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Research Article

Thermal consequences of the formation of a slab window beneath the Mid-Cretaceous southwest Japan arc: A 2-D numerical analysis

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Abstract The development of voluminous granitic magmatism and widespread high-grade metamorphism in Mid-Cretaceous southwest Japan have been explained by the subduction of a spreading ridge (Kula–Pacific or Farallon–Izanagi plate boundaries) beneath the Eurasian continent and the formation of a slab window. In the present study, the thermal consequences of the formation of a slab window beneath a continental margin are evaluated through a 2-D numerical simulation. The model results are evaluated by comparison with the Mid-Cretaceous geology of southwest Japan. Of particular interest are the absence of an amphibolite- to granulite-facies metamorphic belt near the Wadati–Benioff plane, and significant melting of the lower crustal-mafic rocks sufficient to form a large amount of granitic magma. Because none of the model results simultaneously satisfied these two geological interpretations, it is suggested that subduction of plate boundaries in Mid-Cretaceous southwest Japan was not associated with the opening of a slab window. According to previous studies, and the results of the present study, two different tectonic scenarios could reasonably explain the geological interpretations for Mid-Cretaceous southwest Japan: (i) The spreading ridge did not subduct beneath the Eurasian continent, but was located off the continental margin, implying the continuous subduction of very young oceanic lithosphere; (ii) ridge subduction beneath the continental margin occurred after active spreading had ceased. Consequently, in both tectonic scenarios, the subduction of plate boundaries at the Mid-Cretaceous southwest Japan was not associated with a slab window, but very young (hot) oceanic lithosphere.

Key words: Mid-Cretaceous igneous activity, Ryoke–Sanyo granitoids, Sambagawa Metamorphic Belt, slab window, spreading-ridge subduction, thermal modeling.

INTRODUCTION

It is generally accepted that high-temperature (T) metamorphic belts associated with the production of great quantities of granites at convergent plate margins, as typified by the Circum-Pacific regions of southwest Japan and North America in the Mid-Cretaceous and Early Cenozoic, are indicative of the ancient subduction of 'hot' oceanic lithosphere (e.g. Uyeda & Miyashiro 1974). The Cretaceous and Early Cenozoic are also characterized by Farallon–Izanagi or Kula–Pacific spreading in southwest Japan (e.g. Engebretson *et al.* 1985; Maruyama & Seno 1986; Fig. 1) and Pacific–Farallon spreading in North America (e.g. van Wijk *et al.* 2001). It is conventionally thought that ridge subduction is required to form such high-T metamorphic belts, because petrologic estimation of the metamorphic field gradient of high-grade (amphibolite- to granulite-facies) metamorphic belts cannot be explained simply by the thermal structure of the arc-trench system in 'normal' or

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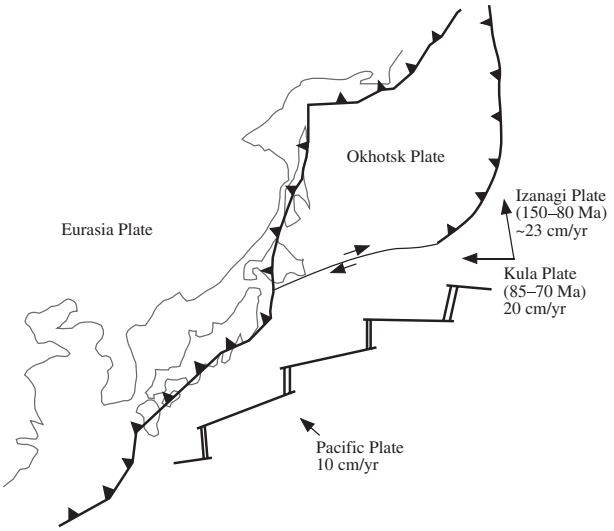

Fig. 1 Paleogeographic reconstruction of the Eurasian continental margin area during the Late Cretaceous (simplified from Maruyama & Seno 1986, with permission). The Kula–Pacific ridge is located off the Eurasian continental margin.

steady-state subduction (Peacock *et al.* 1994; Iwamori 2000; Okudaira 2002).

Along the Mid-Cretaceous Eurasian continental margin, the subduction of a ridge is interpreted by many geological observations in southwest Japan, including the voluminous along-arc activity of the Ryoke–Sanyo granitoids and *in-situ* basalts with mid-ocean ridge basalt (MORB) affinity within Cretaceous accretionary complexes (Banno & Nakajima 1992; Kiminami *et al.* 1993; Kinoshita 1995, 2002; Nakajima 1996; Brown 1998). Based on these geological circumstances, Iwamori (2000) suggested that the formation of the paired metamorphic belt of southwest Japan; that is, the high-pressure (P) Sambagawa Metamorphic Belt (oceanic side) and the low-P Ryoke Metamorphic Belt (continental side), occurred as a result of ridge subduction at the Cretaceous Eurasian continental margin. However, this model in fact represents the subduction of a very young slab rather than an active spreading-ridge, because the spreading rate of the ridge in his model is zero.

When a slab window opens during ridge subduction, the asthenosphere is assumed to rise and fill the space formerly occupied by the slab. If such a high-T section had indeed been subducted along the Wadati–Benioff plane, the forearc region, in particular near the Wadati–Benioff plane, would exhibit extreme thermal perturbations with the resultant formation of an amphibolite- to granulite-facies metamorphic belt. However, the rocks of the Sambagawa Metamorphic Belt, which is considered to be formed near the Wadati–Benioff plane during subduction of the plate boundary beneath the Mid-Cretaceous Eurasian continental margin, do not indicate such low-P/high-T metamorphic conditions (e.g. Banno & Nakajima 1992; Enami 1998; Aoya *et al.* 2003). For the subduction-type Sambagawa metamorphic rocks, most of the prograde P–T paths proposed so far indicate that metamorphic pressures (reflecting the depth of the rocks) increase with increasing temperatures (e.g. Enami 1998; Aoya *et al.* 2003). Because such P–T paths of the rocks would represent their subduction-stage tectonic history, it is believed that they were located near the Wadati–Benioff plane during their subduction stage.

