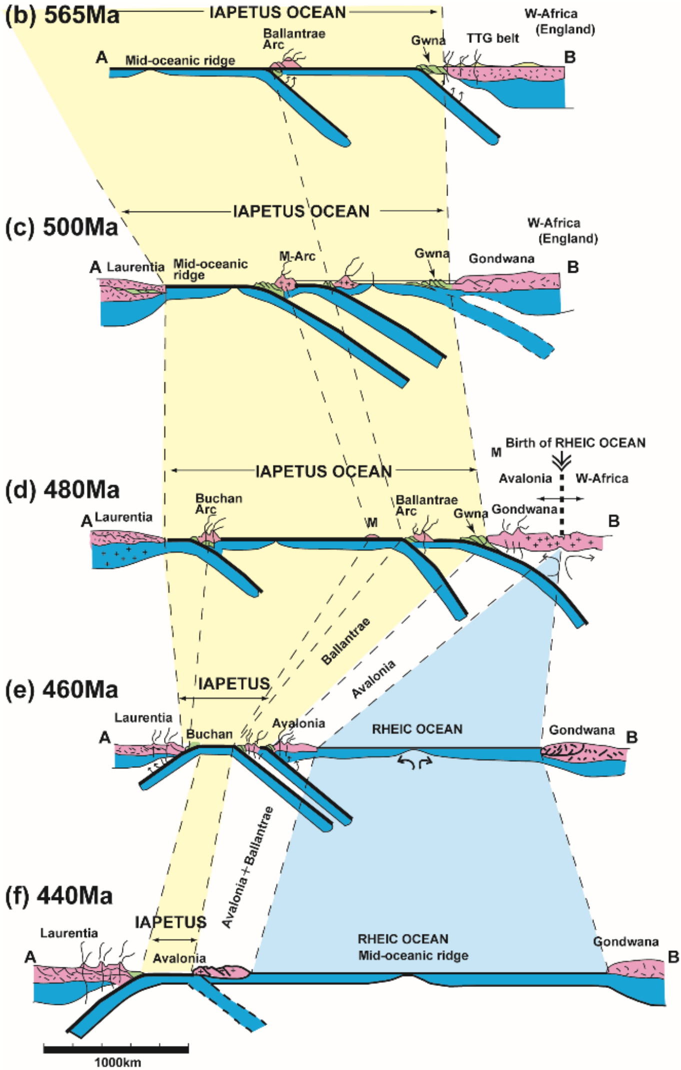


Fig. 37. Continued.

A key point of Pacific-type orogeny is the balance between the rapid collapse of continental crust and its subsequent rapid growth. IT is very convenient to use the Japanese Islands as an example because this region has been very carefully studied by hundreds of Japanese geologists and their cooperative partners from many countries in Asia, Europe, and America during the last 40 years. The Japanese Islands were born at around 600 Ma and since then have experienced at least five episodes of Pacific-type orogeny, which made the crust of Japan grow oceanward and also five episodes of strong tectonic erosion (Figs. 33-36). Every time retreat and advanced ocean ward or continent ward, just repetition (Figs. 33, 34). During the last 520 Ma the continental crust of the Japanese Islands increased five times compared to that of today. However, the tectonic erosion caused the removal of four times of the mass of continental crust to the mantle to leave only one landmass of the present-day Japanese Islands. The question is what was the cause for such strong tectonic erosion what was the cause of large-scale continental crust growth in short time? An answer comes from the well-constrained history of Asian paleo-geography during the last 200 Ma in respect to the reconstruction of oceanic plates. Of key importance are the detailed geology of the Japanese Islands, in particular, during that period, the formation of accretionary complex, the episodes of tectonic erosion and the formation of regional metamorphic belts. A lot of information about those events is stored by sandstones allowing reconstruction of their hinterlands keeping material of former TTG belts. The frequency distribution curves of U-Pb zircon ages should be compared with those of hinterlands, both from sandstones and TTG, to understand which arcs/belts completely disappeared in the past (Fig. 36). The comparison of paleo-geographic reconstructions and U-Pb age frequency distribution curves of zircons separated from sandstones led to the following 7 conclusions.

(1) Well correlated timings of progressive regional metamorphism and tectonic erosion. This means the transportation of tectonically eroded sedimentary rocks to the deep levels of subduction zones. The formation of accretionary complex in the same period indicates that a part of trench deposits remained in the surface. The sediments overlapping the oceanic slab may, of course, be brought into the deep mantle together with the slab, suffer progressive metamorphism and never return back to the surface. But the amount of sediments transported into subduction zones must be huge at certain conditions (see sections 4.1, 4.5). Those sediments might stop at 50-60 km depths and get removed through their underplating to the hanging wall. After that they must return to the surface in a direction opposite to that of the descending slab. This way, although a bit enigmatic, is necessary to generate a deep-seated subduction-related metamorphic complex (Chapter 2, Fig. 9).

(2) Formation of granitic rocks is related to ridge subduction into trench (Fig. 34b). This idea strongly explains why slab melting is a major reason to form orogenic granites. The timing of ridge subduction roughly

corresponds to the ascent of a deep subduction-related metamorphic complex to mid-crust levels. The pressure from approaching mid-oceanic ridge shallows the angle of subduction thus creating a major driving force of the exhumation of HP metamorphic rocks, first of all, blueschists and eclogites.

(3) Formation of accretionary complex starts when mid-oceanic ridge get subducted. Ridge subduction fixes a regional metamorphic belt at mid-crustal levels and it jack-ups to form a folded mountain belt, i.e. orogen. The timing of that orogeny corresponds to the formation of huge volumes of accretionary complexes (Fig. 33e, 34d).

(4) Based on 1-2 we can propose the following holistic scenario linking tectonic erosion, formation of regional metamorphic belt, batholith belt, accretionary complex and subduction of mid-oceanic ridge. When mid-oceanic ridge approaches trench, strong tectonic erosion causes subduction of sediments to a depth of about 50-60 km beneath the hanging wall of the mantle wedge together with descending slab. At that depth, those sediments must recrystallize under a HP-LT geothermal gradient and start to extrude from underneath the mantle wedge by the shallowing subduction angle because of the approaching mid-oceanic ridge. Those recrystallized regional metamorphic belts are still ductile (because the temperature above the brittle-ductile transition is 350oC) and therefore they can move a direction opposite to that of the descending slab at a rate of 2-3 mm/year, which is very slow. After the ridge subduction, the 2 km thick flow of the ductile channel moves to the surface, the angle of subduction becomes steeper and accretionary complex can form. That thick accretionary complex can be underplated at mid-crustal depths beneath the exhumed regional metamorphic belts and push it up to induce orogeny of formation of a non-volcanic arc (Figs. 7b, 9d, 13c).

(5) Considering those issues we can predict orogenic episodes older than 200 Ma, because the timing of mid-oceanic ridge subduction corresponds to the progressive subduction-related metamorphism. If the first episode of strong tectonic erosion happened at 250-200 Ma, then the next older tectonic erosion must be at ca. 350-300 Ma (Fig. 33). These two ages suggest that the mid-oceanic ridge subducted underneath proto-Japan.

(6) The above mentioned empirical rules do not look consistent with the ubiquitous tectonic erosion along the Pacific trenches. If issues 1-5 are correct and approaching mid-oceanic ridge defines strong tectonic erosion and formation of accretionary complex after ridge subduction, then marine geophysicists should consider more detailed observations in the regions of ongoing tectonic erosion or stopped tectonic erosion.

(7) Dewey and Bird (1970) proposed the model of Cordilleran-type orogeny and Pacific-type orogeny (Fig. 38a), but the latter must be sufficiently improved based on the results of very detailed geological studies of the Japanese Islands (Figs. 33-36).

S. Maruyama (1997) first described regional metamorphic belt as less than 2 km thick platy high-temperature intrusion into weakly metamorphosed or non-metamorphosed accretionary units, which represents a core of orogeny together with its other constituents - granitoid batholith belt and fore-arc basin in between. Accretionary complex represents a structural bottom of P-type orogeny. All those four units define what we call Pacific-type orogenic belts (Fig. 1). In addition to those four, ubiquitous tectonic erosion plays a crucial role in the formation of regional metamorphic belts. Evidence for strong tectonic erosion comes from the present and past occurrences of serpentinite mélange belts. Moreover, huge amounts of fragmented and eroded continental crust material are transported into the mantle, which represents one of the most important discoveries.

In the Japanese islands, the oldest orogenic belts now rest on the top of younger Pacific-type orogenic belts. The occurrences of all Paleozoic Pacific-type orogens are fragmental and the total amount of those is extremely small, but still they are present. Tectonic erosion was responsible for the present-day fragmental occurrence of P-type belts, which all once consisted of four components (granitoids belt – fore-arc basin - metamorphic belt – accretionary complex), forming large orogens, similarly to the modern orogenic belts. **The presence of even small fragments of those rock associations or constituents, not necessarily all four, clearly indicate a Pacific-type nature of orogen, which obviously was originally much larger than now.**

A classic model proposed by Dewey and Bird (1970) (Fig. 38a) shows extensive formation of TTG batholith and contemporaneous mélange formation at the trench with the mixing of oceanic materials with trench turbidite. However, those units are emphasized without any concept for the formation of one orogen called Pacific-type with four elements, exhumation of HP belt as an orogenic core, fore-arc basin deposit, accretionary complex underneath, and a huge TTG belt (Fig. 38b). One unit of Pacific-type orogen corresponds to one ridge subduction (Maruyama, 1997). The revised concept of Pacific-type orogeny includes the concept of tectonic erosion. The rapid growth of TTG crust, created by slab melting when a young slab subducts, is accompanied by tectonic extrusion of a high-P/T regional metamorphic belt and followed by extensive tectonic erosion every 100 m.y. As a result, the continental margin never grows.

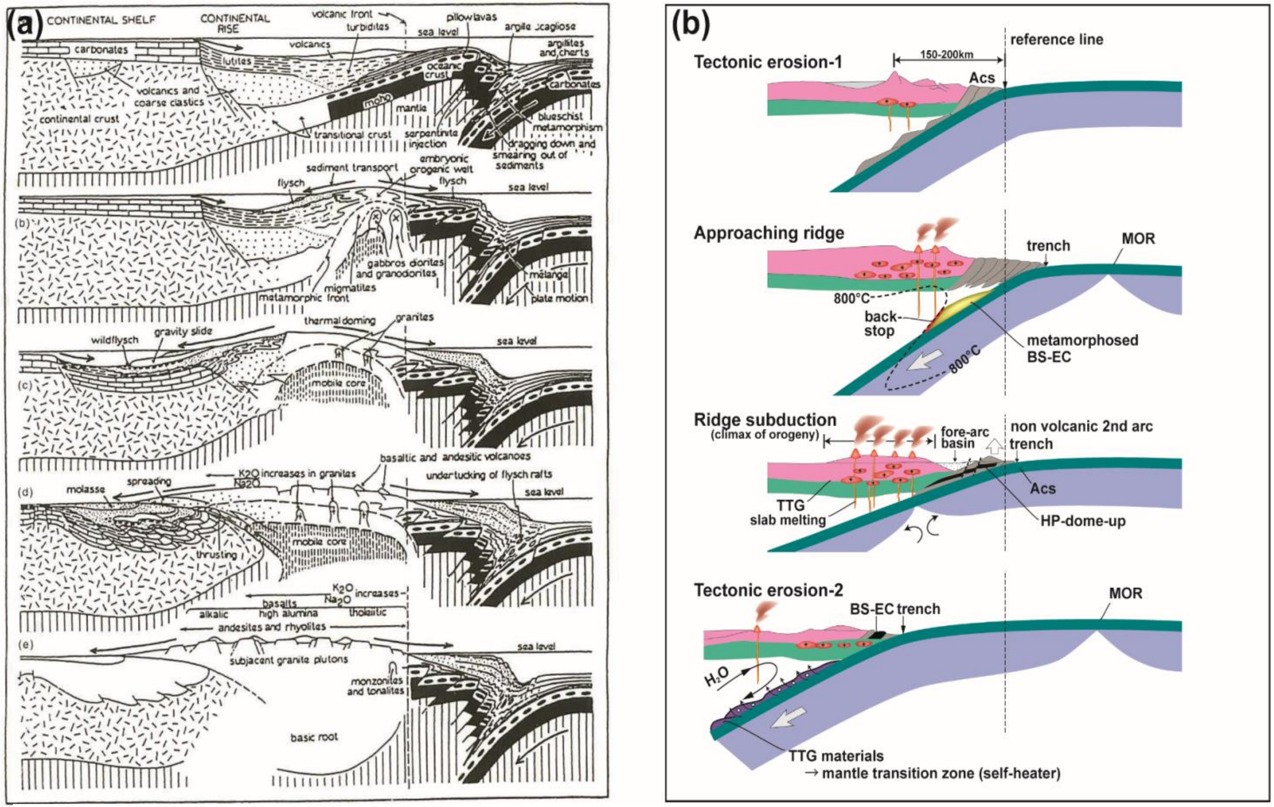


Fig. 38. A classic model proposed by Dewey and Bird (1970) shown in (a) and a revised model of Pacific-type orogeny (b).

4.7 From Pacific-type to collision-type orogeny plus tectonic erosion

The proto-Izu intra-oceanic arc collided with Honshu arc of Japan in Miocene time (see above section 4.5 and Fig. 3). That collision caused string tectonic erosion and changed the general pattern of regional geology, which was partly destroyed. The Izu arc must have subducted directly underneath the Honshu arc to nearly 100% and now the subduction is ongoing with the descending slab of the Philippine Sea plate into the mantle. So, the subduction of the arc leaves not a piece of island-arc crust on the hanging wall. If a descending slab consisted of continental crust, like in the case of the India-Eurasia collision, most of geologists believe that accretion of continental crust must be 100% or that continent cannot subduct into the mantle. Is that true? This is a question. More than 20 intra-oceanic island arcs, about 90% of the total number of those on the Earth, are now in the modern western Pacific region, which occupies only 7% of the total globe surface (Juteau and Maury, 1999). Most of the SW Pacific arcs could be destroyed by the northward moving Australian continent. In about 50 m.y. from now they may become a part of the Amasia supercontinent (Maruyama et al., 1989; Senshu et al., 2010; Safonova and Maruyama, 2014). However, most of the intra-oceanic arcs in the western Pacific hardly can be accreted to the hanging wall of the Asian continental margin. Almost 100% of those arcs will probably disappear through subduction into the deep mantle.

Let us discuss the fate of intra-oceanic arcs in terms of the India-Eurasia collision. India collided with Asia at ca. 50 Ma (Hall et al., 2002). But that time the Himalayan orogeny did not yet start. An intra-oceanic arc, Kohistan, was present between the Indian and Eurasian continents before their final collision at 50 Ma. First, the Cretaceous Kohistan intra-oceanic arc collided with the Indian continent, remaining a Tethyan sea between Eurasia and the arc. The presence of the Kohistan arc is constrained by the amphibolite crust underneath, the absence of surface sedimentary cover therein, and by calc-alkaline igneous rocks. On the surface, the Kohistan arc has survived as small fragments only, one mountain size, with a thickness of probably less than 1 km. Therefore, about 99% percent of the Kohistan arc must have subducted to the deep mantle. Pacific-type orogenic belts and several microcontinents are present to the north of the Himalaya orogenic belt. There are 100 to 200 km wide Cretaceous batholith belts paired with a narrow parallel glaucophane schist belt of close age. That region is very narrow, like a suture, as it is characterized by serpentinite and glaucophane schist belts, but no jadeite and quartz. In between the batholith and blueschist belts there is a Cretaceous fore-arc basin (Maruyama et al, 1989). Pacific-type orogenic belts are ubiquitous in Tethyan orogenic belts. Therefore, it is reasonable to consider collisional belts as formed by two large continents, once separated by a large ocean. But to consume such an intervening ocean, Pacific-type orogenic belt must be present. In addition to those Cretaceous intra-oceanic island arc and, the large Indian continent, there are small continental blocks and the Kohistan intra-oceanic arc. Both, the arcs and the microcontinents, originally, could be much larger. Those suggestions have been long debated. For example, the concept of Greater India suggests that the size of the Indian continent was twice as large before the collision. The northern half of the Indian continent may have subducted underneath the Himalayan mountain belt. A part of it may still be present under the Tibetan region. The upper continental crust could be a part of Asia, the lower continental crust could be the Greater India, and other pieces of the plate descending northward have been already subducted into the mantle below Asia. More evidence for such a scenario comes from P-wave tomographic images (Huang and Zhao, 2006).

Figure 39 presents high-resolution P-wave tomographic images underneath the Himalayan mountain belt with two north-south cross sections shown by lines. At 50 Ma the original trench was located south of India, near the southern end of the Indian Peninsula. A huge high P-wave velocity anomaly is present underneath India (Greater India I), suggesting stagnant slabs including TTG crusts within it. Since then, the Indian continent indented into Asia to ca. 2000 km north to the present-day suture at the foothills of the Lesser Himalayas. The TTG crust and the underlying tectospheric mantle can be traced under the Tibet (Greater India II). The isolated high P-wave velocity anomaly under the Tarim may correspond to the TTG crust-bearing slabs (Greater India III). There are also low- P-wave velocity and high P-wave velocity anomaly masses, one in the mantle transition zone and another in the uppermost lower mantle, none of which can be traced upward and downward. This suggests the presence of older TTG crusts (Proto-Greater India) in stagnant slabs (bottom of Fig. 39). A similar interpretation is available along line (h). The lowermost boundary could be the subducted Indian continent extended north of the Himalayan mountain belt. Further to the north, a high p-wave velocity could be the subducted continental crust of the Greater India.

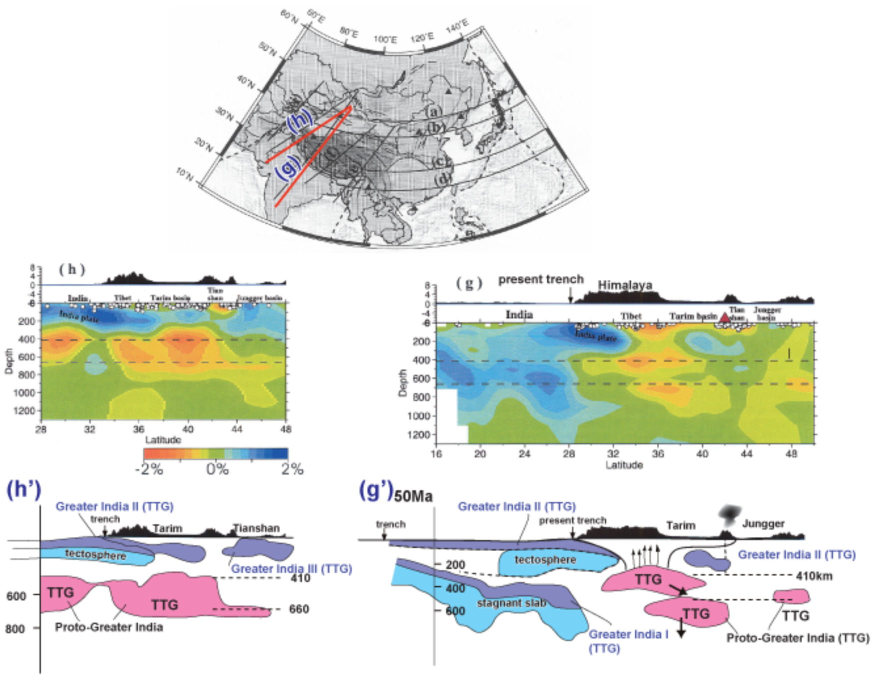


Fig. 39. P-wave mantle tomography in East Asia (Huang and Zhao, 2006) and an interpretation of subducted TTG crusts in the mantle (bottom figures).

Chapter 5. Fate of TTG, role of water and second continents in the MTZ

5.1 How granite can be subducted?

The new view of subduction zone tectonics shows that the formation of accretionary complex parallel to subduction is not ubiquitous. The hanging wall of P-type convergent margin can be destroyed by tectonic erosion. Fragments of continental crust can be transported into the deep mantle together with the slab. If this is correct, how deep granitoids can be transported into the mantle? It has been long believed that the continental crust, which is dominated by relatively light granitic rocks, cannot be subducted to depths greater 30 km (Moho boundary) because of its high buoyancy, i.e. lower density (2.8 g/cm3), in respect to the mantle (3.5 g/cm3). A possibility for low-density materials such as upper continental crust to sink into the deep mantle was pointed out by many scientists (e.g., Irifune et al., 1994; Dobrzhinetskaya et al., 2006; Afonso and Zlotnik, 2011; Kawai et al., 2013; Ichikawa et al., 2013b).

Experimental and theoretical studies showed that the density of continental materials such as Archean TTG is higher than the mantle peridotite in between 270 and 800 km in depth (Irifune et al. 1994; Kawai et al. 2009, 2013; Kawai et al., 2010; Ichikawa et al., 2013a; Kawai and Tsuchiya, 2015). Based on the first principle calculations Kawai with co-authors (Kawai et al., 2009, 2013) estimated the densities of TTG crust, pyrolite, harzburgite, MORB and anorthosite and concluded that beneath a depth of 270 km the rocks of granitoid composition acquire negative buoyancy compared to the ambient mantle due to the coesite-stishovite, olivine-wadsleyite and wadsleyite-ringwoodite transformations at depths of 270, 410 and 520 km, respectively (Fig. 40). Those phase transitions result in higher densities of the TTG-type rocks allowing them to be subducted to the depths from 270 to 660 km, but not deeper into the lower mantle (Kawai et al., 2013). Using Preliminary Reference Earth Model (PREM; Dziewonski and Anderson, 1981) for the MTZ Kawai with co-authors (2009) compared the estimated P- and S-wave velocities of TTG with those of peridotite (pyrolite, harzburgite) and MORB to show that only the TTG fits the PREM model at 270 and 660 km). The tomographic images beneath Japan and East Asia show the active subduction of the Pacific plate and accumulation of the slab materials between the 410 and 660 discontinuities. The low-P-wave zones under the Datong and Changbai modern volcanoes suggest mantle melting near the upper boundary of the MTZ, i.e. at a depth of 410 km (Huang and Zhao, 2006).

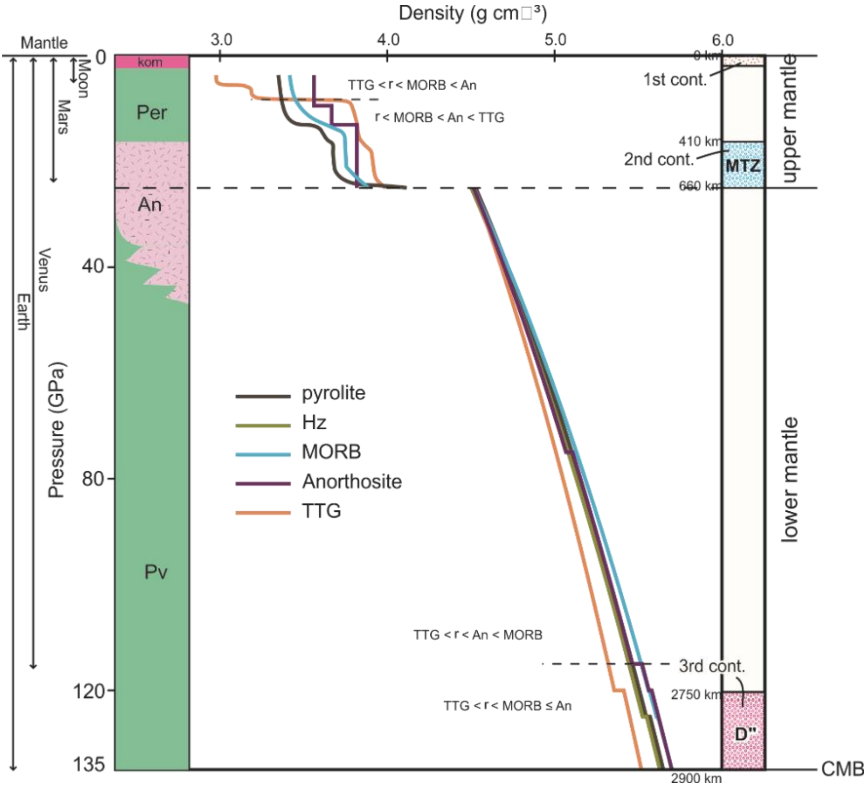


Fig. 40. Density profiles of TTG, anorthosite, MORB, harzburgite, and pyrolite at a whole range of mantle depths (after Komabayashi et al., 2009; Kawai et al., 2009, 2010). Per, peridotite; An, anorthosite; Pv, perovskite, TTG, tonalite-trondhjemite-granodiorite; MORB, mid-ocean ridge basalt.

