Thermochronological investigation of fault zones

Takahiro Tagami

Division of Earth and Planetary Sciences, Graduate School of Science, Kyoto University, Sakyo-ku, Kyoto, 606-8502, Japan.

Email: tagami@kueps.kyoto-u.ac.jp

Abstract

The timing of faulting episodes can be constrained by radiometric dating of fault-zone rocks. Fault-zone material suitable for dating is produced by tectonic processes, such as (1) fragmentation of host rocks, followed by grain-size reduction and recrystallization to form mica and clay minerals, (2) secondary heating/melting of host rocks by frictional fault motions, and (3) mineral vein formation as a result of fluid advection associated with the fault motions. The thermal regime of fault zones consists primarily of the

following three factors: (a) regional geothermal structure across the fault zone and background thermal history of studied province bounded by fault systems, (b) frictional heating of wall rocks by fault motions, and (c) heating of host rocks by hot fluid advection in and around the fault zone. Thermochronological methods widely applied in fault zones are K-Ar ($^{40}Ar/^{39}Ar$), fission-track, and U-Th methods, for which methodological principles as well as analytical procedures are briefly described. The thermal sensitivities of individual thermochronological systems are then reviewed, which critically control the response of each method against the thermal processes. Based on the knowledge above, representative examples as well as key issues are highlighted to date fault gouges, pseudotachylytes, mylonites and carbonate veins, placing new constraints upon geological, geomorphological and seismological frames. Finally, the Nojima Fault is presented as an example for multiple application of thermochronological methods in a complex fault zone.

1. Introduction

Faulting is a major superficial manifestation of geodynamic processes. Hence, the thermochronological investigation of faults and fault zones is critically important in reconstructing the evolutionary history of the Earth and Planets. For example, the knowledge on the timing of fault motions is essential in analyzing a variety of crustal tectonic processes, such as orogenies at plate-convergence margins, continental breakup and rift developments, intra-continental deformation and evolution of sedimentary basins, etc. Besides such academic contributions, reliable dating of the latest motions of an active fault system allows to estimate the recurrence interval of seismicity and thus gives an important clue about when the next big earthquake might take place in the region. The knowledge on the latest fault motions also places critical constraints on the long-term stability of the upper crust, which is one of the key issues for modern environmental research, such as the selection of nuclear waste storage sites. It also provides valuable information about the reconstruction of sedimentary basin history in terms of hydrocarbon exploration.

The classical way of dating fault motions is to indirectly infer the timing from the geological sequence that involves the fault. The time of fault activity is bracketed by dating geological bodies that predate and postdate the faulting. The youngest geological body which is cut by the fault gives the older age limit of the activity, while the oldest body which overlies the fault provides the younger age limit. In addition, the syntectonic sedimentary record preserved in rifts, orogens and strike-slip settings reflects the tectonic history of the area with regard to timing of faulting or other local or regional tectonic events (e.g., Fossen, 2010). An alternative approach is to estimate the age of geomorphological features and processes triggered by recent fault movements. Coupled with radiocarbon analyses, this approach has been applied successfully to estimate the timing of latest fault motions for paleoseismic studies (e.g., Trumbore, 2000).

A direct approach to determine the timing of fault motions is to analyze the fault-zone rocks themselves by thermochronological techniques, which will be highlighted in the present review article. The formation of these rocks is the result of various physical and chemical reactions, such as, frictional heating/melting of host rocks, dynamic recrystallization of minerals, precipitation of mineral veins, as will be described below. The key parameter that characterizes these processes is temperature variation through time and space, which can be reconstructed with confidence by modern thermochronology. The thermal retentivity of radiogenic daughters shows a large variation between individual elements (rigorously, nuclides) utilized as well as between minerals. This fact was widely recognized in 1960's to 70's, and led to the establishment of a new field of geochronology, termed thermochronology, by which the thermal history of rocks can be analyzed quantitatively for geological timescales. See according review volumes (e.g., Reiners and Ehlers, 2005) for historical background, methodological aspects and applications to tackle a variety of geoscientific problems.

The principal aim of the present paper is to review the state-of-the-art about the direct dating of fault-zone rocks using thermochronology. The technical and methodological advancements achieved primarily in the last decade greatly advanced the direct dating, such as, K-Ar ($^{40}Ar/^{39}Ar$) dating on authigenic illite within a fault gouge, zircon FT and $^{40}Ar/^{39}Ar$ laser-probe dating on pseudotachylyte layers, U-Th dating of carbonate veins, etc. In the present contribution, highlighted will be the relevant geoscientific backgrounds, thermal regime of fault zones, thermochronologic overview, and some applications to estimate the timing of fault motions in a variety of tectonic settings.