Based on the shape of metamorphic P–T paths for the low-P/high-T Ryoke metamorphic rocks, Brown (1998) suggested that the low-P/high-T metamorphism was caused by ridge subduction. His ridge subduction model for the Ryoke Metamorphic Belt required that the juxtaposition of the high-P/low-T Sambagawa Metamorphic Belt against it was a result of terrane amalgamation, and the paired metamorphic belt of the Ryoke and Sambagawa Metamorphic Belts represents laterally contemporaneous terranes, rather than across-arc components of a trench/arc 'paired' system. If this interpretation were true, the petrologically derived P–T conditions of the Sambagawa metamorphic rocks could not be used to evaluate the simulated results. However, tectonic processes forming the paired metamorphic belt in southwest Japan have not been understood fully (e.g. Brown 2002; Iwamori 2002), and further geological information about them is required. In the present study, we adopt the traditional view that the paired metamorphic belt for the Ryoke and Sambagawa Metamorphic Belts is representative of the geological situation in southwest Japan during the Mid-Cretaceous (e.g. Uyeda & Miyashiro 1974; Banno & Nakajima 1992; Iwamori 2002); we adopt this stance because lateral displacement along the major fault (Median Tectonic Line) during the Late Cretaceous has been estimated to be ~500 km, which is less than the lateral extent of the Ryoke Metamorphic Belt (~700 km) (e.g. Yamakita & Otoh 2000).

The thermal consequences of the formation of a slab window are not yet fully understood, although the subduction of very young (hot) oceanic lithosphere has been investigated in this regard (e.g. DeLong *et al.* 1979; Peacock *et al.* 1994; Iwamori 2000; van Wijk *et al.* 2001; Miyazaki & Okamura 2002). In the present study, we used 2-D numerical

modeling to evaluate the thermal structure beneath an arc-trench system that arose because of the subduction of a spreading ridge. The analysis focuses on the role of model parameters, such as the convergence velocity of a spreading ridge and the spreading rate at the ridge center, in the development of the thermal structure beneath the arc-trench system. Results of the model are discussed in terms of geological interpretations used in southwest Japan, notably: (i) The extent of development of the amphibolite- to granulite-facies metamorphic rocks near the Wadati–Benioff plane; and (ii) the time scale of partial melting of lower crustal rocks needed to promote long-lasting granite activity.

MODEL

The setup for the numerical model used to calculate the thermal structure beneath an arc-trench system is shown in Figure 2. In the model, a 300×200-km finite-difference grid is set as the model domain (subduction domain). Before subduction of the oceanic lithosphere in the subduction domain, the thermal structure of the oceanic lithosphere is calculated independently using a different model domain of 2000×70 km (ocean domain). In the subduction domain, the subduction is assumed to continue for 70–100 My, using a set of boundary conditions calculated in the ocean domain. The subducting oceanic lithosphere has been modeled to be in contact with the continental lithosphere and mantle asthenosphere along the Wadati–Benioff plane.

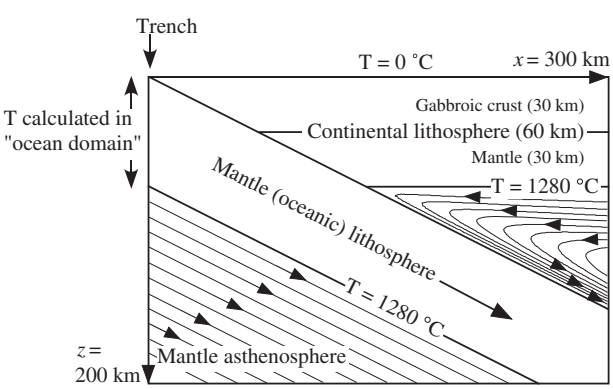

Fig. 2 Kinematic model used for calculation of temperature in the subduction domain. Arrows show the direction of subduction of the oceanic lithosphere and the direction of viscous flow in the mantle asthenosphere.

The heat-transfer equation is solved to obtain the variation in temperature (T) with distance from the trench (x) and depth below sea level (z), as a function of time (t):

$$\rho c\,(\mathrm{d}T/\mathrm{d}t) = \kappa\nabla^2 T - \mathbf{v}\cdot\rho c\nabla T + Q \qquad (1)$$

in which k is thermal conductivity, ρ is density, c is heat capacity, $\mathbf{v}$ is velocity and Q is the heat production rate. On the basis of the equation and conditions described below, the thermal structures of both simulation domains are calculated using an explicit finite difference method ($\Delta x = \Delta z = 1000$ m; $\Delta t = 3.15 \times 10^9$ s).

In the ocean domain, the thickness of the mantle lithosphere is set constant at 60 km, considering the known depth of the lithosphere–asthenosphere boundary (~65 km) and its invariance with age (Hirth & Kohlstedt 1996; Karato & Jung 1998). The initial (pre-spreading) thermal structures of the oceanic lithosphere are approximated using a 60-km-thick plate-model with a basal temperature of 1280°C, equivalent to a mantle heat flow of 0.058 W/m^2. An adiabatic gradient of ~0.3°C/km is incorporated into the vertical temperature structure in the mantle asthenosphere. Values of thermal conductivity, radiogenic heat production, rock density, and heat capacity of the mantle lithosphere and mantle asthenosphere are shown in Table 1.

The spreading-ridge is modeled as follows. A region with a constant asthenospheric temperature (1280°C) within the oceanic lithosphere beneath the ridge center is considered. The shape of the high-temperature region is assumed to be triangular, with a basal length of 2 km and a

Table 1 Physical properties used in this study

	Thermal conductivity (W/m/K)	Heat production (μW/m^3)	Density (kg/m^3)	Heat capacity (J/m^3/K)
Gabbroic crust	2.5	1.6	2950	940
Mantle lithosphere	2.7	0.0	3250	1000
Mantle asthenosphere	2.7	0.0	3250	1000

Data source: Turcotte and Schubert (1982) and Wang *et al.* (1995).