TTG is the most buoyant rocks near the surface, but the heaviest between depths of 300 km and 660 km, suggesting the strongest slab-pull force is in the mantle transition zone, if it is subductable below 300 km. However, it is most buoyant across the whole depth range in the lower mantle, although the density difference in the uppermost lower mantle is negligibly small. This suggests that the TTG mass accumulates at the bottom of the upper mantle to form a second continent. Anorthosite follows a similar trend as TTG, but it becomes the heaviest material throughout the depth of the lower mantle, suggesting that an anorthosite layer covering the Hadean Earth must have been deposited on the core-mantle boundary (CMB) at a depth of 2900 km by mantle convection after the Hadean. Metamorphosed anorthosite may appear a major component in the D" layer at the bottom of the mantle (Fig. 40).

A question is how to deliver the crustal material to the depth of 270 km after which it can sink down further to the MTZ (Fig. 40). The continental material, in spite of its buoyancy, can be subducted through subduction channels developing at the interfaces between the subducting (oceanic) and overriding plates (Von Huene and Scholl, 1991; Yamamoto et al., 2009; Stern, 2011). Recent finite difference method based numerical simulations of a subduction channel have shown that such a subduction of continental material may proceed by a mechanism of "viscous drag" (Fig. 41) implying a dragging force greater that the buoyant force (Ichikawa et al., 2013b). Those authors show that a 2-3 km thick channel can provide a flux of continental materials of 2.2 km3/yr and suggest that almost all of the continental material that is subducted through the channel is capable of reaching the MTZ.

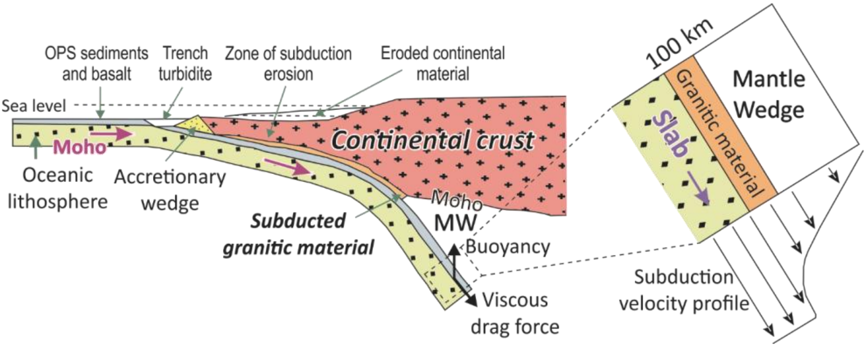


Fig. 41. "Viscous drag" model for the subduction of buoyant granitic material (modified from Ichikawa et al., 2013a). The eroded granitic material is subducted via a subduction channel located between the slab and the mantle wedge. At a 2000–3000 m thickness of the subduction channel, the slab is able to drag the continental granitic materials of an effective thickness of about 500 m. OPS, ocean plate stratigraphy; MW, mantle wedge. For details of the numerical simulation, see Ichikawa et al. (2013a).

Thus estimated amount of the continental material, which have been transported to the deep mantle during the 4 Ga of Earth's history, is about 1010 km^3, i.e. greater than the volume of the present continental crust (Ichikawa et al., 2013b). A recently performed systematic series of three-dimensional numerical simulations examining the effects of viscosity behaviors near the base of the MTZ has shown that the viscosity reduction of subducted crustal material leads to a separation of crustal material from the slab and its stagnation in the MTZ (Tajima et al., 2015). Therefore, we argue that the previous opinion about the impossibility to subduct granitic material deep to the mantle is not absolutely true anymore (Safonova et al., 2015).

5.2 The role of water

Old oceanic plates must be highly water-saturated from the top (sediments) to partly hydrated slab peridotites. Subduction of an old plate means that the slab must be rather cold and that the hydrous Mg-silicates of slab peridotites may remain stable inside the subducting slab, which then can transport the surface sea water into the deep mantle. Deep seismicity in the subducting slab can be explained by the reactions of dehydration of those hydrous Mg-silicates or dense hydrous Mg-silicates (DHMS) by the heating of the descending slab by the ambient mantle to generate free fluid through dehydration embrittlement of DHMS. That process has been studied by experiments in diamond anvil by R. Jeanloz (UC Berkley, USA), who confirmed the sound of dehydration embrittlement indicating that the dehydration of serpentine causes the birth of fluid together with the sound, i.e. brittle fragmentation. The dehydration causes sudden drop of the temperature of partial melting and of the viscosity by the birth of fluid. Therefore, subduction zones, even in a low-temperature state, can generate magma under high water pressure to activate/promote convection in the mantle wedge.

The Western Pacific triangular region (Fig. 42a) is the place where the Pacific plate is subducting from east to west and the Indo-Australian plate is subducting from south to north, i.e. double-side subduction. Both plates are very old and can be traced back to 200 Ma (Jurassic time). The double-sided subduction zone in the Western Pacific region represents an entrance of water into the deep mantle. The western Pacific region has been refrigerated by the subducting cold oceanic plates since 450 Ma.

However, the region is also characterized by the presence of many oceanic microplates less than 1300 km across, as well as active magmatism. Petrochemical data from drilled basalts of DSDP from the Philippine Sea plate show that the source mantle for oceanic basalts is rich in water ca. 0.2 wt.%, and is 50–60 °C lower than that for MORB. The extensive melting is due to the high water content in the source mantle. Marginal basins may be greatly deeper than major oceans. Age–depth calculations based on a model of transient half-space cooling at given parameters of temperatures of mantle and surface, 1280 and 0 °C, and the thermal diffusivity, 1 mm2/s shows that the correlation of age-residual depth from a mid-oceanic ridge is consistent with the bathymetric data (Park et al.,1990; Komiya and Maruyama, 2007). Moreover, the mid-oceanic ridge may be relatively deep because this region is underlain by the cooler mantle. Addition of water to the mantle peridotite lowers the solidus temperature and viscosity. Melting experiments of hydrous peridotite show that addition of 0.2 wt.% H2O content lowers the solidus temperature by 150 °C. As a result, the

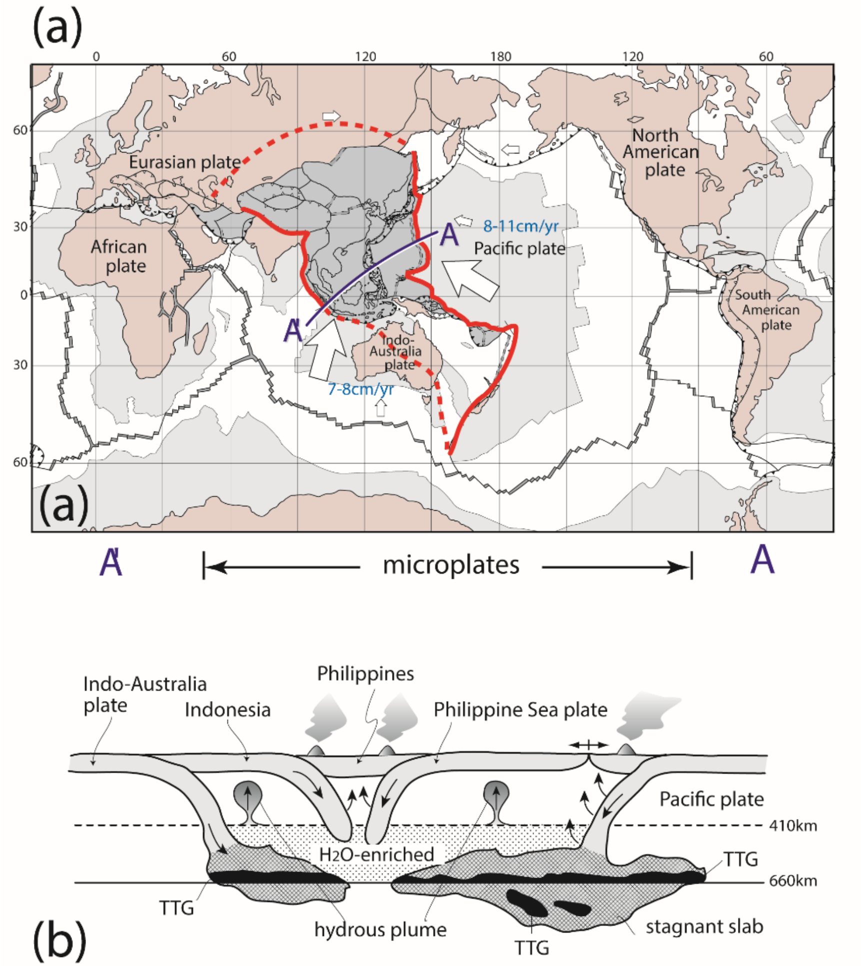


Fig. 42. The Western Pacific triangular zone (a) and the structure of the upper mantle underneath (b).

mantle under the region may practically correspond to a ca. 90 °C hotter mantle than normal MORB-source mantle in terms of magmatism and rheology. Numerical simulation for a hotter mantle suggests that many small plates should be formed because of extensive heat release by active magmatism, consistent with many microplates in this region.

This is the reason why the western Pacific mantle is a region of the most active mantle convection causing frequent events of seismicity (earthquakes) and volcanism (eruptions). The deformations on the orogenic zones of the western Pacific are most active and ongoing. The continental crust in eastern Asia is fragmented to consist of more than 10 continental microplates. There are also fragments of more than 10 microplates in the western Pacific oceanic region and they are characterized by different types of plate margins (divergent, convergent, transform). This explains well why 90% of the world intra-oceanic island arcs and marginal basins are concentrated in this small region (7% of the area of the Earth), where more than 20 marginal basin and island arcs are present (Fig. 42a) (see also Chapter 4, sections 4.6, 4.7).

Thus, the Western Pacific triangular zone faces the presence of double-sided subduction zones caused by the Pacific subduction (Fig. 42a) from the east and Indo-Australian plate subduction from the south. It marks the entrance for water draining into the mantle (Fig. 42b), i.e. represents the entrance for TTG subduction caused by tectonic erosion at P-type convergent margins or direct arc subduction to form then a second continent in the MTZ. All of these cause the formation of a supercontinent by the double-sided subduction zones and its breakup afterward by melting temperature drop and viscosity drop to activate mantle convection by the entrance for water and the heating of the lowered-T sub-continental mantle by the entrance of TTG.

5.3 Second continent in the mantle transition zone

However, even the most remarkable orogenic activity cannot be explained only by water. As follows from the previous section the western Pacific triangular region has been cooled down by double seismic subduction zones during more than 100 m.y. at least. At an initial stage magmas can be generated by the lowering of melting temperature by the addition of water. However, it cannot be continuous over a long time. The further cooling and extraction of magma components should stop magmatism, but the magmatism and seismicity in the western Pacific triangular region are continuous and ongoing. To solve the problem we need an additional heat source, i.e. the presence of second TTG continent at the bottom of the upper mantle. The focus of this paper is tectonic erosion, which is rather ubiquitous at Pacific-type convergent margins and related subduction zones (see Chapter 4). The second issue is direct subduction of at least five intra-oceanic arcs to the deep mantle (Yamamoto et al., 2009). These two observations from the western Pacific indicate that huge amounts of granitic materials including intra- oceanic arcs must have been subducted during several hundred m.y. to get accumulated at the bottom of the upper mantle (Senshu et al., 2009; Safonova et al., 2015). In addition, the geology of the Japanese islands shows that older orogenic belts record several episodes of strong tectonic erosion suggesting a continuous supply of TTG material into the subcontinental mantle, not only in eastern Asia, but also in the western Pacific. During the last 500 million years as a result of those periods of tectonic erosion four volumes of the present-day crust of the Japanese islands must have been transported into the deep mantle over (Suzuki et al., 2010). The Izu-Mariana intra-oceanic arc system and a half of the Honshu arc to the north could have been tectonically eroded by the subduction of both the Philippine Sea plate and the Pacific plate (Hirata et al., 2010). Mechanisms of transportation of granitic material to the deep mantle were discussed in (Yamamoto et al., 2009; Kawai et al., 2009, 2013; Ichikawa et al., 2013). In brief, they are based on First principle calculations and later UHP experiments suggesting that granitic material possesses higher density than the upper mantle, but lower than the lower mantle. Therefore, the subducted granitic rocks, if reaching the depth of 270 km, get denser than the ambient mantle due to the coesite-stishovite phase transition and helped by the slab-pull force (viscous drag; Ichikawa et al., 2013b) (Fig. 43).

The subducted granitic material together with the descending slab may be separated at 520 to 660 km depths by the olivine-wadsleyite and wadsleyite-ringwoodite transformations, respectively. Then those materials must be gravitationally stable in the upper mantle to grow a second continent at its bottom (Kawai et al., 2013). Ten two questions appear. How much granitic material has been subducted into the mantle throughout the whole history of the Earth? How much granitic material can be present in the mantle transition zone between the depths of 520 and 660 km? The amounts were estimated by seismological and geological methods.

Kawaii and co-authors (2013) used PREM one-dimensional models to calculate P-wave and S-wave average velocities (Fig. 43). The PREM indicates that the lower half of the mantle transition zones is an area of fast P- and S-velocities. The two first P- and S-velocities in the upper and lower mantle were compared. The compositional heterogeneity inferred after the correction of temperature differences have long been an important issue in seismology debates. One model suggests enriched basaltic crust transformed to granite with a model amount of stishovite of ca. 10%. Another model considered the presence of harzburgite (Irifune et al., 2008) which can be segregated at MTZ depths. However, both models phased a series of problems. These two models can explain P-wave anomalies, but not S-velocities. But granitic rocks can accommodate the model amounts of SiO2 phase as high as 90%. Then, both P- and S-wave velocity anomalies can be explained by the granitic materials stored in the MTZ in an amount 6 to 7 times higher than that of the surface continental mass (Kawai et al., 2010, 2013).

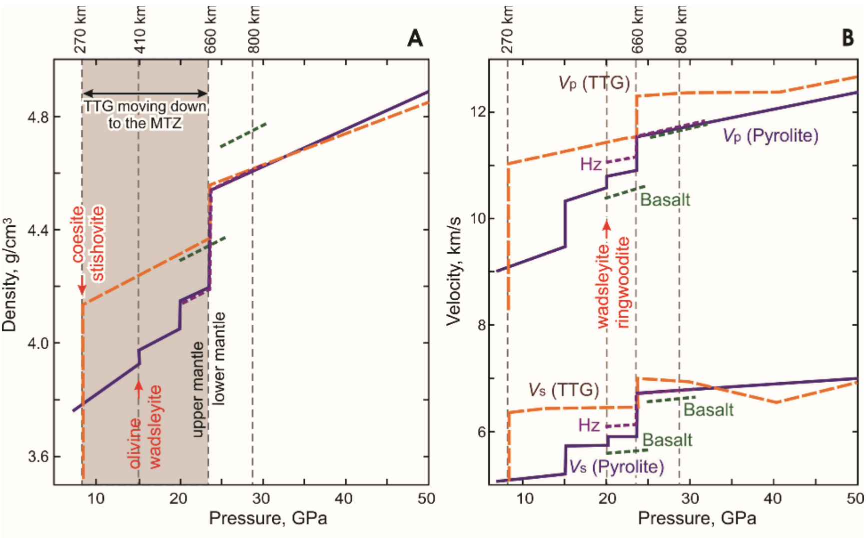


Fig. 43. Evidence for the possibility of continental material subduction to the deep mantle. A, B, first principle calculations and PREM-based models showing, respectively, the density (A) and seismic wave velocity (B) “jumps” of pyrolite (blue), TTG-type material (red), harzburgite (violet) and basalt (green) due to mineral transformations. Tectonically eroded and subducted TTG-type continental crust material becomes gravitationally stagnant between depths of 270 and 660 km and therefore can be accumulated near the bottom of the upper mantle, i.e., at the MTZ (modified from Kawai et al., 2013).

Ichikawa and co-authors (2013) assume that in the Archean, the temperature of the upper mantle was 170 K higher than the present upper mantle (Iwamori et al., 1995). Then the velocity of the oceanic plate was higher than the present velocity (Komiya, 2004; Komiya and Maruyama, 2007). By supposing that the plate velocity is three times higher than that at present based on a simple estimate from the average lifetime of oceanic plates (Hargraves, 1986), the subduction rate in the Archean should be 2.4 km3/yr at the temperature 200 K higher than that of the present, the slab velocity of 24 cm/yr, and the dip angle of 45°. The reduction of the flux due to the effect of this high temperature is counteracted by the effect of the high slab velocity. Hence, if the subduction mechanism of continental materials in the subduction channel has continued for about the past 4 Ga, about 1010 km3 of continental materials might have been transported to the mantle transition zone, which is bigger than the total amount of the present continental crust, 7.18×109 km3 (Cogley, 1984). If plate velocity in the Archean is slower than the present velocity as discussed by Korenaga (2006), the amount subducted during the Archean is considerably less because of its high temperature. In any case, the cumulative volume of the continental crust after 2.5 Ga is as much as the volume of the present continental crust.

Both approaches (Kawai et al., 2013; Ichikawa et al., 2013) suggest that the volume of the crust probably stored in the MTZ is seven times bigger than that of the present continental crust is consistent with the estimate by Rino and co-authors (2004, 2008), who calculated the amount of new granitic crust produced in the western Pacific regions hosting many intra-oceanic arcs. If those arcs were many in the Archean, they must have produced huge amounts of TTG mass. A question remains how effectively TTG rocks can be segregated if they form by the melting of descending slab. Anyway, the upper half of the mantle transition zone and the topmost lower mantle may carry floating granitic bodies. Those bodies are enriched in SiO_2 and therefore should be identified as anomalously dispersed seismic heterogeneities. A seismic wave method was proposed by Kaneshima and Helfrich (2003, 2009). They investigated anomalous later phases within a time window from 10 s to 120 s after direct P waves for deep and intermediate-depth earthquakes at circum-Pacific subduction zones and interpreted the anomalous phases as S-to-P scattering waves (wavelengths~10 km) from heterogeneities in the shallowest lower mantle (depths≤950 km). Several S-to-P scattering objects where elastic properties of the rocks must substantially change within several kilometers were detected in the shallowest 300 km of the lower mantle beneath the Circum-Pacific area. The researchers identified several dispersed seismic heterogeneities or scatterers (anomalies) in the top of the lower mantle and speculated that those are floating pieces of oceanic crust. However, the amount of SiO_2 phases in those rocks (<10% in volume) seems to be insufficient to cause such anomalous dispersion of seismicity. Granitic TTG material containing much more silica phases (up to 90%) can easily explain the anomalous dispersion by (Kaneshima and Helfrich, 2003).

Thus, the subducted continental materials are expected to accumulate around the base of the transition zone because continental materials are denser than the ambient mantle materials in the depth range between 270 and 800 km for tonalite–trondhjemite–granodiorite (TTG) materials (Kawai and Tsuchiya, 2010; Kawai et al., 2009, 2010, 2013). There is also a possibility that the subducted continental materials are distributed to the lower mantle at all depths via entrainment of the slab subduction. However, continental materials are considered to contribute to the evolution of the endmember enriched component—recycling of the delaminated old continental lithosphere in order to explain EMI and subducting sediments to explain EMI (Aizawa et al., 1999). However, the determination of the depth where the enriched materials are derived is rather difficult. In any case, the subducted continental materials may have an important role such as mantle plume formation due to its high radioactivity (e.g., Senshu et al., 2009).

5.4 First vs. second continents

If tectonic erosion of TTG crust has been common through the geological history of the Earth, how much of continental crust can be present in deep Earth's interior and where it is? By tracing the generation of TTG material at subduction zones through the Earth history and its tectonic erosion at trenches, we assume that the amount of TTG accumulated in the MTZ is equal to 7 to 10 continents of the surface. TTG material must be gravitationally stable at the bottom of the upper mantle. Through the geological history, the second continent must have increased in volume and its bulk-chemical composition must be similar to that of the surface continental crust. This is a reason why Kawai and co-authors called it second continent (Kawai et al., 2009). The first continents are gravitationally stable on the surface and move horizontally moved according to plate tectonics to cause supercontinent break-up, dispersion and final amalgamation. What is the driving force of the horizontal movement of continental plates then?

Continental crust forms and grows at convergent plate boundaries. Then the continental material can be tectonically eroded and descend together with directly subducting intra-oceanic arcs (see Chapter 4). All those must be transported together with descending slab to the deep mantle and accumulate and segregate from the slab at the bottom of the mantle transition zone, deep from the trench. The accumulation of TTG material in the MTZ is unidirectional. However, if the segregation is insufficient, the fragmental pieces of the TTG mass can drop partly together with the descending slab into the topmost lower mantle. Even if so, the segregated granitic material can selectively move upward to get underplated to the bottom of the second continent (Fig. 44). The depth of the segregation of granitic material in the lower mantle may be maximum 1500 km, if we assume the identification of SiO2-enriched rocks by (Kaneshima and Helfrich, 2003, 2009). Therefore, the pieces of the second continent may float at the depths less than 1500 km. The TTG material in the MTZ should gradually expand in time. During the accumulation, the TTG material can move as a ductile body horizontally and/or the heavy subducted slab with peridotite causes rifting of already formed second continent, similar to the continental rifting on the surface (Fig. 44).

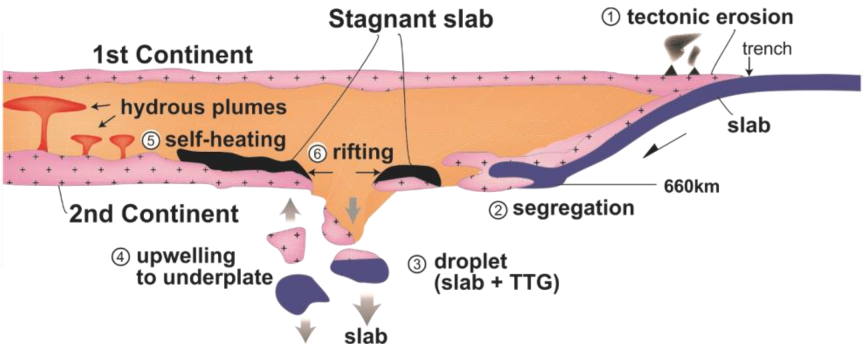


Fig. 44. A cartoon illustrating six major stages to develop a second continent in the MTZ. (1) Tectonic erosion of active margin/arc hanging wall and direct subduction of intra-oceanic arcs transport TTG materials into the MTZ. (2) in the MTZ, TTG materials can be segregated by bending slabs and density differences. Moreover, slab penetration is stopped by the presence of a TTG sheet at the bottom of the MTZ. (3) In time, the stagnant slabs, where phase transformations proceed should gravitationally collapse to rift the second continent, which drops into the lower mantle. (4) TTG fragments, if separated from the down-going blobs, would turn to rise to get underplated at a depth of 660 km to develop a second continent. (5) Those long-lived accumulations of TTG material would hear up the hydrous MTZ by self-heating at a rate of ca. 100 K per 200 m.y. (Senshu et al., 2009). (6) The generation of hydrous plumes (5) and related mantle upwelling can break-up the surface (first) continents as seen in post-Miocene East Asia hosting numerous rift zones and intra-plate volcanic fields (Yarmolyuk et al., 1995, 2014; Krivonogov and Safonova, 2017).

Figure 45 shows distribution of the second continents in the MTZ based on P-wave velocities showing seismic anomalies (Zhao, 2009). If the second continent is composed of 100% pure granite and if the time is not sufficient to self-heating, i.e. less than 100 Ma after the subduction, then the temperature will increase only to 100 K. That temperature increase is insufficient to regard the second continent as a 100% P-wave high velocity mass. For example, the Greater India as a high-velocity anomaly region underneath the Himalayan Mountains corresponds to the recently subducted continental crust (Fig. 39). The low-temperature slab consisting of peridotite plus minor MORB hardly can be separated from granitic material, and both may be seen as high-velocity anomalies. Therefore, we must compare the MTZ and oceanic regions in respect to the continental regions. Those anomalies are absent in the Pacific, Indian and Atlantic oceanic regions (Fig. 45). In the case of Asia, the subcontinental mantle transition zone is very heterogeneous showing contrast features, i.e. both positive and negative velocity anomalies. The seismic tomography patterns clearly indicate second continents under first continents. There are two particular regions: the first one is between the double-sided subduction zone in the western Pacific, including Asia; the second one is in western North America, up to the middle America trench, including the north-eastern corner of the Pacific Ocean, similar to that of the western Pacific, but back to 600 m.y.

In the patterns (Fig. 45) we cannot separate granitic second continents from low-temperature slab, if it contains P-velovity negative anomaly, both stored in the MTZ. But it cannot be continuous to upper and lower depths. The isolated regions of P-velocity anomalies may correspond to the mixture of lower mafic crust plus upper crust TTG. A mass consisting of 100% granite should yield a compositional anomaly more effective than a temperature anomaly. In other words, the effect from the mixing of slab peridotites and mafic rocks is higher than the effect from the heat generation of granite and the heat capacity of the mixed rocks. Therefore, at the moment such an approach is not much competitive as based on several assumptions, but represents an important message for future research as we need a new seismological technique to obtain a real pattern of the distribution of second continents in the MTZ. However, we can confidently conclude, that those MTZs beneath continents and several oceanic regions are special for remarkably heterogeneous structures with clear P-velocity anomalies. Among them are the western Pacific triangular region and similar but older tectonic domains in the northwestern corner of the Pacific including Alaska and Arctic polar oceanic realm.

P-wave velocity mantle tomography shows that those phenomena of mantle dynamics including the distribution of second continents (Fig. 45) are most important and critical underneath East Asia and western Pacific as estimated by (Huang and Zhao, 2006). Those tomographic data have

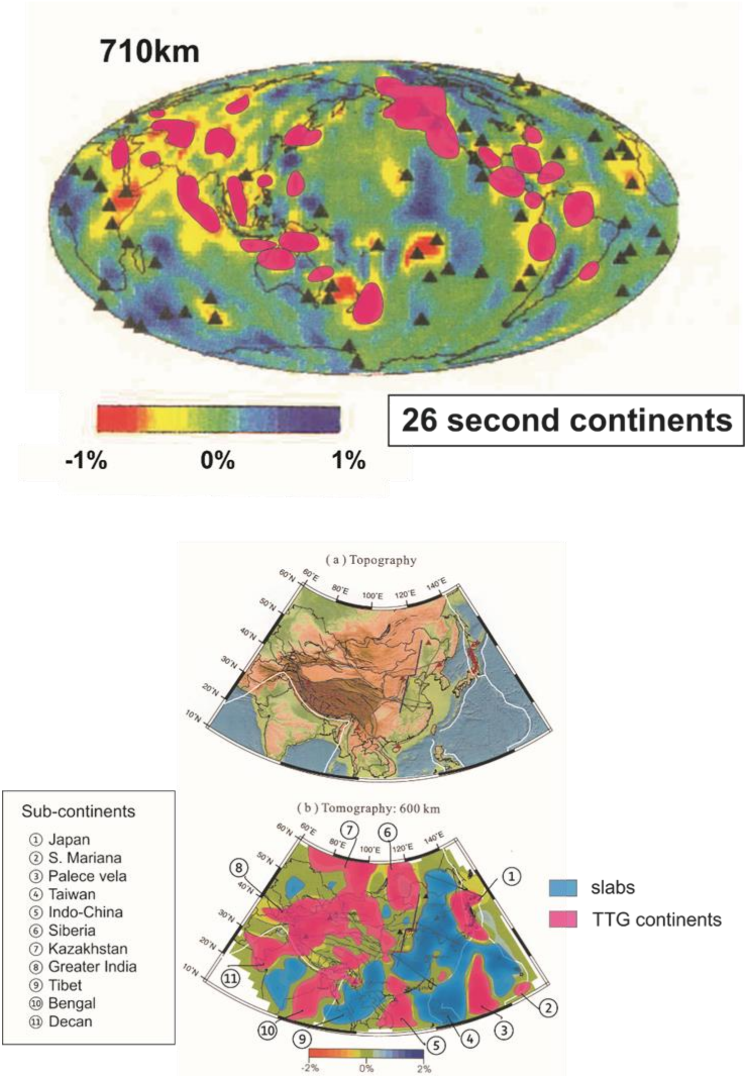


Fig. 45. The distribution of second continents in the MTZ at a depth of 710 km estimated by P-wave high-resolution whole mantle tomography after Zhao (2009). The low-velocity anomalies, which do not continue upwards and downwards, are nearly within the transition zone. These are regions where TTG materials has dominated during long time; hence, they are heated up by the TTG self-heating. Note that second continents occur mostly under the continents, in particular, beneath the youngest and largest continents, Eurasia to Australia, and North-central-South America. Those anomalies are absent under the Atlantic, Indian, and Pacific major oceans though, except for their margins. If this

rather high space resolution due to the large number of seismometers in the whole region, in particular, in China. At UHP conditions deeper than 270 km TTG rocks consist of 90% of stishovite. The values of P- and S-velocities depend on temperature and composition. Therefore, pure granitic material should be visible as an anomaly.

Let's discuss in more details tow to segregate TTG from lower crust mafic rocks or slab peridotites then? In other words, how much amount of a particular mass, in which granite occupy a certain portion together with slab peridotites, is dominating. This is the most important problem to understand tomographic images. Until recently the spatial resolution of tomographic images has been more than several kilometers. The level of several hundred km3 is a spatial resolution. To understand the spatial distribution of second continents in the MTZ, we make several assumptions. First assumption: the velocity anomaly in the upper part of the upper mantle, i.e. above the 410 km phase boundary, cannot extend to the lower mantle. In that case, the masses shown by tomographic images and satisfying those conditions, can be second continents. If a low velocity anomaly is present, a high-temperature plume must ascend to the surface to give intra-plate (hot-spot) volcanoes. A track from hot-spot volcano must be continuous even if it starts at the core-mantle boundary. Second assumption: anomalous P-velocity zones are not related to the CMB. If an anomaly occurs within the transition zone and if slab peridotites dominate in that anomalous region, the temperature issue is more effective to differentiate TTG, mafic rocks and peridotites. In that case a low velocity anomaly can be seen. On the contrast, if TTG dominates, high-velocity anomalies should be present.

The high-velocity anomaly under Asia (Fig. 45), partly below the 660 km boundary, may be related to second continents linked with the stagnant slab once subducted along the eastern margin of Asia as was predicted from the surface geology (Chapter 4, section 4.5). Figure 45 presents a whole Earth P-wave tomographic image for a depth of ca. 700 km depth. If this "snapshot" reflects a distribution of second continents, we can differentiate 26 continental masses. Two megaregions of second continents can be highlighted: the first extends from inner Asia, to Indonesia to New Zealand; the second extends from the north-west to the middle of the United States along its western coast.

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pattern is true, the distribution of second continents reflects the history of continent dispersal on the surface driven by plate tectonics. The bottom figure shows the distribution of second continents under Asia, based on P-wave high-resolution mantle tomography (Huang and Zhao (2006). At a depth of 600 km 11 continents are identified. However, young TTG must show high-velocity anomalies as it is still stuck with stagnant slab.

Two second continents are detected beneath the Japanese Islands and more to the east of the subducted slab. The second continents beneath Australia are really interesting, because the Australian continent got separated from Antarctica at ca. 65 Ma, then drifted to the north to move the Java-Sumatra trench. Therefore, in the past, those trenches probably occurred more to the south as was clearly shown by the tomographic images from another research group. The northward drift of Australia and the northward migration of the Java trench to a distance of at least 3000 km in respect to its original position suggest the TGG materials subducted during that time may still remain in the MTZ, which is now beneath the Australian continent. Of course, at the moment this idea of the distribution of second continents on the Earth is rather speculative. However, it can be confirmed in future by higher resolution seismic tomography including the reflection depths of discontinuous planes for granitic material, which must be different from that of peridotite already partly shown in (Kawai et al., 2009, 2013). Note, the distribution of second continents is highly restricted and they are absent under major oceans: Pacific, Indian and Atlantic. This is a very important point, because such a distribution is consistent with the reconstructed history of the opening of all those oceanic basins through time.

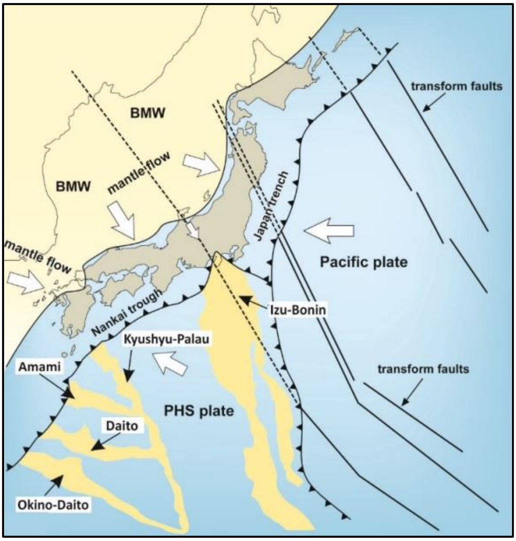
5.5 Accretion or subduction?

Tectonic erosion moves TTG material from the surface to the bottom of the upper mantle. The TTG material accumulates in the MTZ to make second continents. How stable are second continents, how they amalgamate and disperse? The first question is how to scrap off such granitic material during the transportation from the trench to the bottom? Does the transportation of granitic crust go with or without lower crust mafic rocks and slab MORB and underlying peridotites. Slab is covered by a 6 km thick layer of MORB above peridotites. When oceanic lithosphere begins to subduct from the trench axis to the mantle, for example, at the well-documented Mariana trench, the upper plate is cut by normal faults because the coupling of two plates is very weak as the subducting Pacific slab is very old, 150-200 Ma. Therefore, the hanging wall is destroyed under the effect of gravitation to collapse finally along listric normal faults and the rocks start sliding down to fill the trench. Those olistostrome-like deposits would subduct together with the down going Pacific slab, which will transport them to the mantle. At the same time, the hanging wall at its front must continuously subside under the extensional stress and finally the volcanic front collapses at its backside. Under all those factors we can suggest that the whole fore-arc region must be transported into the deep mantle. With time, the trench would move landward and a new volcanic front also would move landward, i.e. backward, through time. In case of the Mariana arc, the volcanic rocks of the older arc will be eventually cut by normal faults at a future trench. This is the case of the Izu-Bonin arc, which formed at 48 Ma and during the next 20 m.y. was evolving as an intra-oceanic volcanic arc. The Izu-Bonin arc now faces the trench. The Mariana Tertiary arc is separated from the trench by a gap. Direct drilling of the inner wall of the Mariana arc strongly indicates the complete missing of the Tertiary arc probably due to tectonic erosion (Hirata et al., 2010).

The density of granite is about 2.8 g/cm3 is very low compared to the 3.4-3.5 g/cm3 density of the mantle. Therefore, the TTG material, which is derived from volcanic and plutonic rocks, should be gravitationally very stable suggesting that continental crust cannot move to the mantle. This is a traditional idea and most geologists believe that the subduction of granitic material is physically impossible. However, this is definitely wrong. Subduction of lower density material is not a function of density only. It can be stacked with heavier material, crushed and kept in the grabens on the surface of old slab subducting at high angles, which are formed by normal faults (see Chapter 4, section 4.1). Another important function is size or volume. The Izu-Bonin arc, which is older than 40 Ma, is still growing by subduction zone magmatism. There are several other intra-oceanic arcs riding over the Philippine Sea: Amami plateau (early Cretaceous), Kyushu-Palau (50-30 Ma), Daito ridge (Tertiary), and Okino-Daito (Eocene). All those formed as immature arcs on the Philippine sea plate, but have almost completely subducted without any accretion to the hanging wall of the Eurasian plate (Yamamoto et al, 2009, a, b) (Fig. 46a). The collision of the Izu-Bonin arc and Honshu Island (Japan) formed negligibly small amounts of accreted units (Fig. 46b). But the hanging wall of the Honshu arc has been tectonically eroded and shortened to about 50 percent northward. Therefore, the upper hanging wall of the Honshu arc has been probably transported to the mantle. All of these observations suggest pencil-like shapes of the intra-oceanic arcs on the Philippine Sea plate. The thickness of the continental crust is about 20-30 km. In general, their geometric parameters, width, thickness, length, are not enough to provide accretion to the hanging wall, compared to another side of the Philippine Sea slab.

The present day size of India, without the Greater India underneath the collisional zone, may be sufficient enough not to be subducted though. Of course, arc-arc collisions are much smaller in scale than that

A



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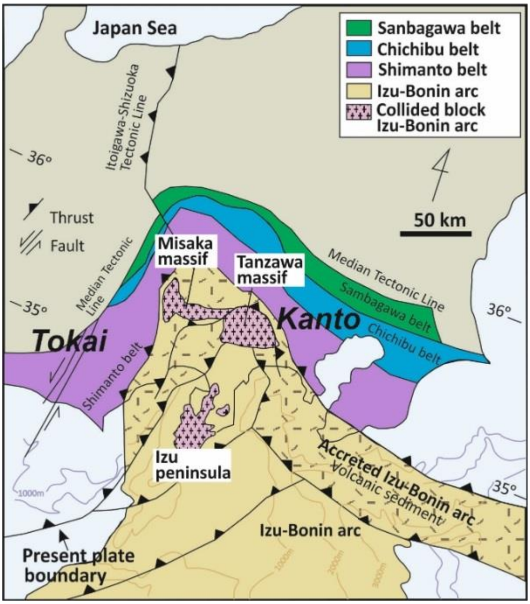

Fig. 46. A, Intra-oceanic arcs directly subducting into the Nankai trough with minor or nil accretion over the subduction zone. B, plan view of Izu-Bonin-Honshu collision zone in Japan (modified after Yamamoto et al., 2009).

of India and Eurasia, but they have been documented not only in south-western Japan, but also at the Sulawesi Island in Indonesia, where two parallel arcs collided to form embryonal continents. The parallel alignment of two colliding arcs is a key issue. In the case of collision of parallel intra-oceanic arcs, for example, in Sulawesi and northern Virginia, there is no subduction of arcs; they are scrapped-off the hanging wall. Parallel collision could provide double growth of island arc crust. These examples show that accretion or subduction do not depend on density only, but also on other parameters (size, volume, geometry).

5.6 Dispersion and amalgamation of second continents

The next important issue is tectonics of the mantle transition zone (MTZ) in the upper bottom part of the upper mantle. When TTG materials subduct down to the depth of 270 km, the phase transitions from quartz (3.0 g/cm3) to stishovite (4.2 g/cm3) and from feldspar to garnet+Na-K-hollandite+stishovite occur, and the granitic mass, composed dominantly of quart and feldspar become denser than the ambient mantle (Komabayashi et al, 2009, Kawai et al, 2010). All these minerals are high pressure phases. The next important barrier for the TTG granitic material descended together with the slab into the deep upper mantle is the 660 km boundary separating the upper mantle and the lower mantle. If TTG material and oceanic slab material experience phase transitions at depth, they must acquire densities higher than that of the upper mantle, and, therefore, physically can penetrate to the lower mantle. However, even if TTG is 100% transformed by high pressure phases, the density still will not be greater than that of lower mantle peridotites (Fig. 43a). In in the bottom of the MTZ, TTG material is slightly heavier than the lower mantle material, but it is still closely stuck with the slab suggesting that both TTG and slab should collapse to the lower mantle (Fig. 44, stage 3). If the slab in the topmost lower mantle falls down deeper and if TTG crust is scrapped off, then only granitic material will return upward, opposite to the material of subducted slab. Finally, that TTG material would get underplated to the bottom of the upper mantle second continents (Fig. 43, stage 4). If at the depth of 660 km the slab resists perovskite-wustite phase transformations, the micron grain size would promote a higher viscosity of the slab material (containing very fine-grained perovskite + wustite aggregates) and its related easier segregation of TTG crust. Therefore, the transformation of TTG material should always be discussed not only in terms of density of the subducted slab or granitic crust material, which would promote the growth of granitic continents at the bottom of the MTZ, i.e. at 660 km, but also in terms of the changes of viscosity, the size of eroded TTG material, the segregation of TTG and slab materials, which all are possibly not pure kinetic processes, we will get a different picture (Fig. 44).

Thus, the subduction and accumulation of oceanic slab, arc mafic crust and TTG materials in the MTZ cannot be provided only by a factor of the difference of their densities. An important issue is the mechanism of their segregation in the topmost upper mantle, which is also possible, and must be understood in more details and more quantitatively in future. If the TTG material gets finally gravitationally stable at depths of 520 to 660 km, that region must become a stability field of second continents. Therefore, second continents must grow right above the 660 km boundary, which is a gravitationally stable region. At that level, the arrived slab must be stopped by those second continents, both physically and rheologically (Fig. 44). However, when the total slab material gets accumulated right above the 660 km boundary and become too big, the second continents can be rifted away, with or without acquiring ductile properties depending on the temperature. They should become split into two pre more pieces. Then the previously stagnant slab may collapse into the lower mantle, with or without the second continents. If TTG material gets segregated from the oceanic slab material during the down going slab-avalanche, it may come up again and gets stable at ca. 660 km depths and/or gets underplated. Such an implication is now applicable only to the western Pacific regions, where the world best-known stagnant slab has been documented, however, it can be found in other regions and, possibly, reconstructed in older orogenic belts and/or beneath former continents. The next question is if such a process has been common through the geological history of the Earth, how much of continental crust can be present in deep Earth's interior.

5.7 Self-heating of second continents and its effect

The next issue is the self-heating generated by second continents, which was first evaluated and calculated by (Senshu et al., 2009). The authors show that if granitic material is present within the mantle during more than 100 m.y., the heating would increase the temperature to 200 K in Archean time. The initial concentrations of U, K and Th are presented in Table 2. In Phanerozoic time the temperature increase would be 50% lower, i.e. to 100K. The reason is the half-life time of decay of radiogenic isotopes. The modern western Pacific is a region of most active mantle convection. Since at least the Tertiary, five intra-oceanic arcs are directly subducting under south-western Japan, and the subduction is still ongoing (Yamamoto et al., 2009). The Japan, Nankai and Izu-Mariana trenches compose a 150 km wide and 1000-2000 km long zone of strong ongoing tectonic erosion. The hanging wall of the continental crust is being permanently destroyed. The surface geology of the Japanese Islands clearly show at least five episodes of strong tectonic erosion since 520 Ma. The volume of the TTG crust formed since then must be four times bigger than that of the present-day landmass of the Japanese Islands. Since the Cambrian at least four "Japans" must have been tectonically eroded and transported into the mantle under eastern Asia (see Chapter 4, section 4.5). However, roughly same amounts of granites were formed by slab melting during subduction. The volume of continental crust, which disappeared by tectonic erosion, has increased to a similar extent during the next 500 m.y. The position of the trench again shifted oceanward to a distance of approximately 100 km and retreated again 100 km back through tectonic erosion. The repeated occurrence of such kind of trench migration oceanward or continent-ward finally made the trench take its present day position. But huge amounts of granitic materials still must have been transported into the mantle.

Table 2

Radioisotopes in modern, 1 Ga, and 2 Ga mantle

	^{40}K	^{232}Th	^{235}U	^{238}U	Total
Half life time (10^9 years)	5.24	0.704	14.0	1.25	
Present mantle					
Abundance (ppm)	28.0	76.4	0.143	20.1	
Heat flux (10^{-12} W/kg)	0.816	2.02	0.082	1.90	4.81
1 Ga mantle					
Abundance (ppm)	48.7	80.3	0.384	23.4	
Heat flux (10^{-12} W/kg)	1.42	2.12	0.218	2.22	5.97
2 Ga mantle					
Abundance (ppm)	84.7	84.4	1.03	27.3	
Heat flux (10^{-12} W/kg)	2.47	2.23	0.585	2.59	7.87

Data from Van Schmus (1995).

The total mass of TTG crust of the second continents, which are stable around the 660 km depth, must constantly increase. If we accept that the rate of TTG production on the surface has been continuous during 3 or more GA and will continue in future, the rate of tectonic erosion must become higher than that of TTG. The crust of all “first” continents on the surface must be removed to the MTZ, i.e. they may finally disappear on the Earth. The TTG material moved to the MTZ is enriched with U, K, and Th, which are radiogenic elements capable to generate heat. From bottom and up the viscosity of MTZ mantle decreases to promote break-down reactions, including the transformation of hydrous ringwoodite and hydrous wadsleyite, which are stable in the lower and upper parts of the MTZ, respectively. The breakdown of these minerals may generate fluid and wadsleyite compositions close to that of magma and may finally generate hydrous plumes and trigger melting at a depth of 410 km (Fig. 44). Wadsleyite breaks down to form olivine plus water at the 410 km depth can then produce dehydrated fluid, dominated by water. But high-temperature water also contains big amounts of silicate components enough to generate magma. Without those reactions of dehydration, the overlying mantle would be dry, as olivine cannot contain water in its structure. The appearance of fluid or magma strongly changes viscosity. If the portions of melt are small and scattered, the viscosity of mantle material will remarkably increase. At 410 km a hydrous plume can be generated through the heating by the second continents and by the water stored in the MTZ.

Chapter 6. Supercontinent cycle and mantle dynamics

6.1 Supercontinent cycle

If the fragments of continental crust are clustered in several places similarly to the assembly of microcontinents to form continents on the surface, will they all gather in one place? What is the fate of those second or lost continents? May they follow the supercontinent cycle like that on the Earth? As granitic material is enriched in radiogenic heat elements, such as U, Th, K, may the lost continents within the Earth's interior play a critical role in mantle convection? If so, the lost continents may contribute to the generation of superplumes and enforce or affect the supercontinent cycle.

Supercontinent Pangea consisted of northern Laurasia and southern Gondwana parts. The first Gondwana formed at about 540 Ma and soon started rifted (Fig. 47). The rifted continents, such as North China, South China, India, smaller continental blocks of Indo-China and Taimyr and numerous microcontinents could move northward and finally amalgamate with each other and North America, East Europe (Russian Platform), and possibly Siberia to form another supercontinent Laurasia about 300-200 Ma, i.e. in rather short time, a core of future Eurasia (Paleo-Asia). Later, Laurasia connected with the southern remnant of Gondwana at middle latitudes to form supercontinent Pangea split into the eastern and western parts. The western part included North America and South America, the eastern part included Africa, East Europe (Russian platform) and Paleozoic orogenic belts to open the Atlantic Ocean. Thereafter, the former fragments of Gondwana successfully collided and get accreted to the southern margin of Paleo-Asia. The present six continents on Earth may form another supercontinent in future, in 200-300 m.y. from now, with Asia as a central nuclide (Safonova and Maruyama, 2014). In general, during the Phanerozoic, at least two supercontinents got rifted into 10 continents which then amalgamated into only one supercontinent taking a time of about 400-500 m.y. or even longer. Since the late Neoproterozoic at least three supercontinents existed: Rodinia back to 1 Ga, Gondwana at 540-400 Ma and Pangea at 250 Ma.

If the continents float on the mantle, continental rifting and dispersion must be controlled by mantle convection. Paul Hoffman proposed an "Inside-out" model back in 1991. The Rodinia supercontinent broke-up at about 700-800 Ma to make 8 to10 rifted continents, which finally amalgamated to form the Gondwana supercontinent, centered in the southern pole (Fig. 46a). P. Hoffman speculated that the Rodinia supercontinent was split into two in the center by a north-south trending rift to give birth to the Paleo-Asian or Paleo-Pacific Ocean. The Paleo-Pacific Ocean expanded but another Ocean on the opposite side of the Earth reduced and finally all oceanic plates subducted into the mantle, underneath the continents. Such highly speculative process triggers one supercontinent cycle. P. Hoffman named the process "inside-outs". However, such a process cannot explain the rifting of Rodinia and the drift of the South American continent. Then, what about a future supercontinent Amasia?