height of 53 km ('constant asthenospheric temperature region' in Fig. 3). This high-temperature region beneath the ridge is kept within the oceanic lithosphere during the calculation, with different parameters set in the ocean and subduction domains. The half-spreading rates (V_s) for the ridge center are set at 10, 30 and 50 mm/year, corresponding to slow, medium and fast rates of plate separation at presently active oceanic spreading centers (Lonsdale 1977). The faster (160 mm/year) and slower (60 mm/year) cases of ridge convergence velocity (V_c) are considered in this study. The modern convergence velocity of the Philippine Sea Plate to the Eurasian Plate is estimated to be ~60 mm/year (Seno *et al.* 1993), whereas that in Cretaceous southwest Japan for convergence of the Kula plate orthogonal to the trench axis of the Eurasian continent is considered to have been ~200 mm/year (Engebretson *et al.* 1985; Maruyama & Seno 1986). The kinematic parameters used in the model are summarized in Table 2. For example, for a convergence rate at the ridge center of 160 mm/year and a half-spreading rate of 50 mm/year, the opposite ridge flanks will be subducted at velocities of 210 and 110 mm/year (Fig. 3b) in the subduction domain. In contrast, in the ocean domain, because thermal structures in the opposite ridge flanks should be symmetric, and a relative ridge convergence velocity with respect to the continental lithosphere could be zero ($V_c = 0$ mm/year), thermal calculations have been solely associated with the half-spreading rates (Fig. 3a). Therefore, for the calculation in the ocean domain, the position of the ridge center is fixed as a boundary condition and thermal structures within a ridge flank are calculated. The thermal structure of the oceanic lithosphere near the spreading ridge is modified by diffusional heat flux from vertical thermal boundary layers. In the ocean domain, even after 30 My of system evolution, the thermal structures remain relatively unchanged and approach a steady state. The steady-state thermal structures are used for the oceanward boundary condition in the subduction domain. The age of oceanic lithosphere entering the subduction domain is directly proportional to the half-spreading rate of the ridge center and the distance from the ridge in the ocean domain. For example, for half-spreading rates of 10, 30 and 50 mm/year, the oceanic lithosphere is 30 My in age at distances of 300, 900 and 1500 km away from the spreading center. In the

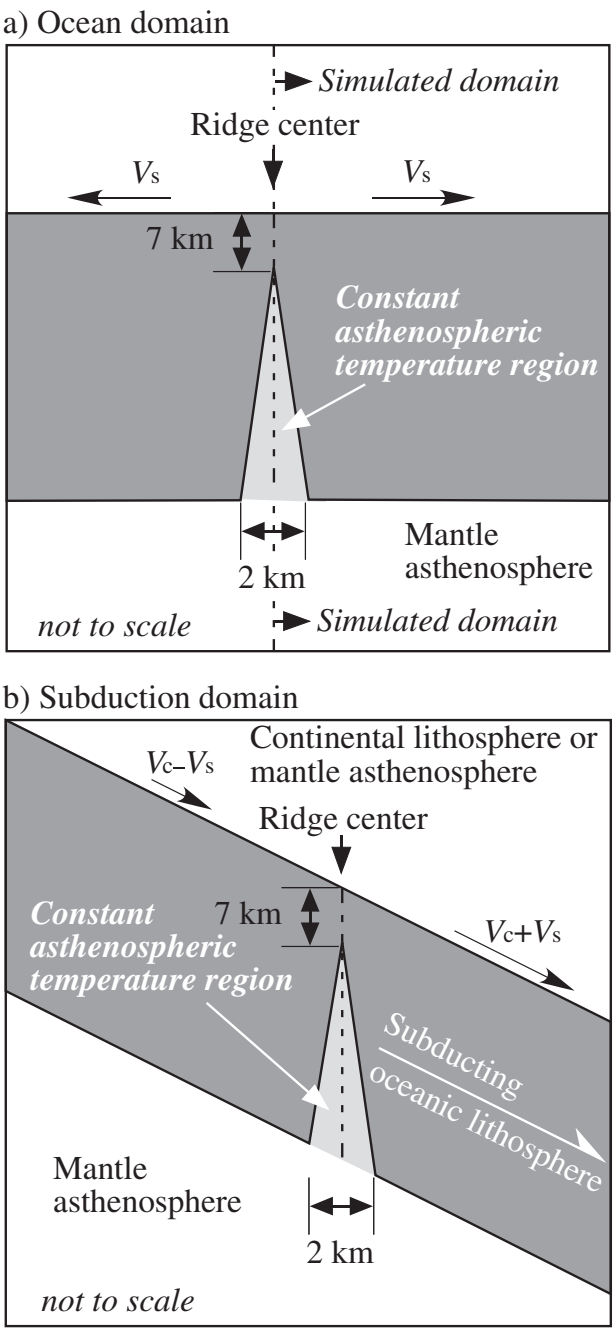

Fig. 3 Schematic diagram showing the geometry of the spreading-center within the oceanic lithosphere. (a) Ocean domain, (b) subduction domain.

Table 2 Kinematic parameters used in this study

	Velocity
Ridge convergence velocity V_c (mm/year)	
Slow	60
Fast	160
Half-spreading velocity V_s (mm/year)	
Slow	10
Medium	30
Fast	50

present study, the oceanic ridge axis approaches the trench parallel to the trench axis. In fact, the consistent Th–U total Pb ages on monazite in both the west and east of the Ryoke Metamorphic Belt are believed to show that intrusion of the granitoids occurred synchronously along the length of the belt (Brown 1998; Suzuki & Adachi 1998), suggesting that subduction of the spreading axis of an oceanic ridge should be almost parallel to the trench axis.

In the subduction domain (Fig. 2), the continental lithosphere is fixed in space and the velocity of the subducting lithosphere is given by **v** ($V_c + V_s$ for the leading slab and $V_c - V_s$ for the following one) in Equation 1. We have assumed that the opposite ridge flanks subducted at different velocities, and then a 'thermal' slab window, defined by the region with a temperature of ~1280°C, could be formed during the ridge subduction in the subduction domain. The initial (pre-subduction) thermal structures of the continental lithosphere are approximated using a 60-km-thick plate-model with a basal temperature of 1280°C. An adiabatic gradient of ~0.3°C/km is incorporated into the vertical temperature structure in the mantle asthenosphere. After onset of subduction of the oceanic lithosphere, the thermal structure in the mantle asthenosphere caused by viscous flow is calculated using the analytical solution of McKenzie (1969). On the landward boundary, the temperature in the mantle asthenosphere above the point at which the flow direction of mantle convection changes in the wedge is calculated according to an adiabatic gradient and fixed, whereas that below the point is calculated based on the stream function of McKenzie (1969). The angle of the subducting oceanic lithosphere is set at 26.6°C, approximating the average dip of modern subduction zones (e.g. Central Chile, Peru and Colombia) associated with a young slab with high convergence rate for depths of 0–100 km (Jarrard 1986). In the present study, frictional heating is not considered, because previous studies have shown that the temperature increase caused by frictional heating is very small in many subduction zones (Furukawa & Uyeda 1989; Wang *et al.* 1995). Hence, the third term on the right-hand side of Equation 1 solely represents radiogenic heat production. The values of physical properties of the gabbroic crust, mantle lithosphere and mantle asthenosphere used for modeling are shown in Table 1.