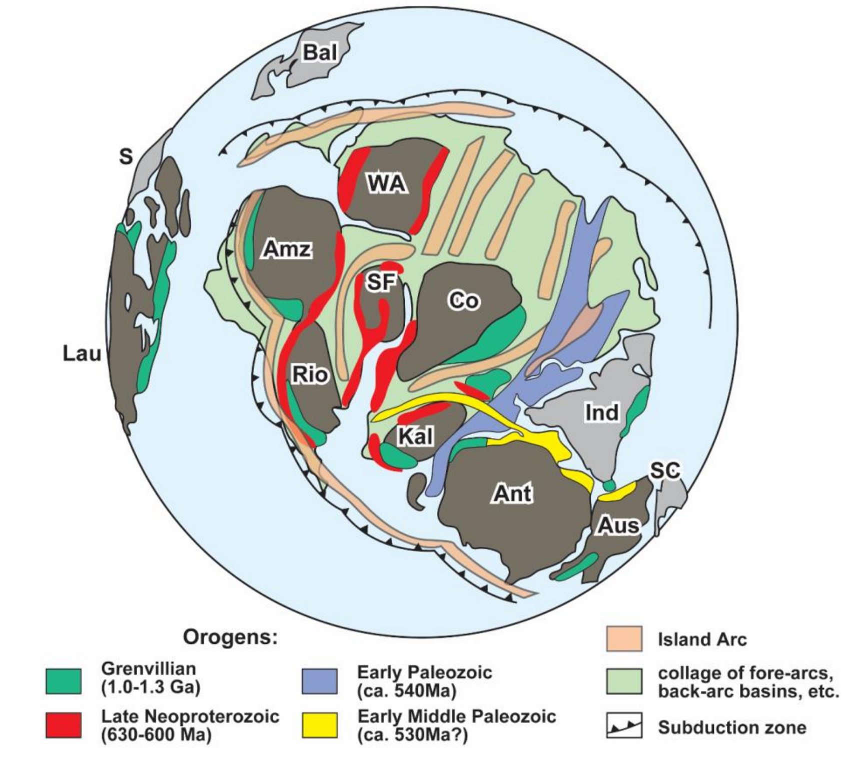

Fig. 47. Selective summary of the Gondwana supercontinent and its composing continental blocks at 540 Ma. Abbreviations for continental blocks: Amz, Amazonia; Ant, Antarctica; Aus, Australia; Bal, Baltica; Co, Congo; Ind, India; Kal, Kalahari; Lau, Laurentia; Rio, Rio de La Plata; S, Siberia; SA, South America; S, Siberia; SC, South China; SF – San Francisco; WA, West Africa.

6.2 Role of double-side subduction and triangular junctions in the history of Asia

Asia is the youngest continent in the world formed by collision and amalgamation of numerous continental blocks during late Neoproterozoic–Palaeozoic through Jurassic closure of the Palaeo-Asian and Tethyan ocean basins, followed by late Mesozoic Circum-Pacific and Cenozoic Himalayan orogenies (Figs. 16, 49). The tectonic evolution of Asia has witnessed numerous events of continental crust formation during the last 800 million years, including the breakups of Rodinia and Gondwana, amalgamation and breakups of Laurasia and Pangea supercontinents, and opening and closure of the Palaeo-Asian and Palaeo-Pacific oceans and their branches (e.g. Maruyama et al. 1989; Zonenshain et al. 1990; Sengör and Natal'in 1996; Dobretsov et al. 2003; Jahn 2004; Yakubchuk 2008; Safonova and Santosh 2014). The tectonic pattern of Asia is built around seven cratonic blocks or 'nuclei': Siberia, Kazakhstan, Tarim, North China, South China, India, and Indochina. The most prominent orogenic features of the present-day Asian continent is the Central Asian Orogenic Belt (CAOB) formed by the closure of the Paleo-Asian Ocean, the collisions of the Siberian, Kazakhstan, Tarim and North China continental blocks and amalgamation of Laurasia (Windley et al., 2007; Senshu et al., 2009; Safonova et al., 2011) (Figs. 47b, 48, 49).

Since the Triassic, the growth of continental crust in Asia has become even more active: its volume has increased by almost one order of magnitude, in particular, due to the formation of thick accretionary complexes in the western Pacific/East Asia (Permian–Cretaceous), amalgamation of northeastern Siberia (late Mesozoic), India–Eurasia (Eocene), and Arabia–Eurasia (Miocene) collisions (e.g., Parfenov and Natal'in 1986; Maruyama et al. 1989, 1997; Hall 2002; Ali and Aitchison 2008; Isozaki et al. 2010). Thus, since the beginning of Cenozoic time, the blocks of India, Arabia, and Africa collided with Eurasia and are continuing to move northward. In the future, Asia will experience further growth by accretion of oceanic seamounts and microcontinents currently sitting on the Pacific plate and moving westward toward the subducting margins of the western Pacific. Moreover, Asia is surrounded by two huge subduction zones with different polarities, the western Pacific and the India–Sumatra, which are moving to the northwest and northeast, respectively.

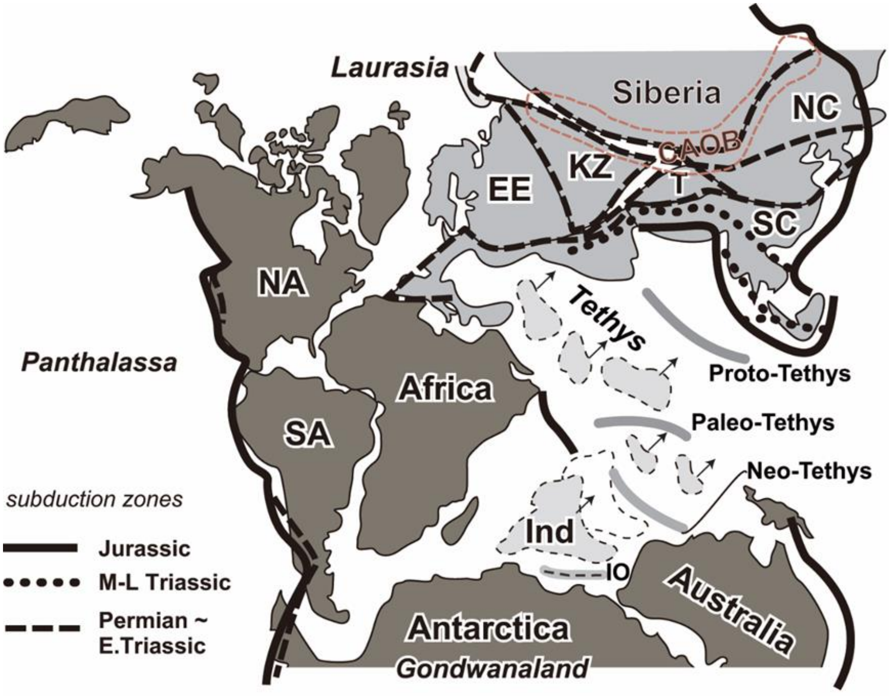

Fig. 48. Selective summary of the Laurasia to Pangea supercontinents and their composing continental blocks at; 300–250 Ma. Colors: light grey – Laurasia Group, dark grey – Gondwana Group. Abbreviations for continental blocks: Ind, India; EE, East Europe; KZ, Kazakhstan; NA, North America; NC, North China; SA, South America; T, Tarim; SC, South China.

Thus, the Paleo-Asia formed a nucleus of the Siberian continent until ca 300 Ma and then the North-China, Kazakhstan, East Europe, Tarim, and Indo-China successively collided and amalgamated by 200 Ma. At ca. 50 Ma the Indian continent collided with Asia. At ca. 200 Ma South Arabia collided with Europe to give rise to the Zagros Mountains. At ca. 6 Ma Africa collided with Europe to form the Alpine fold belt. As Asia is a region of double-sided subduction zone, it is now a frontier of a future supercontinent (Maruyama and Sakai, 1986; Maruyama et al., 1989; Hall, 2002; Windley et al., 2007; Safonova and Maruyama, 2014). To explain the supercontinent cycle in terms of P. Hoffman's "inside-outs", we need other important subduction zones, i.e. not only the Atlantic and Pacific oceans, but also the Indian Ocean to provide the south-to-north

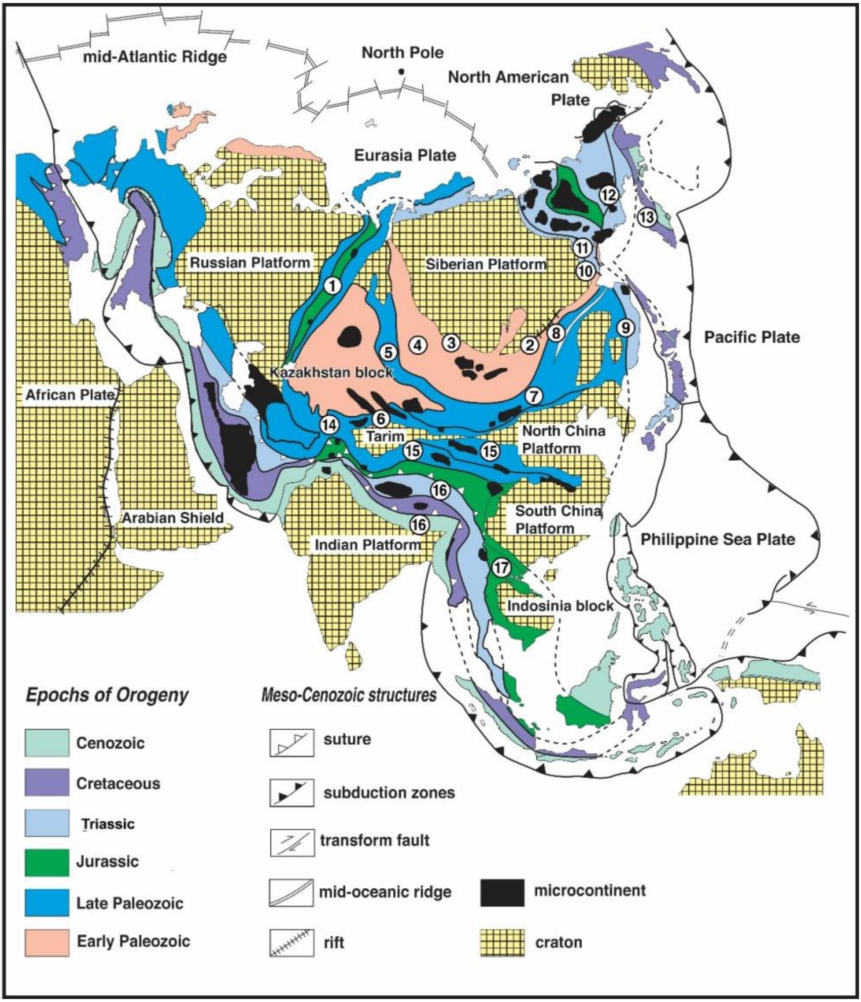

Fig. 49. Tectonic map of Asia. Numbers in circles are for major orogenic belts of Asia (selectively; modified from Maruyama and Sakai 1986; Maruyama et al. 1989): 1 = Uralian, 2–6 = Central Asian (2 = Baikal–Muya, 3 = Yenisey–Transbaikalia–North Mongolia, 4 = Altay–Sayan–NW Mongolia, 5 = Irtysh–Zaysan, 6 = Tienshan), 7 = South-Inner Mongolia, 8 = Mongol–Okhotsk, 9 = Sikhote–Alin, 10 = Verkhoyansk, 11 = Okhotsk–Chukotka, 12 = South Anuy, 13 = West Kamchatka, 14 = Pamir–Hindukush, 15 = Kunlun–Qinling, 16 = Tibet–Himalaya, 17 = South China.

movement. However, the distance from the south of the Indian Ocean to the north of the Pacific or Atlantic oceans is about 6000 km, therefore their related subduction zones cannot be linked with P. Hoffman's "inside-outs". But the distance from the Gondwana northern margin to southernmost Asia is much shorter. Therefore, we should call that pattern "inside-in". The "Inside-out" patterns may be needed to explain the movements around the whole Earth, like the 40,000 km length at the equator; otherwise we cannot call it "inside-outs". A combination of "inside-ins" and "inside-outs" can explain the supercontinent cycle better (Murphy and Nance, 2005) (Fig. 50).

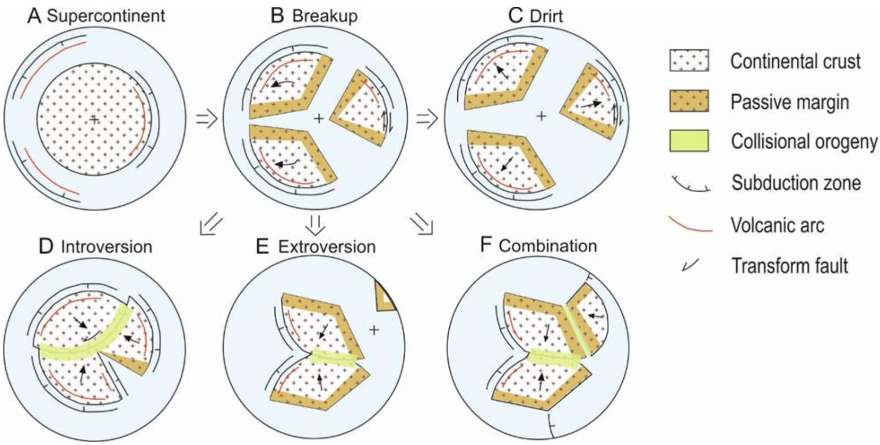

Fig. 50. The mechanism of breaking and making supercontinent by the inside-out and inside-in tectonic models of Murphy and Nance (2005). The simplest model is the inside-out model by Hoffman (1991), but it does not work in the case of supercontinents. A supercontinent is rifted into three blocks (B), which drift away (C) and then should finally amalgamate to form another supercontinent. However, such a process can satisfy perfectly neither the inside-in model (D), nor the inside-in model (E). A reasonable alternative is a combination of inside-out and inside-in models (F).

Let us consider the supercontinent cycle in terms of mantle convection. Numerical simulations of 3D mantle convection based on a random distribution of trenches showed the formation of overlapping subduction zones (Honda et al., 1993). If that happens, the domain for the down welling convection and the size of convection cell would expand twice, suggesting that the down-welling would "merge" down-going plumes to develop a much bigger plume, similar to a "black hole". That numerical simulation illustrates the fate of the double-sided subduction zones, which are now active in southeastern Asia. For example, one triangular region extends from the Kamchatka Peninsula to the south through the Japanese islands to New Zealand. The Pacific Plate is subducting along that line (Fig. 41a). The Indo-Australian Plate is subducting along the consuming boundary extended from New Zealand through Indonesia to the Himalaya. Therefore, a double-sided subduction develops underneath the western Pacific triangular region. Moreover, numerous much smaller microplates occur on the present day Earth (Komiya and Maruyama, 2007). This is a unique point in SW Asia, which is the only place on the present day Earth representing a frontier of a future supercontinent, as all continental plates, both big continents and microcontinents are currently drifting to that specific triangular point.

So, the present-day western Pacific faces the frontier of the future supercontinent Amasia, because Asia is characterized by double-sided subduction zones: one moves the Pacific plate from the east to the west; and another one moves the Indo-Australian plate from the south to the north. The Indian Ocean never acted as an inside-out pattern, but matching rather the inside-in model, because it is a small ocean. On the other hand, the Pacific was once a super-ocean, Panthalassa, and its size will reduce in time to form, finally, Amasia supercontinent in ca. 250 m.y. (Safonova and Maruyama, 2014) to provide the completion of the inside-out track. Therefore, the actual way of the formation of a supercontinent is a combination of inside-in and inside-out models/stages. This is true even in the case of the formation of Rodinia and its breakup to form Gondwana (Fig. 50). The collision and “inside-ins” and “inside-outs” amalgamations of all continents in 200-300 m.y. from now to form a future supercontinent should be interpreted in terms of double-sided subduction zones, the role of the Indian Ocean and mantle dynamics. A new supercontinent will be soon rifted and broken-up. However, what will be a driving force for the rifting? The double-sided subduction zones provide twice more efficient cooling of the underlying mantle due to its related large-scale mantle down-welling. However, to break-up the continents we need the heating of the mantle underneath a supercontinent to initiate the rifting. The modern western Pacific triangular region is the most effectively cooling domain on the modern Earth. On the other side, this is a region of very active magmatism, and, probably, regional metamorphism and tectonics related deformations as recorded by actives seismicity (frequent earthquakes) and extensive sedimentation.

6.3 Hierarchy, structure and history of continents and supercontinents

Nature is full of nesting structures. Let us list up Earth’s structures larger than atoms. A regular arrangement composes a regular mineral of a millimeter scale. Mineral assemblages compose rocks of several tens of centimeters scale. Different types of rocks compose rock bodies with different shapes and sizes of several to tens kilometer scale. The next is an orogenic belt consisting of an assemblage of rock bodies as large as one to several hundred kilometers wide and extending over a distance of more than one thousand kilometers; older orogenic belts are typically fragmented. The next type of nesting structures is continent consisting of a mosaic of orogenic belts of different shapes and different sizes. For example, the North American continent is 3000 km across, and the Eurasian continent is twice larger. The orogenic belts keep the past record of continent amalgamations and breakups.

Continents amalgamate to form supercontinents, which, in turn, can be rifted, dispersed and finally get amalgamated again, after 500 Ma to 1 Ga, to follow the supercontinent cycle. The rifting causes the opening of oceans and the formation of the underlying oceanic lithosphere, which, finally, disappears by the subduction into the mantle followed by the collision of continents. Continents commonly remain on the surface without subduction, if their size is bigger than, approximately, that of the Indian continent. However, the oceanic lithosphere disappears in the deep mantle with only minor fragments preserved within accretionary complexes and metamorphic belts of Pacific-type orogenic belts, but never in collision-type orogenic belts. Therefore, the continental lithosphere and the oceanic lithosphere show different behavior through tectonic cycles. The Earth’s crust is a part of either continental lithosphere or oceanic lithosphere, which have different structures. Oceans and atmosphere are present above the Earth’s crust, and the solid mantle and the core are present under the continental and oceanic lithosphere. That very simplified layered structure of the Earth defines its dynamics by the movement of lithospheric plates, which may include both oceanic and continental crust.

The formation of supercontinents is driven by mantle convection and provided by Y-shaped double-sided subduction zones. The geological map of the Earth is the most important and the only source of information for the reconstruction of the geologic history of the Earth including the history of continent and supercontinent amalgamation (Fig. 2). The cells of mantle convection are bounded by mid-oceanic ridges forming a network as a response to a curtain-like mantle upwelling. The source of the mid-oceanic ridge curtain upwelling was detected by tomographic images at a depth of about 410 km, at the top of the mantle transition zone (Zhao, 2004; Zhang and Tanimoto, 1993). The tomography-based shape of a superplume is different; it looks like a mushroom (Fig. 51).

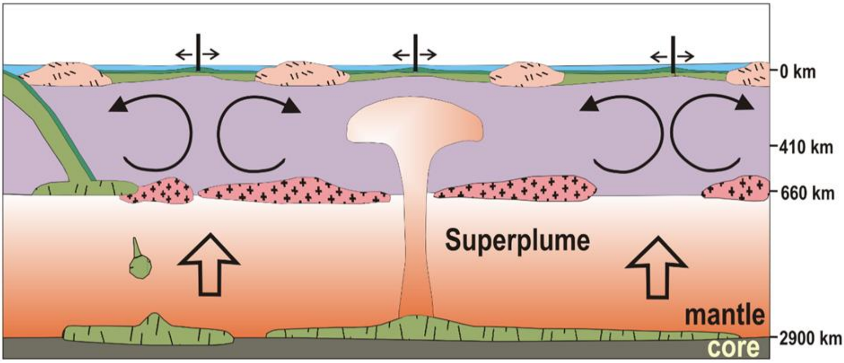

Fig. 51. Shape of a superplume approaching the base of the oceanic lithosphere (Santosh et al., 2014).

Anomalously huge superplumes are derived at the core-mantle boundary (CMB) at a depth of about 2700 km. There are only two superplumes recognized so far: southern Pacific and African. Both are rooted right above the CMB, and their radii exceed 3000 km. The top surfaces of those mantle plumes are present just underneath the continental (Africa) or oceanic (Pacific) lithosphere to upwell the surface to about 300 m. The size of the topmost part of the superplumes is again 3000 km across. Between the top and the bottom, the radius looks reduced, about, 1000 km. The boundary surface is continuous, changing gradually with the surrounding mantle. There are 4 to 5 hotspots in the southern Pacific suggesting that the superplume splits into several smaller mantle plumes. Therefore, the bundle magnitude of smaller plumes forms the entity of one superplume (Figs. 6, 49, 52). The two mantle upwellings, one in southern Pacific and the second underneath Africa, are “balanced” by the super downwelling underneath Asia (Fig. 52), which formed at ca. 250 Ma. These three are major mantle convections inside the Earth, which drive the whole Earth convection system. The two upwellings and the only one downwelling underneath coupled with the Y-shaped double-sided subduction zones mean that all continents on the Earth will be swallowed in time (Fig. 42). The frontal parts of the downwelling are the northward and westward Y-shaped double sided subduction zones, which is the entrance of the continents approaching from the east and south, i.e. the frontier of future supercontinent Amasia (Fig. 42, 53).

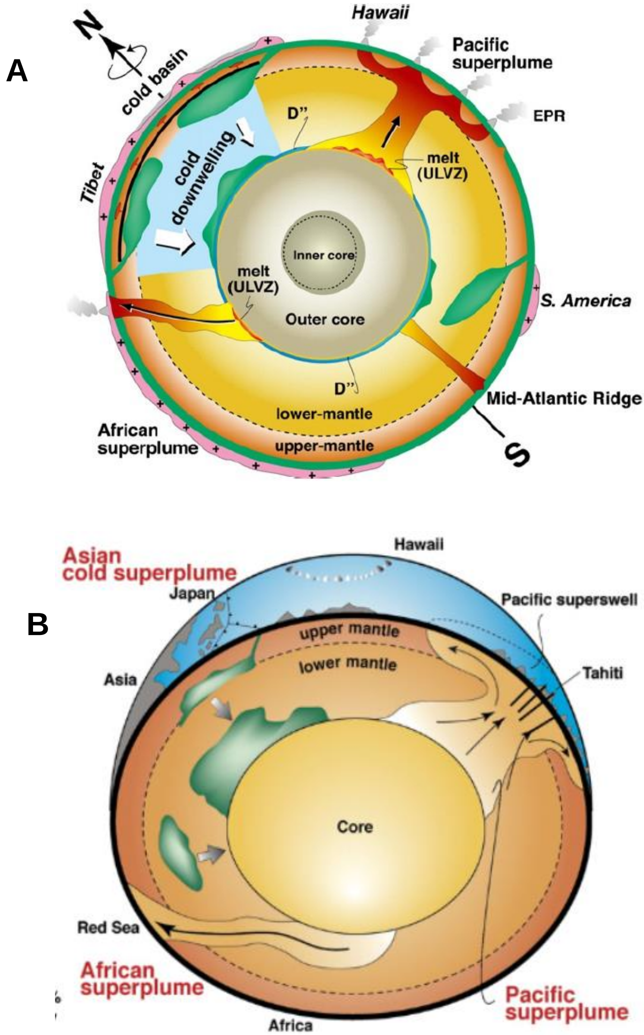

Fig. 52. A, Major convection pattern of the Earth's mantle showing superplumes and one super-downwelling. B, P-wave velocity perturbation suggests two large mantle upwellings in the Pacific and Africa, and one super-downwelling in East Asia. Recent improved tomographic images by S- and P-waves support this classic tomographic image of a superplume (modified after Maruyama, 1994; Maruyama et al., 2007). EPR, East Pacific Rise; ULVZ, ultra-low velocity zone.

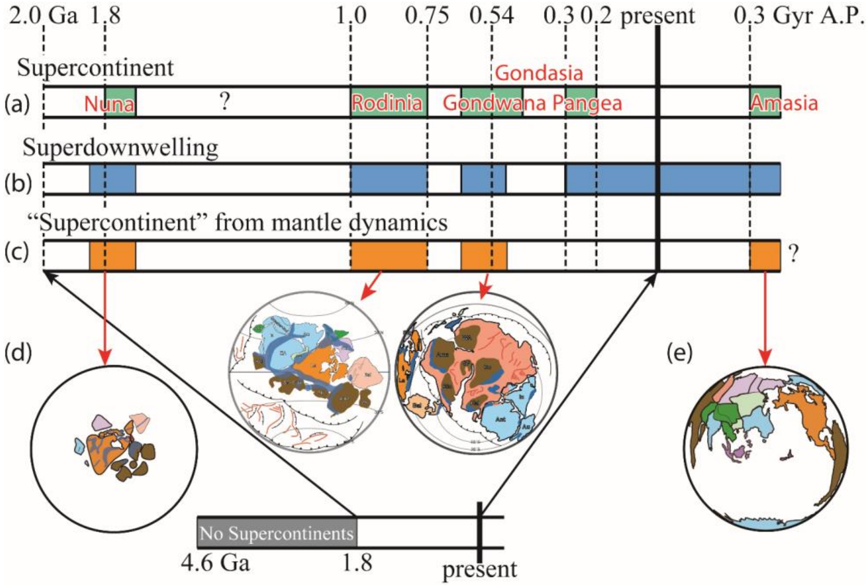

Fig. 53. Formation of supercontinentц through the history of the Earth (modified after Senshu et al., 2009, 2010). (a) Supercontinents: no before 1.8 Ga. The patterns of sutures along which the continents amalgamated indicate double-sided subduction zones in space and time, suggesting the birth of a strong mantle downwelling under a supercontinent (b). This pattern shows whether or not that was a real supercontinent (c). The shapes and relative sizes of supercontinents are shown in (d). In fact, Pangea was not a true supercontinent and should be regarded as a transient state from the breakup at 0.5 Ga to the future supercontinent Amasia (e).