Before 30 My, at which time the thermal structures in the ocean domain begin to approach a steady state, the initial (pre-spreading) thermal structure of the oceanic lithosphere is used for the oceanward boundary condition. After 30 My, the thermal structures in the subduction domain remain relatively unchanged and also approach a steady state. After 30 My, for each successive time step, the vertical temperature distribution in the downgoing oceanic lithosphere on the oceanward boundary was calculated in the ocean domain. The thermal structures of the subduction domain at 30 My of system evolution are shown in Figure 4 for ridge convergence velocities of 60 and 160 mm/year at a half-spreading velocity of 10 mm/year. At the onset of the subduction, continental lithosphere and mantle asthenosphere cool as a result of heat conduction out of the subducting oceanic lithosphere. In contrast, the upper and lower parts of the subducting lithosphere are heated, primarily because of conduction of heat from the surrounding mantle. As shown in Figure 4, the thermal structures of these two scenarios are slightly different. In the case of high convergence velocity, advective heat transfer resulting from the

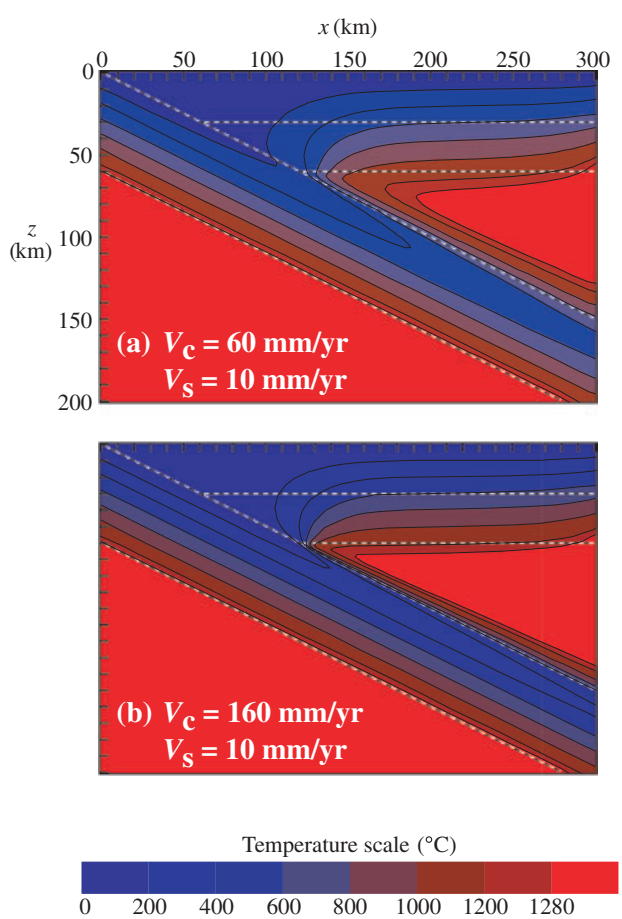

Fig. 4 Temperature profiles in the subduction domain after 30 My of system evolution for ridge convergence velocities (V_c) of (a) 60 and (b) 160 mm/year at half-spreading velocity (V_s) of 10 mm/year.

subducting 'cool' oceanic lithosphere occurs much faster than conductive heat transfer from the surrounding higher-T continental lithosphere and mantle asthenosphere. Hence, the thermal structure of the subduction zone in a higher convergence velocity regime is much cooler than under lower-velocity conditions. In contrast, because viscous flow in the mantle wedge is proportional to the convergence velocity, faster convergence results in higher temperatures in the mantle wedge.

RESULTS

After 30 My of system evolution, the temperature profile of the subducting slab at the oceanward boundary in the subduction domain has changed. The boundary thermal profiles were calculated in the ocean domain for a range of half-spreading rates (10, 30 and 50 mm/year) and the ridge convergence velocities were set at 60 and 160 mm/year in the subduction domain: Six examples are provided in Figure 5. In Figure 5, in order to compare the thermal consequences of spreading-ridge subduction, the thermal structure for each case is represented in terms of the position of the ridge center away from the trench. With the ridge center at –100 km from the trench, there is significant variation in the thermal structure of the slab, as a result of different spreading rates. The thermal structure of the subducting oceanic lithosphere under a fast-spreading regime (50 mm/year) is much hotter than at slower spreading rates (10 mm/year). The age of the oceanic lithosphere at the trench differs according to the distance between the ridge center and the trench (e.g. 10 My at $V_s = 10$ mm/year, 3.3 My at $V_s = 30$ mm/year, and 2 My at $V_s = 50$ mm/year), following from the differences in half-spreading rate.

As shown in Figure 5, a 'thermal' slab window opens when the ridge center moves below the depth of the continental lithosphere–asthenosphere boundary, and with continued subduction the slab window expands. The extent of a slab window is significantly different for the different parameter sets. The temperature structure of the oceanic lithosphere near the ridge differs significantly depending on the convergence and spreading rates, which govern the extent and the duration of a 'thermal' slab window at a particular location. Opening is initiated at slightly different times according to the convergence velocity. These results suggest that in the case of slow convergence and fast-spreading rate, the effect of heat from the surrounding higher-temperature mantle asthenosphere is sufficient for a slab window to form (Fig. 5c,f). In previous works (e.g. Peacock *et al.* 1994; Iwamori 2000; Miyazaki & Okamura 2002), the initial thermal structure of the oceanic lithosphere before subduction was calculated by using an analytical solution based on a steady-state geotherm that is dependent on the distance from the spreading center (i.e. age of the oceanic lithosphere). Because the half-spreading rate is set at 0 mm/year, a slab window cannot open during subduction in these studies, and the thermal structure around the ridge center within the subducting slab could not change substantially. In contrast, the present study demonstrates that the spreading center is still likely to be opening during subduction, which would allow the formation of a region with an asthenospheric temperature (~1280°C) within the slab.