As we mentioned above, during the past of 300 m.y., the Asian continent has been assembling around the nucleus of the Siberian craton by the collisions if six continents located more to the south to form finally the composite continent of Eurasia, the history of which we can elegantly explain by studying the geologic map of the Earth and Eurasia (Figs. 2, 49). Students can enjoy investigating the history of the Earth during last 200 m.y. by using mental scissors to cut off younger elements from geologic map of the Earth reconstruct the past continents. First of all, shall we cut off the oceanic lithosphere of the Atlantic Ocean, because there are no subduction zones, except minor ones, which are negligible, along the passive continental margins of North and South America, Europe and Africa. After we cut off that oceanic lithosphere, we will get four continents combined: North, South America, Africa and Europe. Next let us move to the Indian continent and the Indian Ocean. The northern margin of the Indian Ocean is a subduction zone extending from the Mediterranean Sea through the Himalayas and Indonesia to New Zealand, but its southern margin is of Atlantic type, i.e. no subduction zones. So, we can cut off the oceanic lithosphere and place together Australia, India and Antarctica, which once belonged to Gondwana. If we join the Gondwana blocks and the northern Laurasia, we can make one continent, called Pangea, as there was the westward subducting Panthalassa Ocean connecting with both continents around the central America. In the Pacific Ocean there are no continents at all. Therefore, the reconstruction of the Pacific Ocean does not affect the shaping of 200 Ma continents on the Earth.

To understand the amalgamation of Pangea, let us see the distribution of orogenic belts in the geological map to show those related to Gondwana. Gondwanaland consists of several continental nucleus older than 540 Ma: Congo (Africa), Pilbara (Australia), Dharwar (India) and several other continental fragments surrounded by a matrix of ca. 440 Ma orogenic belts (Figs. 47, 53). That kind of structure is quite similar to the late Paleozoic assembly of Laurasia, which nuclei are surrounded by a matrix of Triassic to Jurassic orogenic belts. The matrix of Gondwana blocks are continent collision-type orogenic belts formed back to 540 Ma. Note that North America collided with Gondwana in middle Paleozoic time. Therefore, sensu stricto, the Gondwana supercontinent was the largest by the middle Paleozoic, but also at the same time, Gondwana land supercontinents has already been rifted and dispersed at that time. Therefore, the timing required to create one supercontinent can be different, because the timing of continental drift always overlaps with the timing of continental dispersion.

The Rodinia supercontinent existed before the Gondwana supercontinent. The timing of Rodinia was a very important to reconstruct Gondwana group continents, many of which preserved their shapes, for example, Africa. However, in the reconstruction of Rodinia the shape of Africa is fully different because continental break-ups, dispersions and amalgamations destroy the present-day shape of continents. Within Asia the shapes of Gondwana and Rodinian group continents have not preserved. Further back in geological time, the Nuna supercontinent existed at. ca 1.9-1.8 Ga (McMenamin and McMenamin, 1990; Hoffman, 1991; Rogers and Santosh 2002). The accuracy of the shapes of the Nuna and even Gondwana supercontinents turns out to be less precise. Therefore, geological constrains are not enough to discuss the shape of Pangea because the oceanic lithosphere is completely gone and the shape of continents were different in the past.

The third difficulty comes from the possibility of tectonic erosion and transportation of big amounts of granitic materials. For example, an interesting constrain is the modern size of the continental crust. Figure 54 shows the growth curve of the TTG crust. The continents older than 1.8 Ga occupy only 30% of the total continental crust of the Earth (Rino et al., 2008). The results of the U-Pb dating of detrital zircons from mouth sands of world major rivers are nearly identical to the curve calculated by the age-distributions of world orogenic belts (Fig. 2). The total amount of 2.6 Ga continents is only 18%. Since then, the amount has been increasing gradually up to the present 100%. The Archean mantle was 150-200 K hotter than the present. The presence of double-layered mantle convection in the Archean suggests that extensive orogenic activities must have created huge amounts of TTG crust in the Archean. Why is this not the case in present-day Earth? The answer is that more than ten times the volume of TTG crusts subducted deep into the mantle and get accumulated at the bottom of the upper mantle as second continents (Chapter 5).

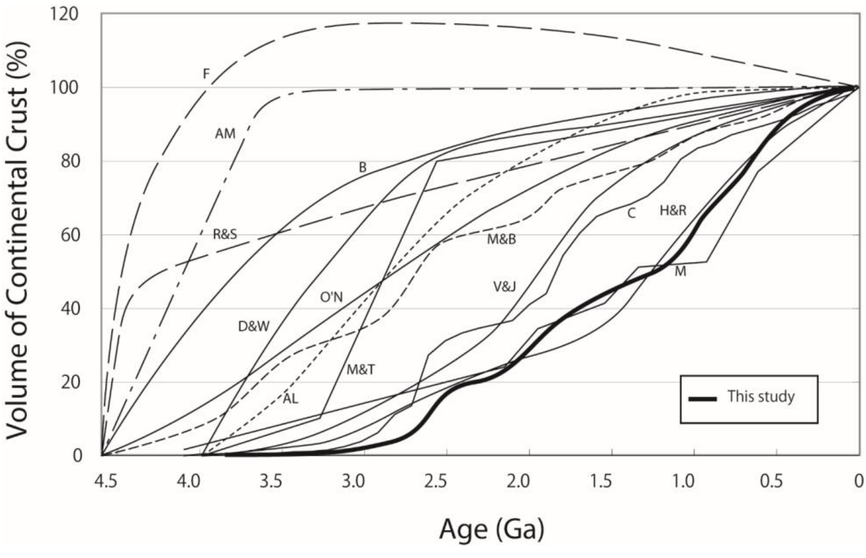

Fig. 54. What does the growth curve of the TTG crust over time mean? (modified after Rino et al., 2008). The numbers of the estimated curves: C, Condie (1998) (integration of his Fig. 1); V&J, Veizer and Jansen (1979); M&T, McLennan and Taylor (1982); AL, All`egre (1982); O'N, O'Nions et al. (1979) (integration of their Fig. 6); M&B, McCulloch and Bennett (1994); R&S, Reymer and Schubert (1984), D&W, Dewey and Windley (1981) (based on the values in their text, Proterozoic section slightly modified (originally zerogrowth) to arrive at 100%); B, Brown (1979) (his model 1); Am, Armstrong (1981) (based on his assessment that all growth took place in the first 1 Ga); F, Fyfe (1978). The bold line corresponds to the U-Pb ages of mouth sand detrital zircons of world major rivers covering ca. 40% of world's continents.

The U-Pb detrital zircon ages and the world geological map (Fig. 2) show that the Neoproterozoic Earth was shaped largely by the Grenvillian and Pan-African orogenies. Out of these, the Grenvillian orogeny has long been regarded to be of minor nature in terms of global-scale orogenic episodes, whereas the Pan-African orogeny has been widely recognized in many continental fragments, although not in major parts of Asia. The Grenvillian orogeny contributed significantly to the formation of the continental crust. The time period between 0.6 Ga and 0.8 Ga marked the climax at the dawn of the Pan-African orogeny. Continental crust formed in this period is concentrated in the Pan-African orogenic belts widely across the globe. These regions were widespread over the half hemisphere of the globe, and were subsequently reduced in size after they moved to form Laurasia. The normalized frequency distribution of zircon ages from river mouth sand over the world clearly demonstrates that Neoproterozoic and (0.9–0.6 Ga) and Grenvillian (1.3–1.0 Ga) peaks define the largest population.

The Neoproterozoic was the most active period of crust formation in the Earth. The cold basins, formed right after the assembly of Rodinia, exhibit a basin chain fringing the northern periphery of Rodinia, which turned into sites of mantle upwellings and led to the rifting and separation of the supercontinental assembly (Fig. 53). The continents then moved northwards after the formation of Gondwana at ca. 540 Ma, and enlarged the northern half of the supercontinent Pangea since 250 Ma.

The reconstruction of supercontinents by P. Hoffman is based only on 30% of continental fragments and is essentially different from the reconstruction of Rodinia and other supercontinents. The orogenic belts older than 1.8-1.9 Ga together with collision-type orogenic belts occupy about 80% of the whole North American continent (Figs. 2, 55). The western margin of the North American continent, from Alaska to California, formed in Phanerozoic time growing to several hundred kilometers in the westward direction.

The north-eastern margin of North America grew southeastward. The related orogenic belts occupy about 20% of its area. The oldest pieces of continental crust of North America represent 6 embryonal continents, cemented by 1.9-1.8 Ga collision-type orogenic belts (Fig. 55). Therefore, the oldest supercontinent is probably Nuna or Columbia (Hoffman, 1991; Rogers and Santosh, 2002). However, term "supercontinent" does not necessarily mean all continents of the Earth to be amalgamated. We suggest that the definition of supercontinent should mean about 80% or more of the continents on the Earth surface to be connected together like those formed Gondwana, Pangea and, probably, Rodinia. These three supercontinents do not satisfy the amount of 100% continents connected together.

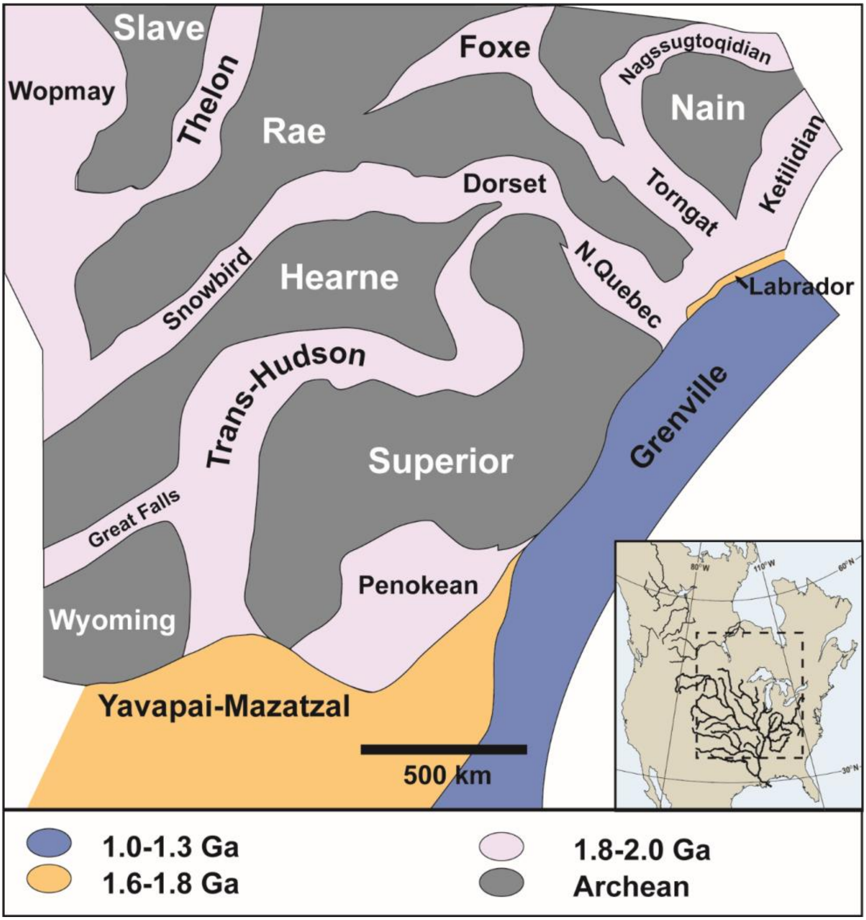

Fig. 55. A geologic map of central North America (modified after Hoffman, 1988). Seven or more embryonic continents formed dominantly in late Archean time amalgamated to form the first supercontinent Nuna on the Earth by 1.8 Ga. Unraveling the Nuna and embryonic continents clearly indicates that the continent formed by intra-oceanic island arcs around an embryonic continent 300 km wide and 1000 km long during a period from 2.6 Ga to 1.8 Ga. The Archean embryonal continents formed by Pacific-type orogeny only.

6.4 Supercontinent cycle

Name “Wilson cycle” was introduced as a credit to John Wilson, a Canadian geophysicist and geologist who achieved worldwide acclaim for his contributions to the theory of plate tectonics. He is the person who suggested a third plate boundary as important, which later Dewey and Bird named “Wilson cycle” implying the birth and death of ocean paired with supercontinent cycle. Some of geologists and geophysicists later were confused to differentiate “Wilson cycle” and “supercontinent cycle”. In this book we follow the definition of “supercontinent cycle” different from the life cycle of ocean, i.e. sensu stricto “Wilson cycle”. By tracing the history of supercontinents we should go back to 1.8 Ga, as discussed before, but still there is a kind of debates about a possibility of formation of supercontinents going further back to at least 2.5 Ga, which is the time of probably the oldest supercontinent Kenoraland, although there is still no enough evidence about that. Thus let us accept that the supercontinent cycle started from 1.9-1.8 Ga by the formation of Nuna/Columbia, then Rodinia, Gondwana and Pangea amalgamated at about 1.0 Ga, 540 and 250 Ma, respectively, to be possibly followed by future supercontinent Amasia (Fig. 53). The distribution of continents records the history of heterogeneous continental lithosphere and oceanic lithosphere. Since Nuna, i.e. at about 2 Ga, the time required for the assembly of another supercontinent has become shorter. For example, the time gap between Nuna and Rodinia was about 1 billion years, then 540 Ma, i.e. approximately twice shorter between Rodinia and Gondwana, and again twice shorter 250 Ma between Gondwana and Pangea. The meaning of that phenomenon is not well understood yet, but it may reflect the secular change of the pattern of mantle convection pattern through time.

If the supercontinent cycle is defined by the change of mantle convection, why the cycled repeated faster than before? However, the supercontinent cycle took only 50% of the Earth history, which started at ca. 4.6 Ga. Then what happened during the first 2.6 Ga of the Earth’s history when there were no continents, i.e. from the Hadean to the Archean.

To answer this question we should consider the most critical information preserved in the Nuna supercontinent formed at 1.9-1.8 Ga. Nuna was the first supercontinent, which size was 3000 km across, i.e. comparable with the thickness of the mantle. If we accept the model of whole mantle convection, even if episodic, it should have started as early as at about at 1.9-1.8 Ga. The most critical geological information is recorded in Laurentia blocks (note, Laurentia is different from Laurasia). The geologic map (Fig. 2) clearly shows the 1.5-1.8 Ga embryonal continents, which are different from cratons in the shape of the matrix and the length of orogenic belts. There are no younger continents similar in shape and size, in particular, those of Phanerozoic age. The Phanerozoic continents are, of course, bigger in size, but also many of them have rectangular shapes, i.e. different from the embryonal continents, which are 100-200 km wide and 700 to 100 kilometers long. Those 1.5-1.8 Ga six embryonal continents formed by the 1.9-1.8 Ga collision-type orogenic belts (Fig. 55). It took only 100 m.y. to make a continent, which size was about 3000 km across, i.e. comparable with the present day North American continent. For example, the Wyoming craton and several others represent intra-oceanic island arcs, which parallel collision formed embryonal continents after 2.6 Ga (Fig. 55). Intra-oceanic arcs consist dominantly MORB-type oceanic crust overlain by deep-sea sediments later incorporated into accretionary complexes. The accretionary complexes also included OPS sediments and alternated sandstones and mudstones of trench or post-orogenic turbidites. All those rocks of accretionary complexes were subsequently intruded by granites and overlapped by dominantly calc-alkaline felsic volcanics. The sedimentary rocks are typically very thin (less 10 meters) and have mafic composition with a subordinate amount of felsic sediments. Those features suggest intra-oceanic arcs similar to those in the western Pacific, which are isolated from continents by back-arc basins and not much elevated above the sea level but produce rather thick turbidite capping the OPS units. The successive parallel arc-arc collision helped the embryonal continents to grow as P. Hoffman demonstrated for North America (Fig. 55).

The formation of continental crust then proceeded in several stages (Fig. 56): 1) formation of juvenile crust at intra-oceanic arcs (eight or more in case of North America but up to 35 worldwide), 2) arc-arc parallel collision; 3) growth of embryonal continents; 4) collision of embryonal continents; 5) growth of a continent up to 3000 km size, first on the Earth (Nuna – Laurentia- North America). Geology of North America suggest that Laurentia appeared at 1.8-1.9 Ga approximately coevally with several other world cratonic blocks of western Australia, south-east Africa, India, north China, Siberia, west Africa and South America. Note that several cratonic nucleus of South America contain Archean orogenic belts and exhibit geological structures similar to those of as North America (Fig. 2; Santosh et al., 2009). Most important messages come from geological information. Big continents grow by intra-oceanic island arcs. The Earth's oldest island arc is recorded in the

Acasta Gneiss of central Canada going back to 4 Ga (Stern and Bleeker, 1998). The geological structure of the Akasta region includes less than few meters across fragments of gneiss containing small fragments of older granite formed, according to the U-Pb ages of zircons, at ca. 4.03-3.9 Ga (Iizuka et al., 2009).

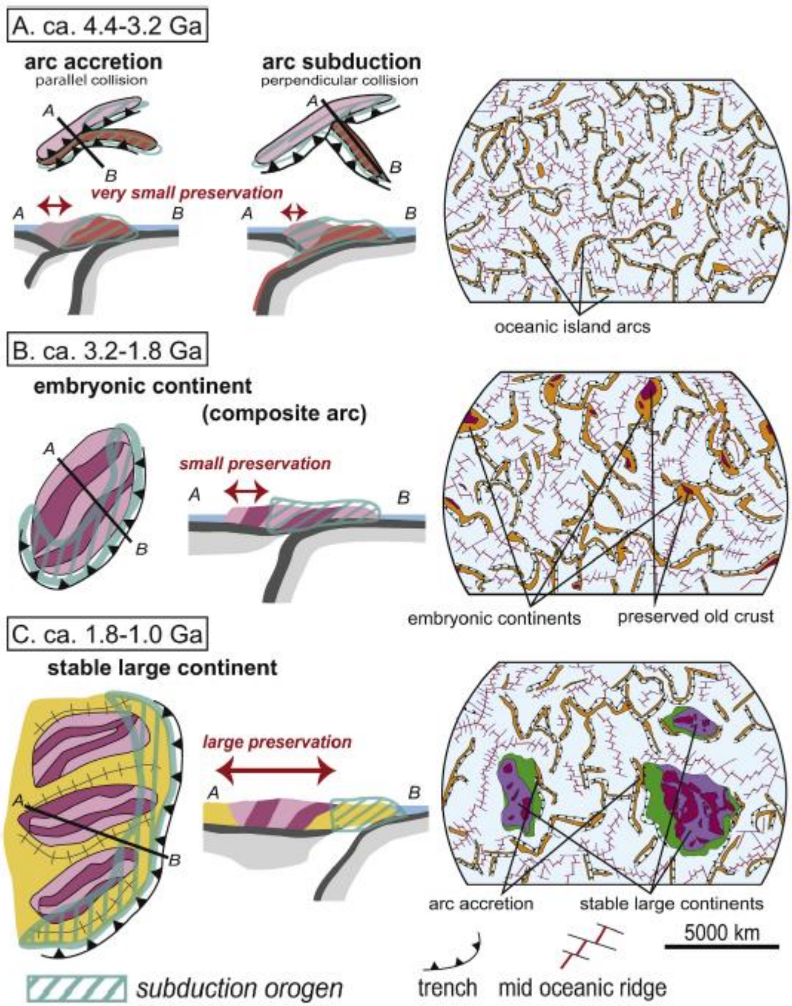

Fig. 56. Schematic models of continental growth and island arc accretion since the Hadean in map view and in profile. For comments see the text.

The next question, which we must address, is when the supercontinent cycle started. As we discussed above, the first supercontinent, Nuna, formed at 1.9-1.8 Ga (Hoffman, 1988, 1997). It was surrounded by intra-oceanic arcs, which parallel collision formed embryonal continents of irregular shapes floating in oceanic regions. However, most of them, in particular, single intra-oceanic arcs must have directly subducted into the deep mantle at subduction zones and therefore disappeared from the surface. During ca. 4.0-3.2 Ga (Fig. 56a), numerous island arcs formed, accreted and subducted. In the case of parallel collision of two island arcs, they easily got amalgamated to each other and grew into minor land masses (arc accretion). In contrast, in the case of perpendicular collision of one arc to the other, the crust of the colliding arc likely subducted smoothly into the mantle (arc subduction).

Moreover, the subduction erosion also occurred to destruct pre-existing arc crusts. Consequently, the preservation potential of the continental crust was very small. During ca. 3.2-1.8 Ga (Fig. 56b), the collided composite arcs formed embryonic continents, which were larger than individual island arcs but smaller than the modern continents without having significant amount of older crust. Tectonic recycling occurred along active continental margins to suppress the net growth of continental crusts. During ca. 1.8-1.0 Ga (Fig. 56c), plural embryonic continents got amalgamated to build larger continents comparable to modern ones. The pre-existing crusts in the interiors were protected from subduction-related tectonics along the active margins to create a possibility of the preservation of old crust, e.g., the formation of the first supercontinent Nuna/Columbia. On the other hand, the accretion of island arcs along the peripheries efficiently increased the total mass of continental crusts.

On the surface of the Earth, the formation of small embryonal continents started back to 2.6-2.7 Ga or earlier (Fig. 56). That time, there were several small landmasses above the sea level with stromatolites of shallow sea environments providing increasing oxygen contents as shown by hematite-bearing banded iron formations (e.g., Windley 1977, 1984, 1995; Maruyama et al, 2007a; Condie, 1982, 2009). However, the formation of large continents (ca. 3000 or more km across), which size was comparable with the thickness of the mantle must had waited until 1.9-1.8 Ga. All these evidences suggest the convection of the mantle started before 2.7 Ga and may have a double-layered structure. The size of the cell of stable convection should be about 700 km, which is the thickness of the upper mantle. By accepting those sizes, the number of oceanic plates on the surface driven by double-layered convection was probably about 400 as calculated by (Komiya et al., 1991). On the modern Earth, the number of major plates is about 10 and the number of smaller plates in the western Pacific region is about 40. The comparison between the present-day major plates and the number of Archean oceanic smaller plates allows us to speculate that the birth of the Nuna supercontinent at 1.9-1.8 Ga and the birth of major plates, which size is close to that of the modern plates, i.e. about 3000 km across, corresponds to the thickness of the mantle. That time, the large cells of convection, similar in size to those of the modern Earth, emerged through mantle overturn (e.g., Fuji and Bougault, 1983; Condie, 1998; Komiya et al., 2004; Rino et al., 2004; Bédard, 2018).

If the mantle convection system is double-layered, the upper mantle never goes down into the lower mantle, and vice-versa. If plate tectonics already operated in the Archean and produced new plates, then oceanic plates must be consumed at trenches. If such a process continues over 1 billion years, then the internal heat would produce magma to cool the upper mantle down. The subduction of cold basaltic oceanic plate would also cool the mantle. Therefore, in time the upper mantle must cool down to stop finally the plate tectonics. On the contrary, the upper mantle covered the lower mantle and acted as a thermal blanket. In addition, considerable amounts of radiogenic elements in the lower mantle must generate internal heat. Therefore, the lower mantle tends to be less cooled down. With the average temperature of the upper mantle either increasing (by the heat from the lower mantle from underneath) or decrease (by oceanic subduction and magma production on the top), with time the temperature of upper mantle would inevitably decrease to make the density of the bottom of the upper mantle and the density of the top of the upper mantle close. This would results in local density overturns, when the upper mantle is replaced by higher temperature lower mantle. That could happen within small regions, but then propagate over the whole mantle. Breuer and Spohn (1995) called such a process mantle overturn.

Thus, the lower mantle material did not melt until the event of mantle overturn at 2.6 Ga, which resulted in the bottom of lower crust to heat up, leading to A-type magmatism, as well as inactive mantle dynamics in the so-called “Boring mid-Proterozoic”. Finally, the Nuna supercontinent began to break up and disperse over the world. The Nuna derived continents finally got amalgamated at 1 Ga to form supercontinent Rodinia in the equatorial region, unlike Gondwana and Pangea. Gondwana existed between 540 and 400 Ma on the southern pole, and Pangea formed at 250 Ma by two groups of continents clustered at the northern and southern poles and connected by a thin junction (Senshu et al, 2009; Figs. 48, 53).

Chapter 7. Pacific-type orogeny, history of the Earth and mantle dynamics

The world geological map (Fig. 2) shows that there were no big continents during the early half of 2.6 b.y. suggesting there were no collision-type orogenic belts in Archean time. However, there were numerous intra-oceanic island arcs well documented within Archean cratonic regions (Figs. 55, 56). Therefore, during 2 b.y. the continental crust has been continuously produced by Pacific-type, not collision-type, orogeny. For example, there are several excellent examples of an Archean Pacific-type orogenic belt in the world. Among them are the 3.8 Ga Isua belt of southern Greenland, the 3.5-3.1 Ga Pilbara craton of western Australia, the 3.6-3.2 Ga Barberton greenstone belt of southern Africa, the 3.6-2.9 Ga Dharwar craton of India and the North American craton (Fig. 55), all dominated by Archean Pacific-type orogenic belts with TTG intrusions and accretionary complexes. From a petrological viewpoint the reconstructed potential temperature of the early Archean mantle (Ota et al., 1992; Komiya et al., 2007) was 200-250 K degrees higher than that of the present mantle. All those geological and petrological data strongly indicate that the mantle convection was more active than in Phanerozoic time to produce huge amounts of granitic crust (Nebel et al., 2018).

The amount of the pre-Proterozoic TTG-type crust must be 7 times larger than that of the present day continental crust of the Earth (Rino et al., 2008; Fig. 54). That estimate was obtained from the modern western Pacific region, which area is equivalent to the whole Archean on-land surface of the Earth. However, that region occupies only 7% of the total surface of the Earth. If we extrapolate that amount to the whole Archean surface, we can estimate the production rate of the continental crust in the Archean using the temperature of the modern mantle and the average age of the western Pacific oceanic plates. If we assume that plate tectonics began at 4 GA, then the total amount of the continental crust should be produced after 2 b.y. However, most of that crust has not survived on the modern surface of the Earth. About 90% of all world's intra-oceanic arcs are concentrated in the western Pacific, however, most of them seem to be subducted directly into the mantle without any accretion on the hanging wall (Fig. 46a). If the Archean Earth was covered by numerous (about 400) oceanic microplates (Fig. 56), along which consuming boundary intra-oceanic arcs formed, most of them must have been subducted to the mantle. Archean island arc pieces may be still in the mantle transition zone to form the second continent (Chapter 5). The 7 amounts of modern continental crust to be produced throughout the Earth's history as estimated by (Rino et al., 2008) is consistent with the estimates based on the PREM model and P and S-wave velocities (Kawai et al., 2009, 2013).

7.1 Archean Pacific-type orogenic belts

Probably all Archean orogenic belts belong to the Pacific-type (Chapters 1, 2). At 2.7 Ga, the collision and accretion of arcs formed small embryonal continents, which amalgamated to form the oldest 2.6 supercontinent of Kenoraland or the 1.9-1.8 Ga Nuna supercontinent about 3000 km in size. Nauna was the first well documented supercontinent, which is clearly seen in the world geological (Fig. 2).

The best world standard of Archean Pacific-type orogeny is the early Archean Isua belt of Greenland (Fig. 57). It was investigated in details by Japanese geologists using the model of ocean plate stratigraphy, which was applied to reconstruct the history and structure of accretionary complexes of Japan (Komiya et al., 1999, 2002, 2003; Maruyama and Komiya, 2011). This 3.8-3.7 Ga world oldest orogenic belt is located in southwestern Greenland. It has a well preserved initial structure less affected by regional metamorphism compared to other Archean belts (Komiya et al., 1999). Empirically, geologists differentiate granite-greenstone belts and high-grade metamorphic gneissic belts. The high-grade metamorphic belts form after granite-greenstone belts. Pacific-type granite-greenstone belt are characterized by the presence of OPS, greenstone (originally oceanic crust) and TTG intrusions. Metamorphism strongly modifies primary greenstone and granite to become high-grade gneiss. The degree of metamorphism is higher than the amphibolite facies; extensive recrystallization goes parallel with extensive deformation. Therefore, it is not easy to identify protoliths and original structures of those granite-greenstone belts, which definitely formed and evolved as Pacific-type orogenic belts (Windley and Garde, 2009).

B. McGregor, S. Moorbath, V.R. Bridgewater and A.P. Nutman wdere the first who studied the Isua region of southern Greenland (Bridgewater et al., 1974; Moorbath et al., 1977; Nutman, 1986). Critical debates were about the presence of peridotite, gabbro basalt, and chert with minor clastic material. Granites and their metamorphosed equivalent gneisses intruded those rocks. Most of them considered those rocks units of geosynclinals origin and, later, as rifted continental regions. These pioneer geologists published geological maps and predicted that that is an accretionary complex with well-preserved duplex structures, bedded chert and other rocks typical of Pacific-type orogenic

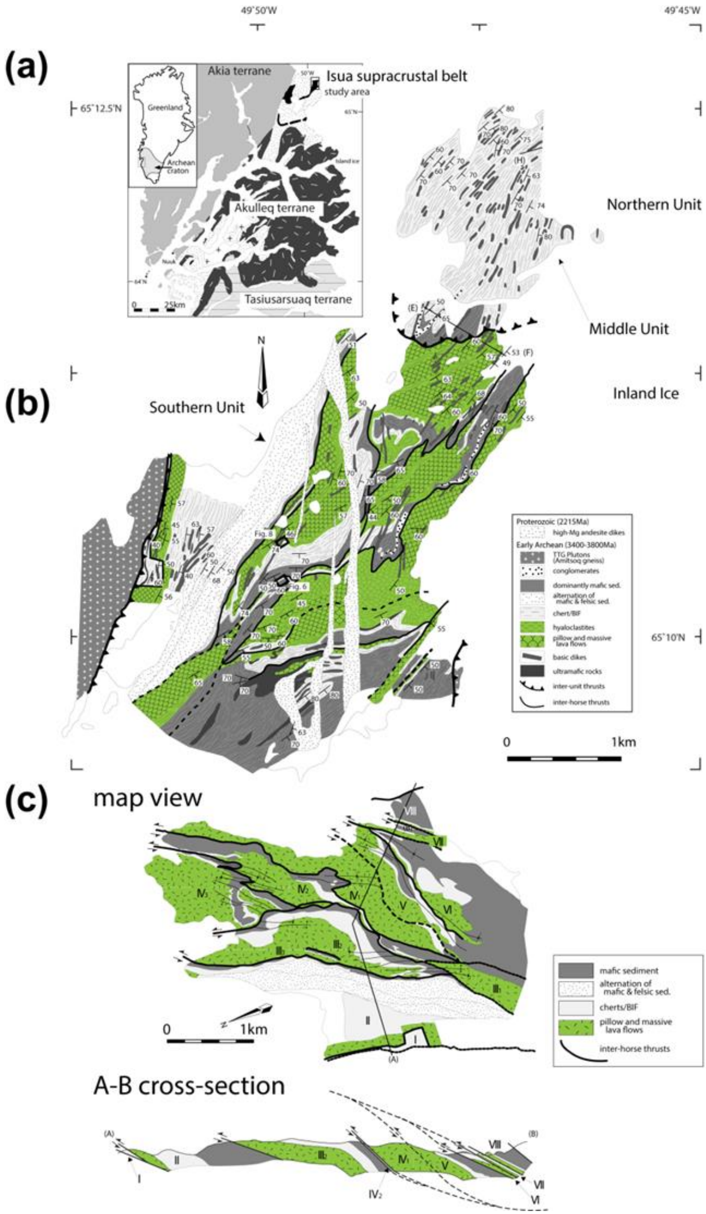

Fig. 57. Geologic map of the 3.7-3.8 Ga Isua belt, Greenland (Komiya et al., 1999).

belts. Figure 57a shows the central part of the geological map of the Isua belt. This oldest orogenic belt on the Earth is a Pacific-type. The geological structure of the belt represents a duplex consisting of 12 horses bound by faults; the boundary thrust clearly indicates the southern convergence (Fig. 57b). The duplex suggests strong horizontal layer-parallel shortening (Fig. 57c). Each horse preserves the characteristic succession of ocean plate stratigraphy (OPS), from bottom to top: pillow basalt, bedded chert and siliceous mudstones, corresponding to hemipelagic sediments and, finally, mafic to felsic mudstones and sandstones and subordinate conglomerate dominated by gravels of mafic to felsic volcanic rocks (Fig. 57c). That structure demonstrates that Archean plate tectonics was already operating at 3.8 Ga in an oceanic domain to form an island arc caused by Pacific-type orogeny. The analysis of geologic structure makes reconstruction of the OPS possible, thereby inferring subduction polarity southward. At the stratigraphic bottom there are low-K tholeiitic pillowed basalts compositionally close to modern MORB, however, the content of FeO is 2-3 weight percent higher (Komiya et al., 1999). The horses consist of well-preserved OPS. Each horse defined in figure 56 preserves the characteristic OPS, and successive accretion of each horse can be structurally estimated from the sequence of duplex formation (Fig. 57). The southward subduction of the oceanic plate in the Archean formed a series of duplex underplatings of a piece of oceanic crust and overlying deep-sea sediments. Two mid-oceanic ridges and one transform fault are estimated to be subducted (Komiya et al., 1999). Those geological and geochemical data suggest that the Isua Pacific-type orogenic belt formed by successive subduction of oceanic plate with minor seamounts (Fig. 58).

The Isua supracrustal belt records the evidence of 3.8–3.7-Ga tectonics similar to modern plate tectonics. It also records the materials formed by the plate boundary process at that time. These have similarities and differences with materials formed by modern plate boundary processes. Differences are (1) more frequent intraoceanic or off-ridge volcanism, as demonstrated by the frequent intrusions within chert beds. This may reflect the higher- temperature Archean and more turbulent mantle upwellings. Petrologic results from Isua basalts also show 150—200 K higher potential mantle temperature and a FeO richer source mantle than that today under the mid-oceanic ridge (Komiya et al., 1999, 2004). (2) More frequent accretion of oceanic basement than in the Phanerozoic. This may reflect thicker oceanic crust than the present. If the crust were thicker than 10 km, the mechanically weak plane easily developed within crust during subduction of oceanic plate, according to recent rheological experiments (e.g., Shimamoto et al., 1993). Therefore, accretion of oceanic materials is not common during Phanerozoic subduction except during deep accretion.

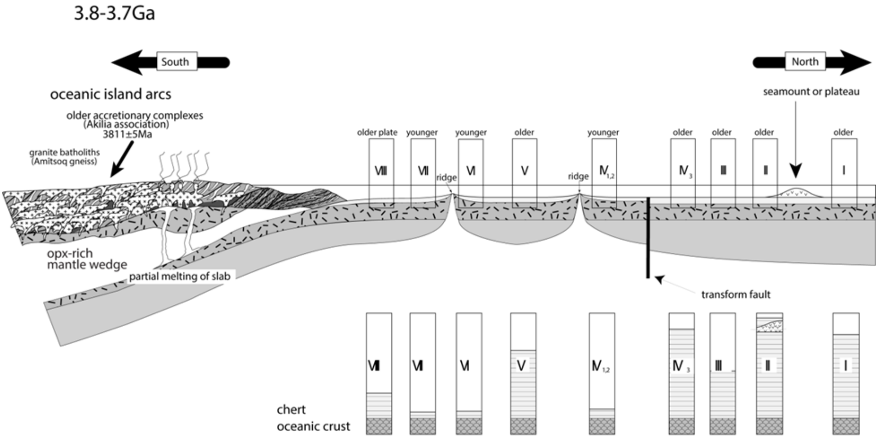

Fig. 58. Geotectonic evolution of the Isua orogenic belt in the early Archean (Komiya et al., 1999).

The characteristic aspects of accretionary complex in the Isua region are as follows: (1) the top-sitting trench deposits are rather thin (less than 10 m); (2) their composition is dominantly basaltic, unlike those dominated by quartz-feldspar dominated sediments of the Franciscan complex in California and accretionary complexes of Japan, where the extremely thick trench sediments are derived from a backward volcanic front dominated by TTG-type granitoids and volcanic rocks. These features make the major difference between Phanerozoic continental margin and intra-oceanic arc accretionary complexes and the Archean Isua accretionary complex. In spite of those differences, many similarities exist between the Archean and the present. The Isua accretionary complex is structurally similar to those formed adjacent to modern immature island arcs, such as Mariana. Although the chert was possibly derived from chemical precipitation at the mid-ocean ridge, the occurrence of chert beds is quite similar to Phanerozoic bedded chert with regard to the obvious bedding planes and the lithological gradation from bottom to top (Fig. 58).

The idea about the intra-oceanic arc origin of the Isua belt makes it similar to other Pacific-type orogenic belts worldwide, in particular, those of Archean age, e.g., the Pilbara Craton in the North Pole Dome area of

Australia (Fig. 59a). The North Pole region has been thoroughly investigated by geologists of Australia, Japan and other countries. The geological map (Fig. 59b) shows seven distinct units defined as Units I to VII. The units represent tectonic piles resulted from layer-parallel shortening caused by subduction-related underplating of oceanic slabs. Therefore, if correct, the apparent-structural top (Unit VII) must be the oldest, whereas the lowest Unit I must be the youngest. The zircon derived U-Pb ages are shown in the traditional lithostratigraphic column (Fig. 59c). The U-Pb ages of zircons separated from bedded chert support the geology shown in (Fig. 59b) (Kitajima et al., 2008). A rhyolite from Unit IV yielded a U-Pb age of 3660 ± 52 Ma. The older zircons from the rhyolite, which were formerly interpreted as xenocrysts, appeared crystallized in-situ igneous grains because they are common and have no overgrowth rims.

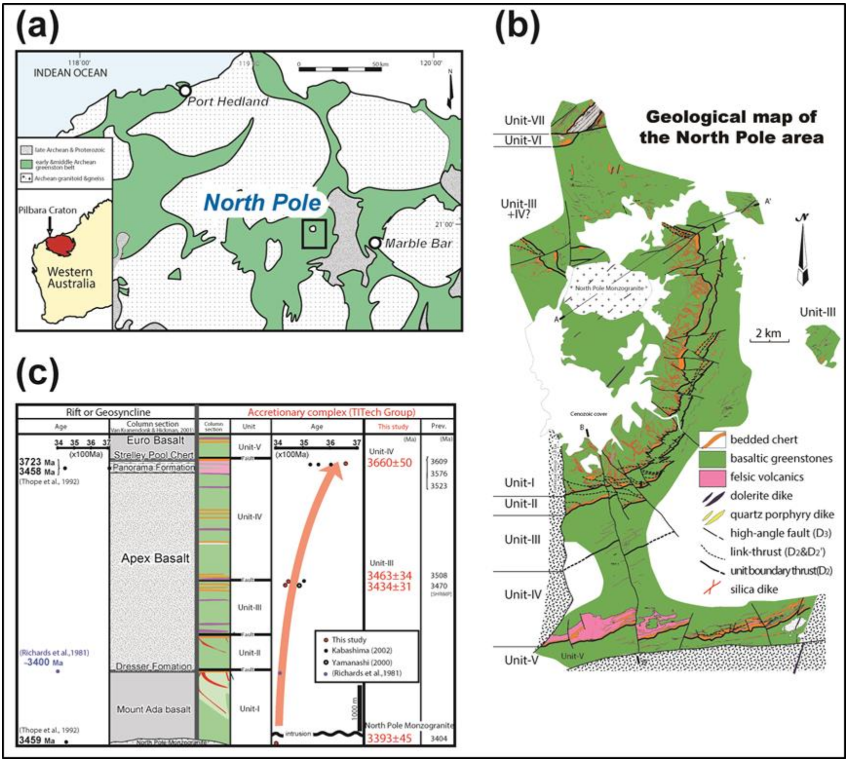

Fig. 59. Geologic maps of Archean orogenic belts in the Pilbara craton, Western Australia (Kitajima et al., 2008).

Figure 60 shows a geologic map of the northeastern area of Unit I, which is a region of duplexing. The duplex consists of a sequence of MORB overlain by chert accreted to the hanging wall of an older accretionary complex. The top thrusts and bottom thrusts can be clearly seen. The authors reconstructed the order of horses from I-10 to I-19 and the relative direction of underthrusting of the descending oceanic slab. These age relations and thrust deformation at the contacts between major units suggest that layer parallel thrusts played a major role in the growth history of the North Pole Dome. The style of tectonic growth history of the North Pole Dome, the overall downward younging stratigraphy, and the right-side-up stratigraphy of the individual units are very similar to those of modern circum-Pacific accretionary complexes. Zircon geochronology for old greenstone/chert successions provides a powerful approach to determine the evolution of Archean greenstone belts, for which radiolarian chronology is not applicable.

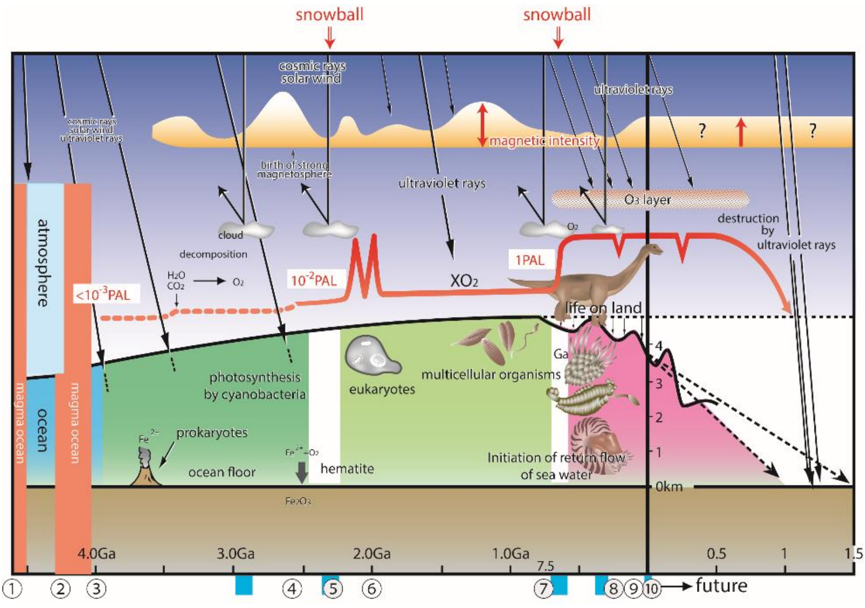

Fig. 60. Comprehensive schematic diagram showing secular variations in geological, biological and environmental factors through the history of Earth (modified after Maruyama et al., 2001, 2007).

The Kaapvaal craton South Africa, the Dharwar craton of India and those of North America also host accretionary complexes. The formation of accretionary complex and subsequent intrusion of huge amounts of granitic TTG magma are characteristic features of Pacific-type orogeny, which make it different from collision-type orogeny. Another point is although granitic crust may be dominant, but present under the sea level in the Archean. The second important aspect is the total length of Pacific-type orogenic belts. In the Archean, there were about 35 cratonic nuclei over the world, but they were much smaller and more fragmented compared to the major Archean cratons, such as North America, Australia, and South Africa. In any case, the size of Archean Pacific-type orogens is always less than 100 km long, suggesting that the sizes of Archean plates were not as big as in the Phanerozoic, probably, they were oceanic microplates. This, in turn, suggests that the mantle convection was double-layered in the Archean (Komiya et al., 2004). The parallel arc-arc collision was responsible for the formation of embryonal continents in the Archean. The subsequent intrusion of TTG magma into a continent composed of arcs and the high-temperature metamorphism at middle to lower crust levels changed those accretionary complexes to form gneisses. The original boundary of the arc-arc collision can be reconstructed by a suture line though (Windley and Garde, 2009). The third aspect is the formation of embryonal continents during a period from 2.7 to 1.8 Ga early to middle Archean (over 2.7 Ga) Pacific type orogenic belts.

In spite of arc-arc collisions in the Archean, most of the arcs probably subducted into the deep mantle. Although the Archean mantle possessed higher potential temperature, the surface area of the geological entities older than 2.6 Ga is very small, only 15% compared to the continents of the present-day Earth. The percentage of late Archean and early Archean terrains is about 10% and <5%, respectively (Fig. 54). However, since 2 Ga through the Earth history, the extensive Pacific-type orogeny was followed by arc-arc collision and amalgamation until the formation of large continents by a scenario when the Laurentia continent, equivalent to present North America, formed at 2.8 Ga by the collision and amalgamation of embryonal small continents. That change from Pacific-type orogeny to collision during a transitional period of 2.7-2.6 Ga, triggered the global scale mantle overturn, which increased Earth’s dynamo and contributed to the stabilization of the surface environment protected by strong Earth’s magnetism. Evidence for such a change of surface environment comes from geology, in particular, from the occurrence of stromatolites at continental margins, which provide some of the fossil evidence of oxygen and suggest that the oxygen came from photosynthesis. i.e. the amount of oxygen in the atmosphere started to increase resulting in the Great Oxidation Event at ca. 2.4 Ga

(Fig. 60). The next remarkable change of the surface environment through the whole Earth's history occurred at 2.4-2.2 Ga and is called "Snowball Earth". "Snowball Earth" means that the Earth's surface was covered by ice or frozen even in the equatorial regions. Therefore, the temperature on the Earth must be very low, close to 0 degrees globally. That event critically damaged the production of oxygen by cyanobacteria and probably primitive algae. The damage of cyanobacteria has a great impact on the evolution of Bacteria and Archaea, prokaryotes, and damaged the whole ecosystem. After the "Snowball Earth" event, the surface environment had been recovering probably until 2.1 Ga, when eukaryotes occurred and diversified (Fig. 60). Figure 59 illustrates the history of life and Earth over the past 4.6 Ga and future 1.5 b.y. (Maruyama et al., 2007). The vertical axis shows surface temperature, atmospheric composition, ocean volume and its chemistry, geomagnetic intensity, mantle temperature and composition, Earth system changes, influences from space such as cratering rates, ore formation, fossil records, and others. Other parameters which may affect the environment and illustrate its changes are geomagnetic intensity, oxygen content in the atmosphere, presence of ozone layer, depth of ocean, snowball states (full glaciations) at 2.3 Ga and at 0.8—0.6 Ga, seawater compositional change (different colors), and life evolution from prokyaryotes through eukyaryotes to metazoans. An emphasis is given to the radiation of galactic cosmic rays from Space as a cause of the Snowball Earth. The level of oxygen changed from < 10 - 3 PAL in the Archean to 10 - 2 PAL in the Proterozoic (since 2.7 Ga) and to 1 PAL in the Phanerozoic with negative spikes of superanoxia events and related mass extinctions (Komiya et al., 2008). Life has been evolving through changes in the Earth system including solid Earth and external forces from the Universe. Note the emergence of a huge landmass at the onset of the Phanerozoic, which initiated the modern-style ocean-atmosphere-landmass material circulation in which high-oxygen and nutrients supplies have been maintained. The supply of nutrients became possible after the emergence of the large landmass and metazoans, which evolved quickly in the Ediacaran-Cambrian time after the second Snowball Earth.

7.2 Proterozoic orogeny

Collision-type orogeny started at 1.8-1.0 Ga parallel to continuing Pacific-type orogeny. The surface of the Earth was occupied to 2/3 by micro oceanic plates separated by several embryonal continents surrounded by intra-oceanic arcs in addition to the Laurentia continent. During that geological period the portion of continental crust increased from 6 to 30% (Fig. 54). At about 1 Ga, the volume of continental crust started rapidly increasing. The present-day volume of the continental crust had grown by ca. 300 Ma. However, the tectonic history of the continents has not been well investigated and must be continued in future. The previous extensive study stopped because most of the orogenic belts are exposed in rather desert and difficult to access areas, e.g., in south-eastern North America, Africa and South America rainforest regions, northern Europe, central Australia, and north-eastern and central Asia. At ca. 1 Ga, the Rodinia supercontinent emerged; it was elongated and extended from east to west in equatorial regions (Figs. 53, 61).