For the period that the spreading ridge is located beneath the continental lithosphere (5 My for $V_c = 60$ mm/year; 1.9 My for $V_c = 160$ mm/year), the thermal structure near the Wadati–Benioff plane will vary dramatically depending on the convergence rate. That is, the temperature of the lithosphere in models with lower convergence rates will be much higher than that in higher-rate models. As the ridge center is subducted, a thermal anomaly is created, causing the isotherms to rise, such that the 1000°C isotherm intersects the Wadati–Benioff plane (Fig. 5). In the slow-spreading cases, an overturned isotherm of 200°C appears in the forearc region. Overall, the influence of the subduction of a spreading ridge on the thermal structure of the crust becomes greater with increasing half-spreading rate and decreasing ridge-convergence rate.

The results of these models contradict those of Iwamori (2000). In his model, higher temperature conditions are predicted around the ridge center within the faster subducting slab, rather than where slower subduction occurs. The difference in thermal structure around the ridge center results mainly from the difference in the mantle-wedge temperature near the interface between the slab and the wedge. The stream function formulated by McKenzie (1969), which was also incorporated in the model of Iwamori (2000), is strongly dependant on the velocity of the subducting slab: Faster subduction will cause faster convection in the mantle wedge and result in higher temperatures in the mantle wedge. In contrast, in the model presented in the present study, thermal structures around

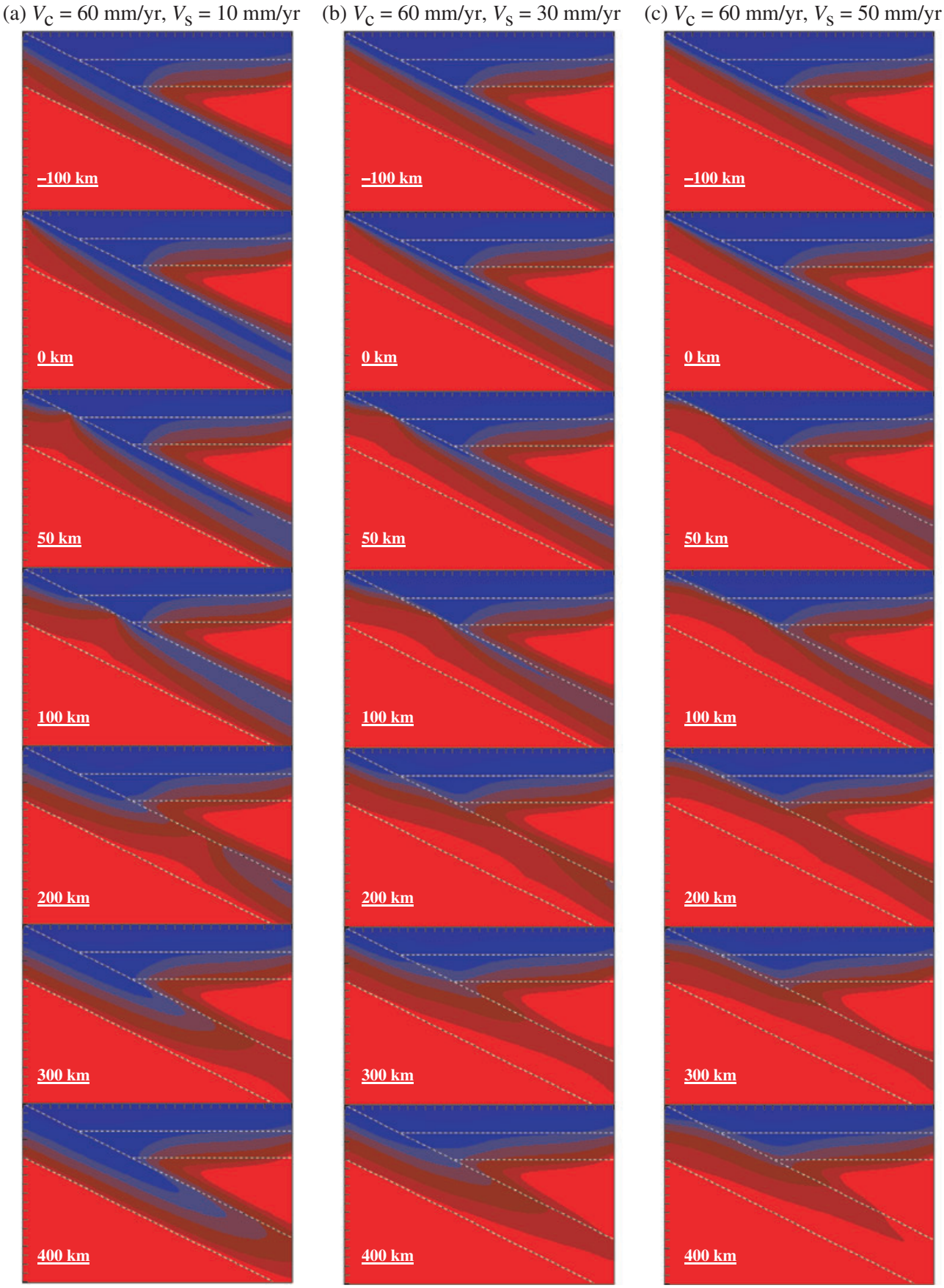

Fig. 5 Thermal structure of the subduction domain in terms of distance between the ridge center and trench for various convergence velocity (V_c) and spreading rate (V_s) scenarios.

(d) V_c = 160 mm/yr, V_s = 10 mm/yr (e) V_c = 160 mm/yr, V_s = 30 mm/yr (f) V_c = 160 mm/yr, V_s = 50 mm/yr