7.3 Phanerozoic orogeny

The Phanerozoic history is closely linked with the amalgamation and breakup of the Rodinia supercontinent, which happened in a period from 1 Ga to 800 Ma or later according to different models (Hoffman, 1921; Dobretsov et al., 2003; Li et al., 2008). Rodinia assembled in the equatorial area (Fig. 61) and therefore was quite different from Gondwana and Pangea, which were placed near the northern pole (Gondwana) or clustered at both the northern and southern poles with a thin junction near the equator (Pangea). Since the pioneering work of Paul Hoffman (1991), the break-up of Rodinia, which split off Gondwana at about 540 Ma (Fig. 62), has been discussed by many scientists who produced a lot of paleogeographic reconstructions and papers discussing the formation of the Gondwana supercontinent, its break-up and final amalgamation. The Phanerozoic era spans a period from 542 Ma to the present. But its geology carries records of the famous Grenvillian orogeny with a peak at 1-0.9 Ga and related orogenic belts. The Grenvillian orogeny was followed by the Pan-African orogeny, which peaked at 600-500 Ma. The second “Snowball Earth” environmental crisis happened at 800-500 Ma (Fig. 60), which was followed by the Cambrian explosion of biota, which led to the birth and evolution of multicellular animals. First metazoan evolved from sponge at 750 Ma. Vertebrates probably appeared at the onset of the Cambrian epoch and there were nearly 35 types of vertebrates by the end of the Cambrian at 488 Ma. By the beginning of the Phanerozoic, the modern style of mantle dynamics had established. During the Phanerozoic a core of the future supercontinent Amasia formed through the formation of two world largest orogenic belts: Central Asian (dominantly Pacific-type) and Alpine-

Himalayan (dominantly collision-type) (Chapters 1, 2). Let us consider those events and epochs in more details.

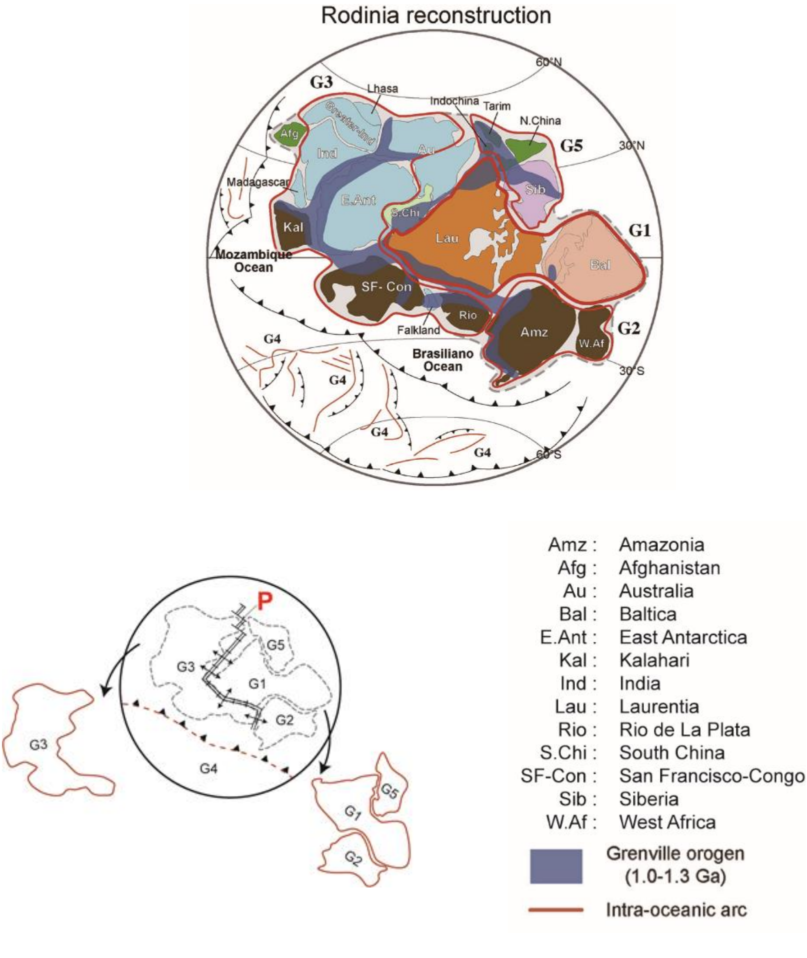

Fig. 61. The paleogeography from 1.0 to 0.75 Ga for Rodinia (modified after Rino et al., 2008). The shapes of the continents match those typically used for Pangea reconstruction. G1 to G5 are continent clusters which can be traced from Rodinia to Gondwana (see below Fig. 62).

The Grenvillian orogeny causes the formation of the supercontinents through the collision of eight continents successfully collided and amalgamated to form Rodinia along the equatorial region (Fig. 61), and all of these were characterized by continent-continent collision occurred and then to have completely different type of orogeny through the Earth's history. However, at the same time active continental margin has been formed by the Pacific type orogeny and as well as the intra-oceanic arc concentrated doubled layered mantle convection regions such as modern western Pacific regions (Fig. 62). This is the entity of the world largest scale production of continental crust since the Grenville orogeny. The Grenvillian Pacific type orogeny has been the most efficient through the Earth's history.

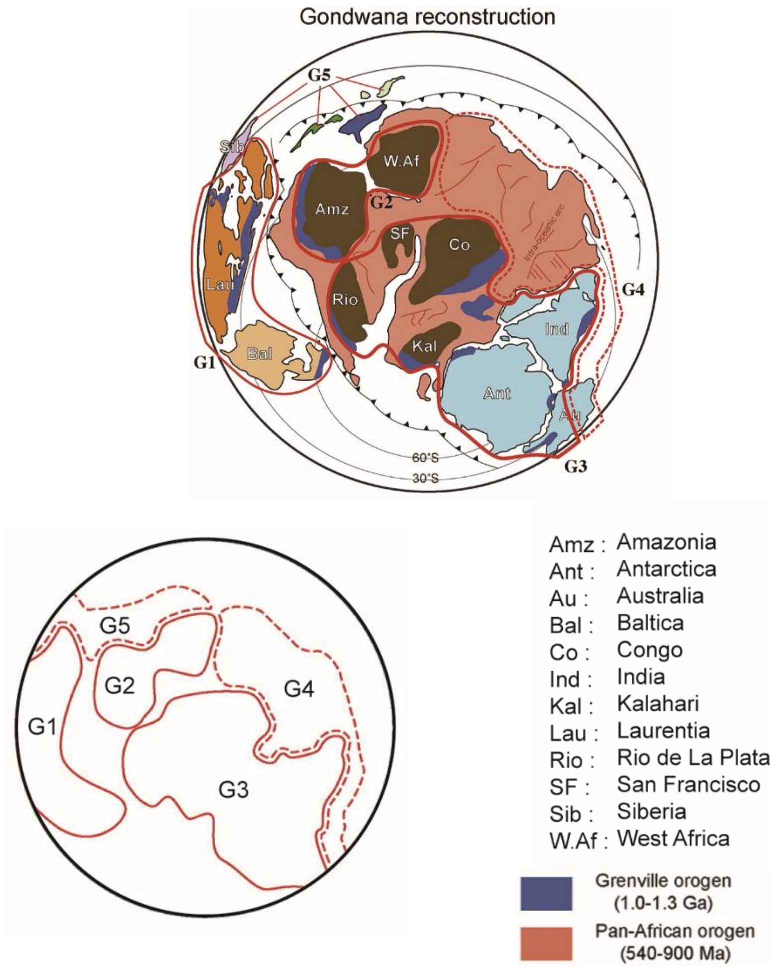

Fig. 62. The paleogeography at 549 Ma for Gondwana (modified after Rino et al., 2008). The shapes of the continents match those typically used for Pangea reconstruction.

In addition to the production of new crust, the Grenvillian orogeny led to the formation of supercontinents through continent-continent collisions: eight continents successfully collided and amalgamated into Rodinia in the equatorial region (Fig. 61). At the same time, an active continental margin formed by a Pacific-type orogeny, and an intra-oceanic arc encompassed regions of double-layer mantle convection similar to the modern western Pacific regions. The Grenvillian orogeny produced the world largest amount of continental crust through the Earth's history, presumably mainly by the Pacific-type orogenic mechanism.

The Pan-African orogeny, which created the Gondwana supercontinent, was another remarkable event (Fig. 62). Since its onset in the latest Neoproterozoic, the Pacific-type orogeny became identifiable in northern Africa and Saudi Arabia and, partly, in Australia, because those continental blocks were connected at that time. The area involved into the Pan-African orogeny comprised numerous intra-oceanic arcs, probably, more than one hundred according to the compilation of Gass (1981). Pacific-type orogenic events were responsible for most of continental crust growth in the Late Proterozoic and for the origin of supercontinents, including Gondwana and Rodinia, because the collisions of continents cannot generate new continental crust. Before the final collision of the continents, the Pacific-type orogeny was apparently a prevailing formation mechanism for supercontinents.

The orogeny that produced the supercontinent of Pangea seems to be quite different from the previous events of supercontinent origin (Fig. 47, 52). Pangea was not a supercontinent in a strict sense, because Paleo-Asia (Laurasia), its northern half, is still growing, and the growth may be completed in about 200-300 million years. The southern half of Pangea (Gondwana), has been rifting off for at least 400 Ma; it continuously drifted away to move the rifted continents all the way to the north. Therefore, the construction of the Gondwana continent is still ongoing and has not been completed yet, though the birth of Laurasia speeded it up at about 200-300 Ma ago. The true formation of a supercontinent means connection of its northern and southern halves. The apparent shape of Pangea suggests the birth of a supercontinent, but it is actually not the case considering mantle processes. To sum up, the Phanerozoic orogeny was quite different from that in the Archean or in the Paleoproterozoic, which mostly occurred as Pacific-type events. In post-Paleoproterozoic time, continental collisions became more frequent and were as significant as the Pacific-type orogeny in the Phanerozoic. However, if we look at the production rate of continental crust, the Pacific-type orogeny has contributed remarkably, while the continent-

continent collisions caused destruction instead of crust growth. In Archean time, continental crust grew rapidly, but direct arc subductions consumed TTG material, which was also the greatest in the Archean (Fig. 55). Therefore, the Pacific-type orogeny contributed a lot to the Archean continental crust growth, but most of the material had been subducted deep into the mantle. On the other hand, continent-continent collisions occurred frequently also during the Phanerozoic and produced large mountain systems such as the Himalayas and the Alps (Figs. 8, 48). The continental collisions destroyed the geological structure of the existing continents but never contributed to the net increase of continental TTG material, which was rather provided by the Pacific-type orogenic mechanism. Thus, the two types of continental orogeny have contributed to the tectonic settings of the Earth's surface in completely different ways. Considering the Archean vs Phanerozoic Earth, the production of TTG crust and direct arc subduction were apparently predominant in the Archean but were far less significant in the Phanerozoic, when the continents had bigger sizes.

In the Phanerozoic, the surface environment of the Earth stabilized and a new type of orogeny started. The Phanerozoic was the time for the loss of oceanic spaces, as a layer about 600 meters thick moved from the surface into the upper mantle. That was the fade of gradual cooling of the Earth, and the removed surface water carried by high-grade slabs (Fig. 63) settled in the mantle transition zone (Maruyama and Liou, 1998). The initiation of return-flow of seawater into the mantle led to the rapid increase of the landmass on the Earth (Maruyama and Liou, 1998, 2005). The estimated P-T conditions of more than 300 world regional metamorphic belt (Fig. 63a) clearly indicate the secular cooling of the subduction zone geotherm, which was gradual from the Archean to the Proterozoic until 1.0 Ga, but then became rapid, in particular, after 0.63 Ga. Those shifts of the subduction zone geotherm overlap in the P-T phase diagram of the peridotite + water system (Fig. 63b), which show that the depths of dehydration became deeper over time (Fig. 63c), and the water content in the hanging wall mantle wedge increased 6.5 times at ca. 630 Ma^ from 1.0 wt % to 6.5 wt % (Fig. 63b). This change was the most critical for hydrating the hanging wall mantle wedge and leading to a rapid decrease of sea level that exposed a huge landmass in the Ediacaran-Cambrian time. With cooling the hydration maximum depth at the bottom of mantle wedge propagated over time, finally reaching a depth of 410 km, which is the link point with the hydrous mantle transition zone where 5-6 volumes of surface ocean water can be stored as hydrous wadsleyite and hydrous ringwoodite. The critical point for such a transition is pass point A (Fig. 63b). Thus, the lower sea level exposed a huge landmass to produce huge amounts of sedimentary rocks that we know from surface geology (Fig. 63c). If the accelerated leaking rate of seawater continues over 1.5 Ga, the Earth will lose its oceans like the present Mars (Fig. 60). The surface record of the height of the sea level at 600 Ma is ca. 600 m above the present-day level, which means the landmass before 600 Ma must be only 5 % (Fig. 63c).

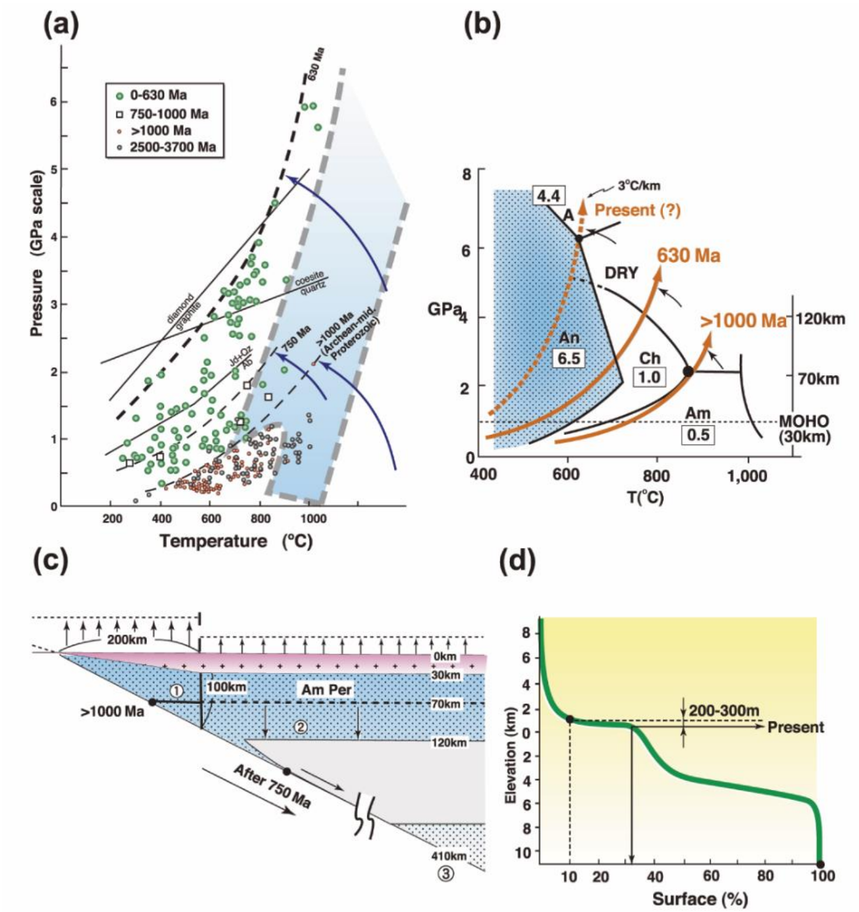

Fig. 63. Links between the initiation of return-flow of seawater into the mantle led, landmass and sea-level. A, P-T conditions of regional metamorphic belts over the world. B, 6.5 times more incorporation of water as antigorite serpentinite dominated after 630 Ma from chlorite peridotite before 1000 Ma at subduction. C, hydration of mantle wedge by serpentinization to push up the continental margin. D, the sensitivity of landmass size against sea-level. The 600 m above sea-level makes 20% decrease of the landmass (Maruyama and Liou, 1998, 2005).

There were no huge rivers and only small amounts of sediments during the Proterozoic to the Archean. Continental platforms were too deep for life to use visible light at the bottom before 600 Ma.

As a result, the temperature at a depth of 30 km (Moho depth) in the subduction zones fell to 650°C and below. Therefore, the stability field of serpentine below the Moho started to enlarge and finally reached the 410 km depth. As a result, the total land surface abruptly became 5-10% smaller at that period relative to the total surface of the Earth, though 30% larger than the modern Earth's surface. Apparently, a huge landmass appeared during the Neoproterozoic, which led to another important change: the shallow continental platform, which was deeper before, reached a depth of 200 meters. Therefore, the regions along continental margins became a paradise or an oasis for large multicellular animals and plants because rivers flowing from large continents were constantly supplying nutrients toward their mouths on the continental margins. Therefore, the Phanerozoic was the time of sedimentation and transportation of nutrients to shallow marine "greenhouse" areas.

The huge amount of deposited sedimentary rocks can be buried, and thus become protected from oxidation, including the organic matter which originally formed by photosynthesis using the sun energy as a driver of life activity (Fig. 64). The initiation of the return-flow of seawater into the mantle in the Neoproterozoic led to a sudden increase of free oxygen in the atmosphere and to the consequent emergence of metazoans (Maruyama and Liou, 1998, 2005). As shown in Fig. 63, with the secular cooling of the solid Earth through consuming plate boundaries, return-flow of seawater into the mantle began sometime around 1.0—0.8 Ga, hydrating the mantle wedge and mantle transition zone (1). The resultant drop of sea level (2) caused a rapid emergence of the landmass (3), because the average topographic elevation of the landmass is very sensitive to sea level change (Fig. 63d). A number of rivers appeared and huge rivers such as the River Mississippi first appeared on the Earth (4) with a resultant evolutional change of material circulation between ocean, landmass and atmosphere, as it is today from being an ocean planet. The continental platform changed, becoming shallow enough for life to use visible sunlight at the bottom of the continental shelf (< 200 m). This change together with the supplies of enriched nutrients from TTG continents were the most critical elements for metazoans. The increased huge landmass brought huge amounts of sediments into the ocean, which buried organic matter (5) preventing them from oxidizing. This has maintained the high oxygen content in the atmosphere since then (6). The increased oxygen level caused the births of large multi-cellular animals on the continental shelf with the help of nutrients and prepared for the most revolutionary evolution of life in the Earth's history. The increased oxygen finally reached the stratosphere to bear the ozone layer (8). This enabled life on land because ultraviolet rays were then cut by the ozone layer. Life conquered land (8).

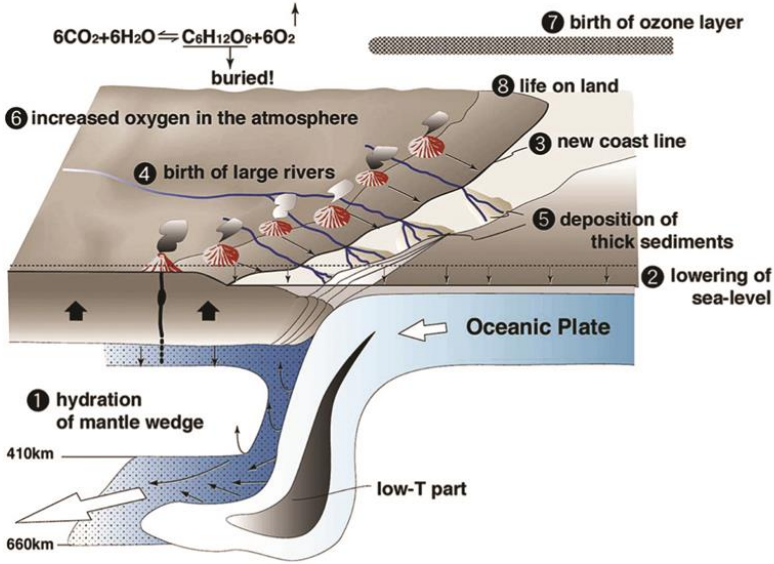

Fig. 64. Process to make the golden period of life evolution on the Phanerozoic Earth, by initiation of return-flow of seawater into mantle and essential mechanisms to increase free oxygen in atmosphere. Numbers indicate the order of genetic link from 1 to 8 (see the text).

The amount of atmospheric oxygen and the size of the ozone layer, which both existed back to 2 Ga though, increased in the Phanerozoic. It protected large multicellular animals and plants and let them conquer the surfaces of continents. The plants evolved from moss, fern, and gymnosperms in the Paleozoic to angiosperms in the Mesozoic, and finally to flower-bearing plants since the Cretaceous.

The Phanerozoic time began from the Cambrian explosion. Most of the ancestral life forms, more than twenty Phylum-level, appeared in a short time from 540 to 520 Ma. Up to 35 Phylum level metazoans appeared by the end of Cambrian at 488 Ma. Since then, diversification of animals and plants proceeded extensively by environmental and geochemical diversifications. This is due to the global scattering of nutrients by the emergence of huge landmass similar to that of the present. Note the comparison of available amount of phosphorous in the Phanerozoic and the Precambrian time, as documented by systematic difference between Precambrian and Phanerozoic sedimentary rocks. Metazoan animals first appeared at the onset of the Ediacaran period at 645 Ma on shallow marine platforms and became widespread as the area of continental platforms expanded. The first fish (known from southern China) dates back to the earliest Cambrian, as well as the famous metazoan called Anomalocaris, and other metazoan species (around 542 Ma). All kinds of vertebrates (34-38 phyla) had appeared by the latest Cambrian. An abrupt increase of nutrient supply began by the emergence of huge landmass, because the amount of water in the mantle wedge must have increased from 1.0 to 6.5 wt%, if subduction zone geotherm began to cut the boundary of the stability field between clinochlore peridotite and antigorite peridotite during the cooling (Fig. 63b). This was due to the initiation of return-flow of seawater into the mantle during the Neoproterozoic.

As plants emerged on the land, the amount of free oxygen in the atmosphere became immediately one order of magnitude greater through photosynthesis. The process was concurrent with large-scale deposition thus making dynamic equilibrium of material input and output, or a biological pump. For example, if all free oxygen generated by photosynthesis was consumed by back reactions, the content of atmospheric oxygen would never increase. Moreover, maintaining high free oxygen in the atmosphere requires the presence of large plants on continents, while dead plants become buried and degrade to organic matter in voluminous sedimentary rocks produced by deposition. These processes have to be coupled to maintain dynamic equilibrium. Free oxygen is not a dynamically stable phase: it oxidizes surface rocks (weathering) and organic matter, i.e., the respective amount of buried organic matter is necessary. This is the cause of the oxygen pump, which was working continuously through the Phanerozoic. Large multicellular animals probably initiated the epigenetics in genome biology, which critically changed their lifestyle. The oxygen-producing pump effectuated by plants controlled the appearance of metazoan animals and the diversification of species living on the land, which depended on the activity of photosynthesis and the production rate of free oxygen. Metazoans evolved and their life became completely different from the previous single-cell life. Most of metazoan animals have coexisted with bacteria inside their bodies as symbionts.

The processes at the dawn of Phanerozoic may have been as follows (Fig. 64). The initiation of the seawater return flow into the mantle caused mantle wedge hydration and the ensuing sea-level fall.

Correspondingly, the coast line moved oceanward thus increasing the size of landmass where huge rivers appeared and carried large volumes of sediments, while voluminous organic matter photosynthesized by algae and cyanobacteria became buried. The burial of organic matter maintained high contents of atmospheric oxygen by preventing it from oxygen-consuming back reactions. The increased oxygen in the atmosphere finally diffused upwards to create the ozone layer, which shielded the incoming ultraviolet radiation from the Sun, thereby making the land habitable for plants and animals. First, cyanobacteria invaded swamps along rivers and around lakes (Gray, 1993). They had gradually evolved to algae, bryophytes and to tracheophytes by the Late Devonian (Scheckler, 2001).