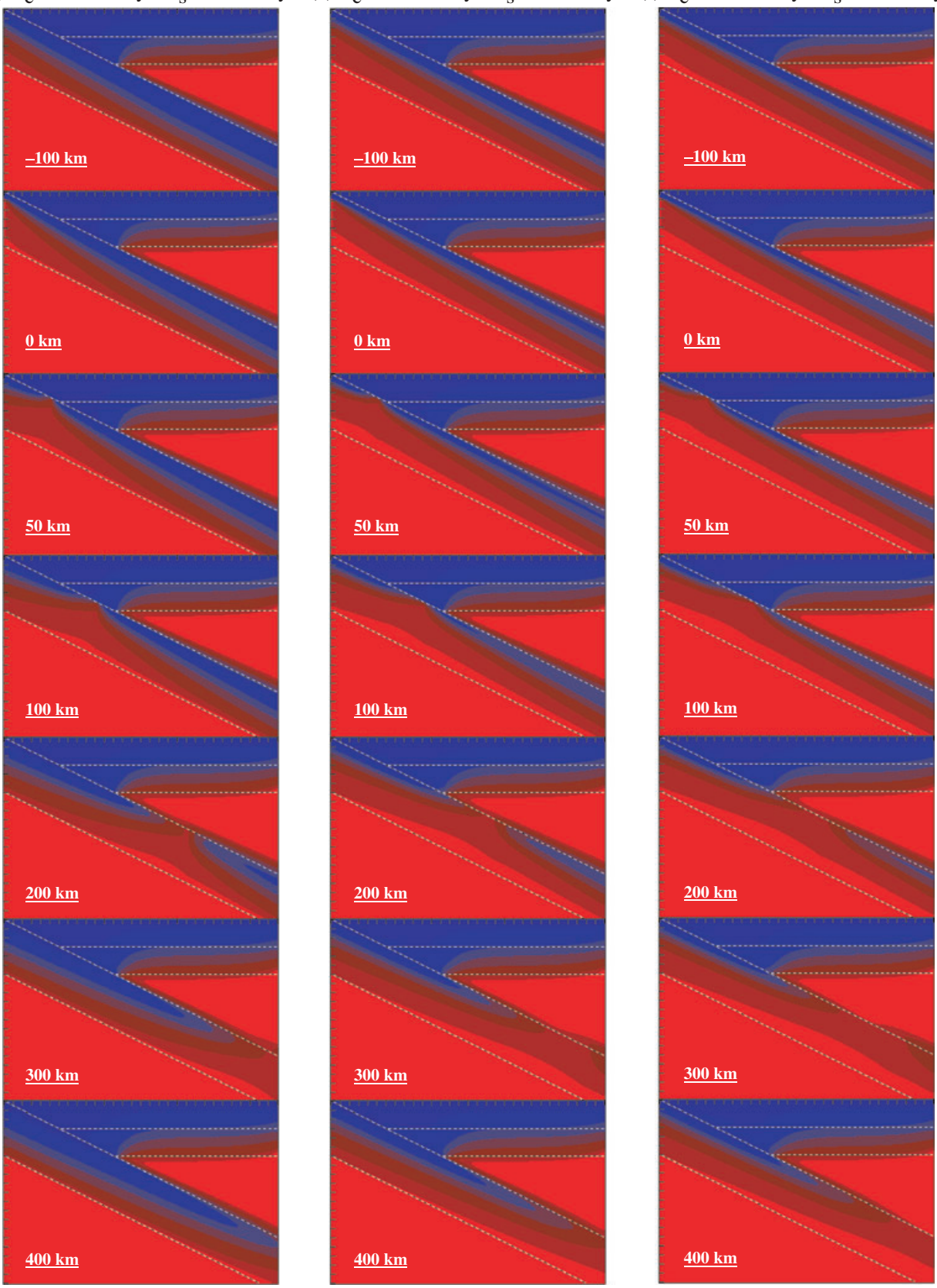

Fig. 5 *Continued*

the ridge center are mainly controlled by the half-spreading velocities that were not incorporated in the model of Iwamori (2000).

DISCUSSION

IS THE SPREADING RIDGE MODEL APPLICABLE TO MID-CRETACEOUS SOUTHWEST JAPAN?

Figure 6 shows the distribution of metamorphic facies in the crust of the forearc region. This distribution is based on the highest temperature–pressure (depth) relationships of the crust after a steady geotherm has been reached before subduction of the ridge, because the P–T conditions of metamorphic rocks are most likely to record when the rocks reached the highest-T conditions during metamorphism. The boundaries of the metamorphic faces follow those of Spear (1993). Iwamori (2000) indicated that the supply and presence of water is essential to drive the metamorphism, yet the ridge subduction can provide water only in a limited period and to a limited space, because it is too hot. Because the model presented here does not incorporate the parameters associated with water production, we cannot evaluate the effect of water on the model results. As shown in Figure 6, in the case of slow convergence rate and fast half-spreading rate, the region of high-grade facies (amphibolite and granulite facies) is enlarged, particularly in the vicinity of the Wadati–Benioff plane. The thermal effect caused by the ridge subduction is limited in the region near the Wadati–Benioff plane, and thermal structures in the region far from the plane do not change significantly. Only in the case of the fastest convergence rate and slowest spreading rate, high-grade metamorphism does not develop in the vicinity of the Wadati–Benioff plane, and metamorphic consequences range from the zeolite facies, through the prehnite–pumpellyite and greenschist facies, to the epidote–amphibolite facies. In this model, because we assumed a crust with a thickness of 30 km, the eclogite facies metamorphic rocks could not form. The metamorphic facies of the Sambagawa Belt range from the prehnite–pumpellyite facies, through the greenschist facies and the blueschist or epidote–amphibolite facies, to the eclogite facies (e.g. Banno & Nakajima 1992; Enami 1998; Aoya *et al.* 2003). This suggests that the formation of the high-P/T Sambagawa metamorphic rocks near the Wadati–Benioff plane requires the subduction of a spreading ridge at a fast convergence rate; results associated with the cases of slow convergence rate are not consistent with this geological interpretation.

The Sambagawa Metamorphic Belt might have been formed near the Wadati–Benioff plane, simultaneously with the Mid-Cretaceous igneous activity; however, the precise timing of the belt formation and igneous activity is a very critical issue. Because of the very high temperature gradient near the Wadati–Benioff plane, only a slight off-set in the timing of belt formation and igneous activity could produce significantly lower temperature conditions, which might explain the Sambagawa metamorphic conditions (DeLong *et al.* 1979; Iwamori 2000, 2002). Unfortunately, we do not know how far the Sambagawa metamorphic rocks were located from the Wadati–Benioff plane when the rocks were metamorphosed under

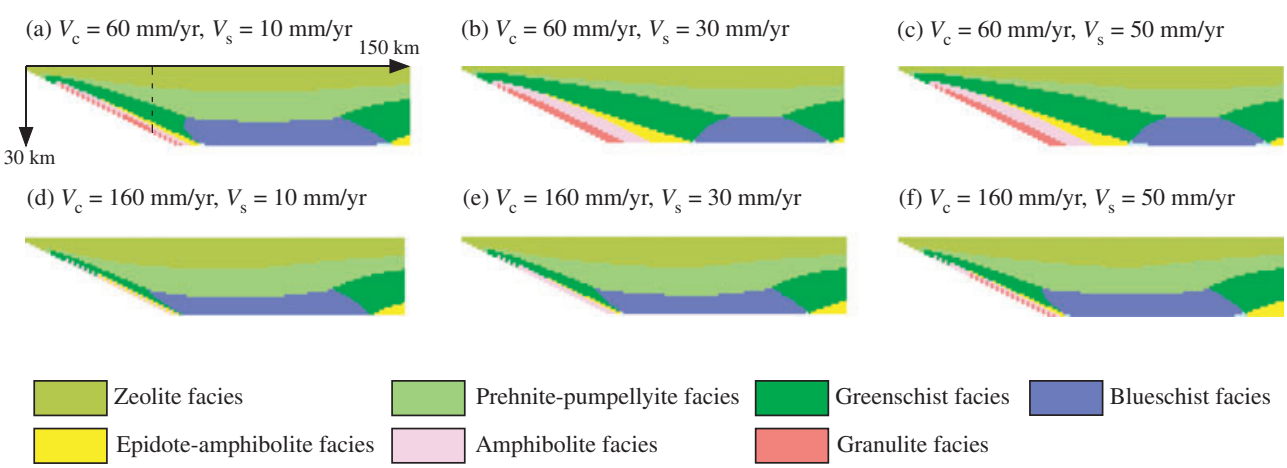

Fig. 6 Distribution of metamorphic facies in the crust of the forearc region under various scenarios of convergence velocity and spreading rate. Dashed line in (a) indicates vertical section representing geotherms in Fig. 7.

the highest-T condition. Iwamori (2002) discusses this issue in terms of calculated pressure–temperature–time (P–T–*t*) paths of the rocks in a forearc wedge when the ridge axis approached. Iwamori (2002) illustrated preliminary results of the P–T–*t* path calculation and stated that the timing of the temperature elevation depends on the position of the rock when the ridge axis approaches. Because P–T–*t* paths for the Sambagawa metamorphic rocks have been petrologically proposed by many researchers (e.g. Enami 1998; Aoya *et al.* 2003), there is a possibility that such numerical analysis would clarify the accurate position of the Sambagawa metamorphic rocks when the ridge or slab window subducted. However, further research in both the petrologic and numerical approaches is needed.

Almost all of the Mid-Cretaceous Ryoke–Sanyo granitoids are I-type granitoids, with major sources being mafic-magmatic rock or metamorphosed equivalents (Kagami *et al.* 1992; Nakajima 1996; Kutsukake 2002). Kutsukake (2002) reported the rare earth element (REE) patterns of these I-type granitoids and suggested that they might have been generated by dehydration melting of amphibolite or hydrous melting of tholeiite at pressures of 1 GPa or higher. Therefore, one of the geological evaluations of the present model is whether the melting of wet tholeiite was sufficient to form a large amount of granitic magma. Because it is difficult to estimate an accurate volume for the granitic magma, based on the surface geology in southwest Japan, we evaluated the model results by using periods of the Mid-Cretaceous granitic magmatism. According to Suzuki and Adachi (1998), granitic activity in the Ryoke–Sanyo zone occurred from 100 Ma to 70 Ma, with the main period of magmatism occurring over ~10 My. Figure 7 shows the geotherms in the crust that lies 50 km from the trench, along with the solidus of dry tholeiite and 5 wt% H_2O tholeiite (Wyllie 1971; Green 1982), and the phase boundaries of aluminosilicates (Spear 1993). Two cases are presented in this figure: The case with the largest thermal effect ($V_c = 60$ mm/year and $V_s = 50$ mm/year) and the case with the smallest thermal effect ($V_c = 160$ mm/year and $V_s = 10$ mm/year). When the ridge center is located at a distance of 50 km from the trench, the temperatures in the lower part of the crust reach hydrous-tholeiitic solidus temperatures under both scenarios. When the ridge center is located 100 km from the trench, the temperatures in the case with the largest thermal effect remain above the hydrous-tholeiitic solidus, whereas temperatures drop below the wet tholeiite solidus in the case with a smaller thermal effect. In the case with the largest thermal effect, temperatures remain above the hydrous-tholeiitic solidus when the ridge center is located 400 km away from the trench (Fig. 7b). These results indicate that the thermal anomaly is restricted in time to less than 1 My for the case of fast convergence rate (160 mm/year) and slow spreading rate (10 mm/year), but might extend for more than 10 My for the case of a slow convergence rate (60 mm/year) and a fast spreading rate (50 mm/year). Even in the case of $V_c = 60$ mm/year and $V_s = 30$ mm/year, temperatures in the crust exceed the hydrous-tholeiitic solidus for less than 4 My. Therefore, in contrast to the interpretation

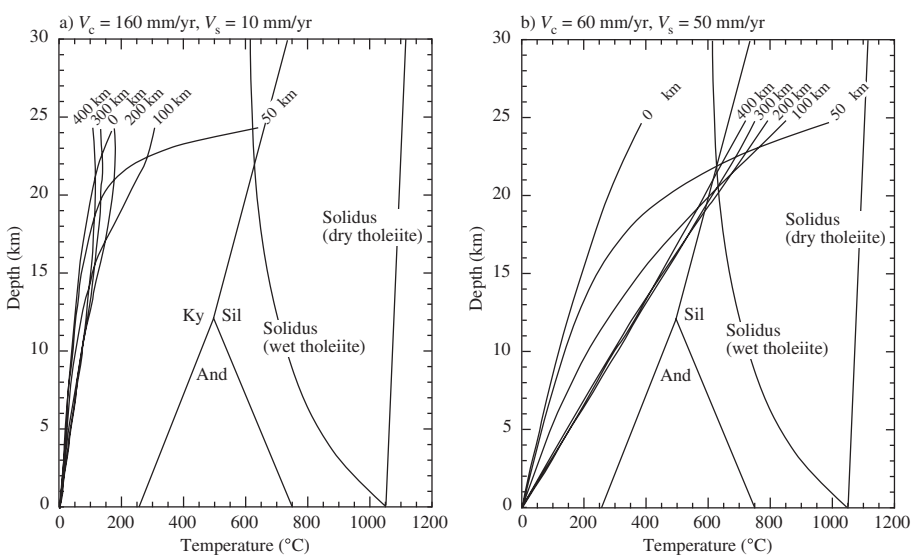

Fig. 7 Geotherms in the crust 50 km from the trench (dashed line in Fig. 6a) under various scenarios of convergence velocity (V_c) and spreading rate (V_s) in terms of the distances between the ridge center and the trench (0 km, 50 km, 100 km, 200 km, 300 km and 400 km). Dry tholeiite and wet (5 wt% H_2O) tholeiite solidi (Wyllie 1971; Green 1982) and aluminosilicate phase boundaries (Spear 1993) are also shown.

of the development of metamorphic facies in the crust, the geological observations implied by Mid-Cretaceous granitic activity requires the subduction of a spreading ridge with a fast spreading rate (50 mm/year) in a slow-convergence regime (60 mm/year). Consequently, in the model of subduction of a spreading ridge studied in the present study, no parameter set can suitably satisfy the two geological interpretations simultaneously, indicating that a slab window did not open during ridge subduction in Mid-Cretaceous southwest Japan.