Thus, it became possible to keep a constant high oxygen level in the Phanerozoic: from 10-35% in the past to 21% at present. It has been working well, except the time of five big environmental changes, when the oxygen content decreased. Metazoan animals and plants coexisted and evolved rapidly for almost 500 million years, but a dramatic change occurred during the snowball-Earth event before the onset of the Phanerozoic. The ability of photosynthesis in plants was severely damaged at least several times during the Snowball Earth period. The rapid decrease of free oxygen was followed by a rapid increase right after the end of the Snowball event. Those processes repeated three to four times during the Snowball period and critically affected on-land plants and animals (Maruyama et al, 2007 a). Such repeated dramatic up-down environmental changes were triggered, among other causes, by sea water flow from the surface into the mantle (Maruyama and Liou, 1998, 2005). The continental crust growth in the late Neoproterozoic preceeded the respective sea level change: the percentage of continental crust increased from 15% to 95% between 1.5 and 1.0 Ga (Fig. 63). The crustal growth was possible only by the Pacific-type orogeny, which played a key role as it provided nutrients stored in the continental crust, e.g., phosphorus, which is concentrated in granitic continental crust but lacks in the mantle, is important for tabulates having their bones mainly composed of phosphorus.

According to a P-T plot for regional metamorphic belts worldwide since the Archean, the first appearance of blueschist was ca. 700 Ma (Nakajima et al., 1990), and subduction zones rapidly became cooler at the onset of Phanerozoic (Fig. 63a; Maruyama et al., 1996; Maruyama and Liou, 1998). The temperature at the Moho depth was above 600°C before 700 Ma, but rapidly changed to below 600°C thereafter, and down to 200°C at present (Chapter 6). Therefore, the seawater return flow into the mantle began in the latest Proterozoic, as estimated by the phase diagrams; the mechanism is sketched in Fig. 63c (Maruyama and Liou, 2005). The observed sea-level drop clearly supports the idea, and the proposed sea-level change curve (Maruyama et al., 2013) shows that a ca. 600 m deep ocean was removed from the surface into the mantle, to the 410-660 km depth transition zone which can store about 5 times the total mass of water in the surface oceans (Maruyama and Liou, 2005).

7.4 Links to mantle dynamics

In this section, we will discuss the fate of TTG material tectonically eroded or subducted into the mantle (see also Chapters 5 and 6). Presumably, the amount of TTG crust transported into the deep mantle was more than ten times greater than the total mass of the surface continental crust. For example, if the target depth is restricted to the lower part of the mantle transition zone at depths of 520 to 660 km, the amount of subducted TTG must be at least 6 to 7 times higher than the present crust judging by the difference between P and S wave velocities in the mantle transition zone and by the PREMA world standard seismic velocities. TTG material is present in the upper mantle or in the topmost lower mantle, for example, under eastern Asia where it forms second continents. TTG present in the topmost lower mantle can be segregated from the slab. The second continents that finally accumulate at the bottom of the upper mantle at about 660 km represent a thick granitic mass, which physical parameters are different from those of the olivine-dominated mantle. The second continents can behave as rigid masses that impede slab penetration. A descending old slab may force rifting and move a second continent horizontally, but not up and down, because of the gravitational stability at the upper mantle bottom. Therefore, this is a kind of continental rifting but at great depths below the surface. On the other hand, a mantle plume reaching the mantle transition zone can cause break-up of second continents. Those processes could be similar to the rifting, dispersal and amalgamation of the continents existing on the Earth's surface.

The distribution of the second continents within the solid mantle could be another important research target, particularly, under Asia, which is the most important characteristic focus region. There may be more than several tenths of second continents (Figs. 42, 45, 65). One more issue is their dynamics and supercontinent cycles in relation to vertical super-upwelling, like the African and Pacific superplumes, which can be combined to discuss the global time and space dynamics in the solid mantle. In this paper, we have shown a preliminary global-scale pattern of 26 distinguished second continents which are extremely unevenly distributed (Fig. 45). There are eastern Asian continents, with the western Pacific triangular regions extending as far as New Zealand and Australia, and another region from Alaska to the northern Hawaii and northern California, as well as the polar oceanic areas. However, second continents are not seen in the seismic tomography images underneath major oceans: Pacific, Indian and Atlantic oceans. That heterogeneous distribution of second continents suggests that supercontinents can form in different geological periods and undergo rifting, dispersal, and amalgamation.

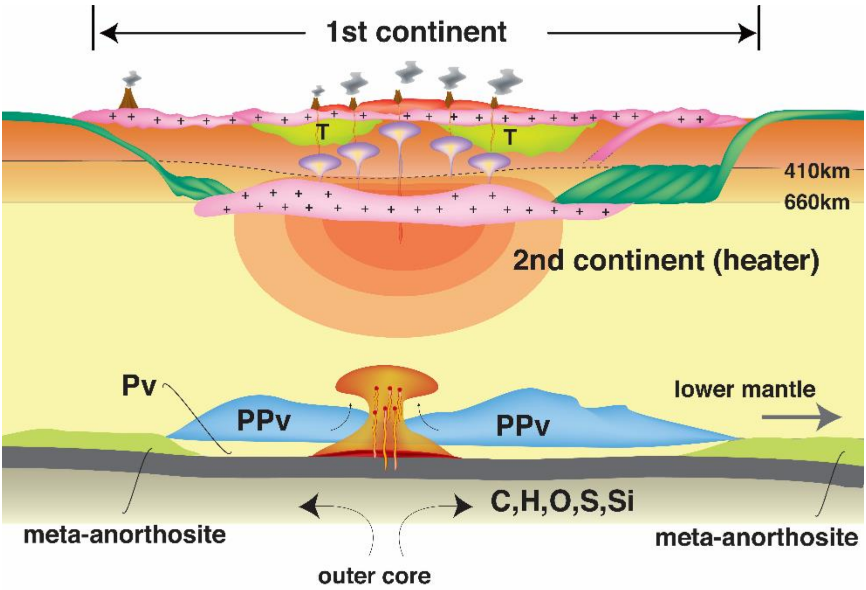

Fig. 65. The role of second continents through the Wilson cycle (modified after Senshu et al., 2009). Accumulated TTG in the mantle transition zone under the supercontinent would warm the overlying wet mantle transition zone to generate hydrous plumes that break up the amalgamated supercontinent, and lead to the birth of a superplume at the core-mantle boundary (Maruyama et al., 2007a, b). If this is the case, the top controls the bottom of the mantle in mantle dynamics.

The second continents dominated by granite enriched in radiogenic elements (U, K, and Th) may also control rifting, dispersal, and amalgamation of the surface supercontinents and may be the most probable major cause for these cycles (Fig. 44). The secular formation of supercontinents on the Earth enriches the upper mantle. The supercontinental cycles are driven by superplumes triggered by the melting of mid-ocean ridge basalt (MORB) right above the CMB. In time, MORB underneath a supercontinent may become segregated and enriched in light elements by the heat released from the metallic core. This may lead to partial melting producing heavy melts that reside right above the CMB, while the MORB residue may rise by buoyancy. Small superplumes generated at the CMB can amalgamate into a bigger superplume. Thus, it is a composite plume, or a superplume, that can break down the surface of a supercontinent (Maruyama et al., 2009).

However, a second continent present on the bottom of the upper mantle just under a surface supercontinent can generate heat, which will be more efficient to break it down. The second continents can be more effective in breaking up supercontinents because basalt contains far smaller concentrations of radiogenic elements than granite. Moreover, the second continents may control not only the upper mantle, but also the whole mantle convection. The heat they release can maintain the birth of plumes at the 410 km depth, at the top of the mantle transition zone. The magma and plumes that can break up the surface supercontinents may form with participation of fluids produced by dehydration reactions. Upwelling mantle plumes may merge to form a superplume and related mantle upwelling. The relationships between smaller plume and superplume upwellings could be another target to be clarified.

7.5 Mantle dynamics and evolution of continental crust through Earth's history

The Pacific-type orogeny must have contributed critically to the origin of large multicellular animals at the dawn of the Phanerozoic. On the other hand, the common sense is that all continental crust had accumulated on the Earth's surface as a result of granite formation during the fade of planet cooling. Yet, nobody has suspected before that granitic crust and, presumably, second continents or third continents formed in the latest Hadean, about 4 Ga. Both the second and third continents may have been the most important control of the mantle convection and evolution through time, but nobody has ever discussed that possibility. The evolution of the solid Earth has long been considered in terms of circular cooling, while the mantle convection and the plume activity were interpreted as responses to that cooling and transfer of internal heat to the space. This is a general idea of the evolution of solid planets: mantle convection as a response to cooling from the space. No one has discussed the role of subducted continental crust in mantle dynamics, the reason why felsic magma, once generated, rose from the mantle through crust and the buoyant material floated on the surface of the Earth. Nobody has suspected, however, as it is discussed in this book, that a total mass an order of magnitude bigger than the surface continents can form huge bodies at a given depth within the mantle, and that the respective supercontinent cycles, with rifting, dispersal, and amalgamation, can occur at that depth. The self-heating ability drives the upper mantle convection, and it presumably works in the lower mantle as well. When the Earth's magma ocean solidified, the thick surface crust enriched in KREEP elements (potassium, rare earths, and phosphorus), together with anorthosite, must have sunk to the mantle bottom, and radiogenic elements such as uranium, thorium and potassium must have enriched the KREEP-bearing magma present under the lower crust or under the primordial continental crust. The primordial continents must have gone by the latest Hadean: possibly, they dropped onto the metallic core and played a critical role as a heat source at the mantle base. Nevertheless, the amount of granitic material increased through the geological time. That material accumulated at the bottom of the upper mantle through tectonic erosion and segregation during subduction to the mantle transition zone. We discussed the amount of continental crust growth in the previous sections. The surface of continental crust in the early 2 Ga half of the Earth's geological history occupied approximately 18% of the present continental masses. Therefore, it was a nearly continent-free water planet, and the climate system must be remarkably different from that in the Phanerozoic. The lack of coarse-grained sedimentary rocks for the first 2 billion years precluded the oxygen increase in the atmosphere (Figs. 64). There were only restricted small stable greenhouse regions suitable for life. At 700 Ma, stromatolites started to develop on a rather global scale and became wider spread by 600 Ma, to control the stable surface environment of the Earth. The stable platform environment formed due to the increasing volume of continental crust, as well as the total volume of ocean (4-5 km thick).

The landmass increased upon the sea level fall caused by the return flow of seawater since ca. 800 to 600 Ma ago. At that period, extensive radiation of the galactic cosmic rays caused abrupt deep and extensive cooling (snowball Earth) and led to mass extinctions which reduced the population of living organisms presumably by a factor of 7-8. Glaciations occurred periodically from 780 million years to 488 million years (latest Cambrian), but predominated in the pre-Ediacaran time which ended at 635 Ma. That period was very hard time for survival on marine platforms and partly on the land. The snowball Earth event was followed by accelerated life evolution, also called Cambrian explosion. It should be noted that the birth of metazoans and the explosive evolution were possible only due to sufficient supply of nutrients from the granitic basement rocks (Fig. 60).

Considering the history of the continental crust, we note that the Hadean primordial continents are 4.6-4.0 Ga. Greenland is the oldest and huge geological body of a 3.8-3.9 Ga age (Fig. 57). On the other hand, the Earth's oldest 4.3-4.2 Ga zircons, which are grains 0.1 to 1 mm in diameter, have been found in the Acasta gneiss of central Canada (Izuka et al, 2009). The very old zircons were also retrieved from Early Proterozoic sedimentary rocks of western Australia, which are mostly younger than 3.8 Ga but some have 4.4-4.37 Ga cores; their host rocks could be granitic melts judging by the oxygen ratio (Wiley et al., 2001).

Many scientists believe that the Hadean Earth was totally covered by a magma ocean, which is a straightforward implication of the planetary formation theory. Our understanding of the primordial continent is developed from the geology of the Moon, and also from the concept of a giant impact. The Moon surface is covered by 50-70 km thick anorthositic crust with a local cover and dikes of KREEP basaltic composition, and is presumably underplated with KREEP-like rock types beneath the anorthositic continent. Both rock units have been interpreted as the final residue of a magma ocean when the Moon was formed by the giant impact. The same impact led to the formation of the early Earth (e.g., Canup, 2004) by collisions of several Mars-sized protoplanets (e.g., Chambers and Wetherill, 1998; Genda et al., 2012). In accordance with the giant impact theory, the Earth must have been completely molten as deep as the core (Tonks and Melosh, 1992). During the gradual cooling of the Moon and the Earth, the final liquid remained near the surface forming the buoyant anorthositic crust, covered or underplated by KREEP magma similar to that observed on the Moon. If this scenario is correct, the Earth developed primordial continents that are similar in composition and possibly in amount to those of the Moon. However, there is no direct evidence for the presence of primordial continents on the present Earth's surface. Therefore, we infer that the primordial continents must have subducted and sunk into deep mantle in the Hadean. If the primordial continent was present on the Hadean Earth, nutrients could be supplied to the life birthplace (Maruyama et al., 2013).

Recent models predict the existence of four of five planet-size orbiting bodies at the final stage of the Earth formation, which collided repeatedly. One of them, called Theiya, collided to form the primordial Earth together with the Moon. That giant impact must have completely melted the mantle and the core. Consequently, the Moon became covered by buoyant cumulate composed of anorthosite together with KREEP and other rock types as minor components. In the case of the Moon, unlike the Earth, which has a large gravity, field, quite different rock types were formed, which consisted of plagioclase equilibrated with magma. In the case of the Earth, there is a density crossover at a pressure around 7 kbar. For example, some rocks at mid-ocean ridges contain plagioclase as phenocrysts, but without olivine. Experiments that reproduce conditions of partial melting in the upper mantle show that olivine is the first liquidus phase that forms before plagioclase. However, plagioclase sinks above the 12 km depth at higher pressures corresponding to the density crossover, but it can float below that depth (Kushiro and Fuji, 1977). Therefore, at a depth of 12 km, single-mineral plagioclase rocks can be a major rock type, because plagioclase sinks down above and comes up below the depths of 12 km. In the case of the Moon, mafic minerals always prefer magnesium first at high temperatures. The enrichment in iron comes through the solidification of magma. The iron-laden melts must be heavy, and anorthosite must start to go up through time in the course of magma solidification, to come eventually to the surface, float on the Moon, and cover it totally. In the case of the Earth, an anorthositic body could have gradually ascended to the surface as well, but a part of anorthosite could have descended by convection in the magma ocean and underwent metamorphism to form a coesite + kyanite + quartz assemblage. Coesite can convert to stishovite at a depth of 270 km. With the ~4.2 g/cm3 density of stishovite, the average rock density would be 3.9 g/cm3. The meta-anorthosite bodies may have sunk to depths of 660 km (Fig. 66).

Plate tectonics made the formation of granite by the ca. 20% partial melting of MORB possible. After the collapse of primordial continents at the mantle base and the onset of plate tectonics, mid-oceanic ridge basalts with the content of FeO2 2-4% higher than in the modern MORB formed a 20-25 km thick oceanic crust as the potential temperature of the mantle was 200 K higher then today. The estimated potential temperature of the source mantle of Isua MORB (Archean) is approximately 1480°C. The higher mantle temperature and thinner oceanic lithosphere in the Early Archean may led to the formation of tholeiitic oceanic plateau-type magmas at low pressure/high degrees of melting (Komiya et al., 2004). The melting produced iron-rich granitic melts during the late Hadean to the Archean. The subducted slabs must have accumulated at the depth of 660 km and were probably dominated by basaltic residue. The phase transformation of those metabasalts carrying anorthosite to the 660 km depth could force the transportation of anorthosite to the lower mantle bottom, but a part of them could have

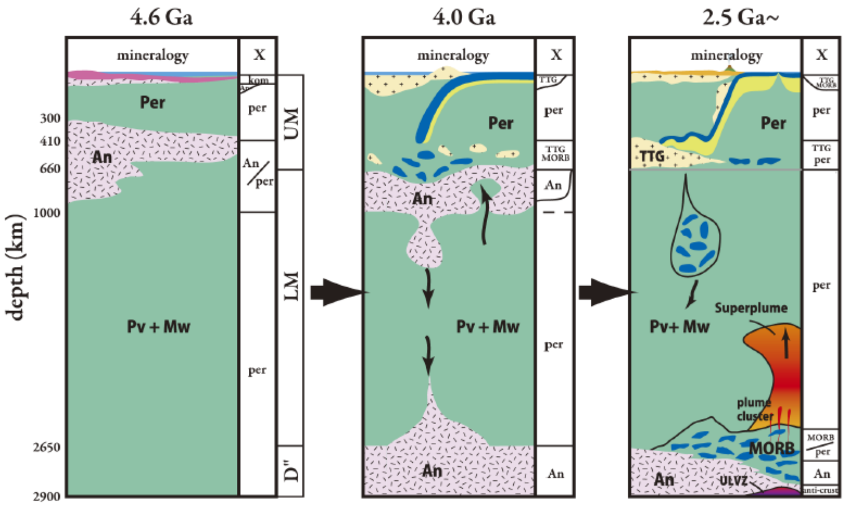

Fig. 66.The consolidation of magma ocean (Kawai et al., 2009). A thick anorthosite layer developed at the bottom of the upper mantle after the consolidation of a magma ocean at 4.6 Ga by phase changes provided by mantle convection (left). The surface of the Earth was covered by a komatiitic skin with a thin anorthosite layer. The birth of a primordial ocean promoted plate tectonics and supplied MORB to the MTZ, which in turn brought the anorthosite layer into the lower mantle down to the CMB, because MORB has higher density at the bottom of the MTZ (middle). Formation of TTG along consuming plate boundaries in Archean time brought abundant TTG into the MTZ to develop second continents. Continued plate tectonics transported huge amounts of MORBs into the mantle, accumulated slabs at the CMB and partially melted those in the D” layer above the CMB to give rise to a superplume enriched in MORB restites, leaving a FeO-rich melt at the CMB to grow anticrust (Maruyama et al., 2007b). After ca. 1.8 Ga superplume cycle began. The second continents grew preferentially in the Archean at the bottom the upper mantle. The volumes of TTG material in the MTZ is now 10 times bigger than that of the surface continents on the Earth.

returned to the bottom of the upper mantle. However, once they were brought to at least the mid-lower mantle, Ca-perovskite became denser than the surrounding mantle and could drop to the bottom of the mantle, to the D” boundary (Fig. 66). However, meta-anorthosite hardly could be brought to the D” layer above the CMB because it is not much denser than the surrounding mantle peridotite. However, up to 25% of the lunar crust is, in places, composed of Fe-rich KREEP. Those KREEP-type abnormally Fe-rich rocks formed the lower mafic crust of the primordial Earth after the magma ocean had solidified. Those rocks definitely could sink as deep as the D” layer and become super-enriched in radiogenic elements: U, K, and Th. The thermal history of the Earth is related to super-enriched radiogenic elements accumulated right above the metallic core. Simple calculations show that several millions of years may be enough for the magma ocean to solidify completely. Thus, the magma ocean had been completely solidified soon after the giant impact.

Recent seismic measurements right above the CMB image possible remnants of the magma ocean at the bottom of the lower mantle, at the D” layer. If so, we need a heat source to generate an ultra-low velocity zone composed of melts, probably enriched in iron. Thus, gravitationally stable melts may be present on the bottom of the mantle. The melts were previously interpreted as recycled MORB, but now we would also emphasize the importance of KREEP-type basalts, which existed on the surface long ago, when the magma ocean was solidifying, and then moved down to the bottom of the mantle. The onset of plate tectonics was triggered by the birth of a primordial ocean. It developed at least since 4.3 Ga to form numerous intra-oceanic arcs, while direct slab melting produced TTG continental crust. However, most of material must have subducted directly to the upper mantle bottom, with rare exceptional cases of parallel arc-arc collisions which produced embryonic elongated irregularly shaped continents. However, most of those intra-oceanic arcs subducted directly to develop second continents at the bottom of the upper mantle, which was also promoted by double-layer convection until 2.7 Ga, when the mantle overturn occurred (see Chapter 6 for details). By the end of the Archean, large amounts of TTG material presumably had accumulated on the bottom of the upper mantle and controlled its dynamics (Fig. 66).

Conclusions and Acknowledgements

This book has evolved from a land-mark paper by S. Maruyama, S. Omori, H. Sensu, K. Kawai and B. Windley "Pacific-type Orogens: New concepts and Variations in Space and Time from Present to Past" published back in 2011 in Journal of Geography, however, in Japanese with English abstract and captions. We give our credit and thank all the authors of that paper.

Since then, the first author has been dreaming to publish those materials in English, in particular, because much more new data from geochronology, petrology, numerical modelling and geophysics have appeared. This became possible after Shigenori Maruyama, Inna Safonova and their colleagues from the Novosibirsk State University and from the Sobolev Institute of Geology and Mineralogy Siberian Branch of the Russian Academy of Science, received megagrant from the Ministry of Education and Science of the Russian Federation on Project "A multidisciplinary study of Pacific-type orogenic belts and development of a holistic model linking evolution of oceans, their active margins and mantle magmatism" (no. 14.Y26.31.0018).

The original paper and this book could not be possible without great contributions of Akiho Miyashiro and John F. Dewey to Earth's Sciences. Shigenori Maruyama enjoyed discussions with both of them in past times and greatly appreciate both, in particular, A. Miyashiro, who encouraged him all the way during the last 30 years. Big thanks also should go to Akira Hasegawa and Dapeng Zhao from the Tohoku University, Sendai, Japan, with whom S. Maruyama also has had many fruitful discussions and who showed and shared published and unpublished data sets on seismic tomography. Finally, S. Maruyama would like to thank very much his colleagues and his supervised students from the University of Tokyo and the Tokyo Institute of Technology, where he has been working over the last 25 years. Inna Safonova acknowledges support from the State assignment of the Sobolev Institute of Geology and Mineralogy SB RAS.

Both authors strongly appreciate all the members of the Laboratory of Evolution of Paleo-Oceans and Mantle Magmatism (LEPOM), which was established at the Geology-Geophysical Faculty, Novosibirsk State University, in the frame of the Megagrant Project. Our colleagues have been permanently encouraging our activities and contributing to our research project. In particular, we cordially thank Project participants Sergey Krivonogov, Alexandra Gurova, Maria Kolesnichenko, Maria Shapovalova and Ruslan Chyorny for typing the text, drafting the figures and critical reading and editing the original and many intermediate versions of the book.

We acknowledge the permanent support and patience of Valery Vernikovsky, the Dean of the Geology-Geophysical Faculty, and the whole and staff of the Faculty, in particular, the Chair of Petrography and Geology of Ore Deposits led by Andrei Izokh. Our cordial thanks finally go to Sergey Galushko, Dmitry Churkin and Sergey Malinovsky, the Vice-Rectors, and to Mikhail Fedoruk, the Rector of the Novosibirsk State University.

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Abbreviations

AC – accretionary complex
An – anorthosite
BMW – big mantle wedge
BS – blueschist
CAOB – Central Asian Orogenic Belt
Chl – chlorite
CMB – core-mantle boundary
C-type – collision-type
DHMS – dense hydrous Mg-silicates
DSDP – deep-sea drilling program
EC - eclogite
EM – Enriched Mantle source
HP-LT– high pressure-low temperature
HREE – heavy rare-earth elements
Hz – harzburgite
Kom – komatiite
LA-ICP MS – laser ablation inductively coupled plasma mass spectrometry
LILE – large ion lithophile elements
LREE – light rare-earth elements
MMF – metasomatic-metamorphic factory
MOR – mid-ocean ridge
MORB – mid-ocean ridge basalt
MTZ – mantle transition zone
OIB – ocean island basalt
OPS – ocean plate stratigraphy
Opx – orthopyroxene
Per – peridotite
PREM – preliminary reference Earth model
PTO – Pacific-type orogeny
P-type – Pacific-type
Pv – perovskite
SEM – scanning electron microscopy
SHRIMP – sensitive high-resolution ion microprobe
Srp – serpentine
SZMF – subduction zone magma factory
Tlc – talc
TSB – trench-slope break
TTG – tonalite-trondhjemite-granodiorite association
TTL – Tanakura tectonic line
UCSS – upper crust shelf sediments
UHP-HP– ultra high pressure-high pressure
UHTP – ultra high temperature and high-pressure

Научное издание

Маруяма Шигенори,

Сафонова Инна Юрьевна

OROGENY AND MANTLE DYNAMICS: ROLE OF TECTONIC EROSION AND SECOND CONTINENT IN THE MANTLE TRANSITION ZONE

Подписано в печать 17.12.2019 г.
Формат 60х84 1/16. Уч.-изд. л. 13. Усл. печ. л. 12,1.
Тираж 300 экз. Заказ № 354
Издательско-полиграфический центр НГУ.
630090, г. Новосибирск, ул. Пирогова, 2.