POSSIBILITY OF SLAB MELTING

In a subduction zone, the geothermal gradient along the slab–mantle interface remains low (=10°C/km). Under such thermal conditions, the subducted slab is metamorphosed and progressively dehydrates in such a way that the dehydration curves intersect with the oceanic lithosphere surface before reaching the solidus at 2 GPa, with the result that the surface of the oceanic lithosphere (i.e. oceanic crust) should be almost completely dehydrated. Melting of such dry tholeiite cannot occur at low temperatures; in fact, temperatures exceeding 1200°C are required at 2 GPa. However, if the subducted oceanic crust reaches 650–700°C before becoming dehydrated, the remnant water will allow dehydration melting to occur, producing adakitic magmas in the subducted oceanic crust (e.g. Drummond & Defant 1990; Peacock *et al.* 1994; Martin 1999; Iwamori 2000).

In the present study, for the case of $V_c = 160$ mm/year and $V_s = 50$ mm/year, the development of high-grade metamorphic rocks along the Wadati–Benioff plane is restricted in space (Fig. 6f). With the passage of the ridge center, a thermal anomaly is created and isotherms rise in such a way that the 1000°C isotherm intersects the Wadati–Benioff plane (Fig. 5f). Because this thermal anomaly can easily exceed the solidus temperature of tholeiite with 5 wt% H_2O, slab melting is possible under these conditions. Therefore, for the case of fast convergence rate and fast spreading rate, if slab melting occurs, adakitic magmas will be formed without the development of high-grade metamorphic rocks near the Wadati–Benioff plane. However, as the thermal anomaly driving the formation of adakitic magmas is restricted in both time (<4 My) and space, the activity of adakitic magma caused by slab melting is too short to generate the extended (>10 My) granitic activity indicated in southwest Japan. In fact, the adakite activity in the Ryoke–Sanyo zone is very restricted (Iizumi *et al.* 2000; Kinoshita 2002), suggesting that the formation of the adakitic magma might not have occurred extensively beneath Mid-Cretaceous southwest Japan.

Iwamori (2000) suggested quantitatively that a normal melt (not adakitic melt) could be produced by slab melting, due to the subduction of a young oceanic slab with a high subduction rate (~250 mm/year). If the slab melting occurs at pressures lower than 1 GPa (~40 km in depth), the melt produced is not adakitic, but normal granite. This shallow slab melting could result in a large amount of granitic magmatism in the forearc region and/or near trench (Iwamori 2000). However, there is no suitably aged granitic activity in the Sambagawa Metamorphic Belt and the Jurassic–Cretaceous accretionary terranes (Chichibu and Cretaceous Shimanto terranes) in the outer zone of southwest Japan, suggesting that the shallow slab melting did not occur in the Mid-Cretaceous.

WHAT IS A PLAUSIBLE TECTONIC SCENARIO?

As described above, two geological constraints on the tectonics for Mid-Cretaceous southwest Japan are: (i) The absence of amphibolite- to granulite-facies metamorphic rocks near the Wadati–Benioff plane; and (ii) significant melting of lower crustal-mafic rocks sufficient to form a long-lasting granitic magmatism. In the model of subduction of an active spreading ridge examined in the present study, no parameter set will satisfy these two geological interpretations simultaneously, which suggests that the ridge subduction that occurred in Mid-Cretaceous southwest Japan did not result in the opening of a slab window. In contrast, the subduction of a very young slab is required to form the high-grade metamorphic belts associated with huge amounts of granitic rock, because the petrologically determined metamorphic field gradient of the high-grade metamorphic belts cannot be explained simply by the thermal structure of the arc-trench system in 'normal' or steady-state subduction (Uyeda & Miyashiro 1974; Peacock *et al.* 1994; Iwamori 2000; Okudaira 2002). According to this suggestion, and the results of this study, two different plausible tectonic scenarios could be suggested to satisfy the geological interpretations for Mid-Cretaceous southwest Japan. The first scenario is that the spreading ridge at the Kula–Pacific (or Farallon–Izanagi) plate boundary, located off the

Eurasian continental margin, did not subduct beneath the continent. In this scenario, very young oceanic lithosphere will be continuously subducted, and the thermal consequences might correspond to those suggested by Aoya *et al.* (2003). As a spreading ridge approaches a trench, the buoyancy of the subducted slab increases. Relative to the asthenosphere, oceanic lithosphere older than ~10 Ma is negatively buoyant and inherently likely to be subducted, whereas lithosphere younger than ~10 Ma is positively buoyant (Cloos 1993). Therefore, in some cases, subduction of a spreading ridge might be stagnant before the ridge intersects the trench. This situation appears to have developed along parts of the East Pacific Rise where it approached North America in the middle Tertiary (Thorkelson 1996). For the ridge subduction to occur, the combined forces of ridge push and slab pull must exceed the buoyant force exerted by the ridge-proximal part of the plate, and the strength of the plate must be sufficient for the slab to subduct without fragmentation (van den Beukel 1990; Larter & Barker 1991). In the second scenario, the spreading ridge associated with the plate boundary ceased spreading before being subducted beneath the Cretaceous continental margin. This scenario is associated with the subduction of very young oceanic lithosphere, similar to the models of Iwamori (2000) and Peacock *et al.* (1994). Consequently, in both the tectonic scenarios, the subduction of plate boundaries at the Mid-Cretaceous southwest Japan was not associated with a slab window, but very young (hot) oceanic lithosphere. However, because we have not evaluated thermal consequences in an arc-trench system caused by the subduction of a very young oceanic lithosphere, we cannot state quantitatively that the tectonic scenarios presented here are suitable for Mid-Cretaceous southwest Japan. Further numerical analyses are required for both tectonic scenarios.

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