2. Fault-zone material

2.1. Faults and fault-zone rocks

The geological description of faults in nature and their tectonic/seismic settings, as well as their mechanical backgrounds and formation processes, have been studied in numerous publications and summarized in dozens of textbooks (e.g., Hobbs et al., 1976; Scholz, 2002; Fossen, 2010; and references therein). Figure 1 illustrates the schematic cross-section of a typical fault zone structure that is widely accepted (Scholz, 1988), where geological features of fault-zone rocks are given along with their mechanical backgrounds. Fault-zone rocks exhibit a variety of geological features, such as breccia, gouges, cataclasites, mylonites, pseudotachylytes, etc., which are classified primarily by their textures, i.e., whether or not rocks are foliated/cohesive (e.g., Sibson, 1977).

Those accomplishments provide the base of this brief review on formation processes of fault-zone material on which radiometric dating methods can be applied. The processes are classified into three categories: (1) fragmentation of host rocks, followed by grain-size reduction and recrystallization to form mica and clay minerals, (2) secondary heating/melting of host rocks by frictional fault motions, and (3) mineral vein formation as a result of fluid advection associated with the fault motions. These processes will be described individually below.

2.2. Grain-size reduction and recrystallization

Frictional fault motion is accompanied by damage and erosion of the host-rock surfaces, which process is called wear. Under the conditions in which the rock-forming minerals are brittle, the wear is controlled by brittle fracture and forms loose particles with angular shapes, called gouge. As faulting and wear progress, the gouge shears as a granular material, with cataclastic flow dominated by grain comminution. This process results in the systematic reduction of the average grain size and forms a fractal (i.e., power-low) grain-size distribution. Due to accumulation of total slip, the fault grows in lateral extent as well as in the thickness of the gouge zone.

Fault gouges generally contain a variety of clay minerals. In igneous and metamorphic provinces, in-situ clay mineralization plays a major role in forming authigenic clay minerals within the gouge. In sedimentary provinces, however, significant components of the clay minerals are also derived from wall rocks by mechanical incorporation of protolith clays. Concerning in-situ clay mineralization, fluid flow has been documented in and around the gouge zone (e.g., Solum et al., 2005). Two mineralogical reactions have been widely utilized for K-Ar and $^{40}Ar/^{39}Ar$ thermochronological applications: the illitization of illite/smectite (I/S) and the neocrystallization of authigenic 1M/1Md illite (Haines and van der Pluijm, 2008; Zwingmann et al., 2010ab; Fig. 2, 3). These reactions are kinetically controlled: for example, the smectite to illite reaction is a function of temperature, K-concentration and time (Huang et al., 1993; Grathoff et al., 2001). Hence, it is likely that the reactions are accelerated by hydrothermal flow episodes in terms of the fault activity. The environmental temperature required for the reactions (i.e., ~150°C) can be generally attained at depths of ~5 km under subnormal geothermal gradient (~30°C/km).

Under conditions in which the rock-forming minerals are semibrittle to plastic, the adhesive wear is dominant in ductile faulting (or shearing) and likely contributes to the formation of mylonites (i.e., mylonitic series and blastomylonites; Sibson, 1977). The crystalline plasticity involves solution- and diffusion-aided processes, such as various types of dislocation creep with associated syntectonic recrystallization, pressure solution and superplasticity. In mylonites formed from granitic protoliths, textural changes are observed as the maximum temperature increases (Simpson, 1985): plastic deformation of quartz, brittle fracture of feldspar, and kinking of biotite at ~300°C (i.e., lower greenschist metamorphic grade), recrystallization of quartz, orthoclase and biotite at ~350-400°C (mid-upper greenschist grade), recovery and recrystallization of all minerals at >~450°C (amphibolite grade). White mica is also formed by synkinematic crystallization at the expense of feldspars in mylonitic shear zones at ~400°C (e.g., Rolland et al., 2007). The newly formed micas are widely employed for $^{40}Ar/^{39}Ar$ thermochronology of ductile faulting.

2.3. Secondary heating of host rocks by frictional fault motions

The mechanical work of fault motion is primarily expended by (a) frictional heating, (b) surface energy of gouge formation, and, if seismic slip occurs, (c) elastic radiation (Scholz, 2002). Hence, a general energy balance is written as

$$W_f = Q + U_s + E_s \tag{1}$$

where W_f is the mechanical work of fault motion including both friction and ductile deformations, Q is heat, U_s is surface energy, and E_s is the energy radiated in earthquakes.

Of these three parameters, U_s can be estimated from the wear model (Scholz, 2002), which gives

$$\xi = U_s / W_{fr} = 2\kappa\gamma / hd\mu \tag{2}$$

where W_{fr} is the frictional work of fault motion, κ is the wear coefficient, γ is the specific surface energy, h is the unspecified hardness parameter, d is the diameter of spherical fragment formed by shearing, and μ is the frictional coefficient. On the basis of frictional sliding experiments, Yoshioka (1986) estimated ξ as ~10^{-4} to 10^{-3}, suggesting that the contribution of U_s is negligibly small.

The third parameter, E_s, can approximately be given as

$$E_s = 1/2\ \Delta\sigma\Delta uA \tag{3}$$

where $\Delta\sigma$ is the amount of the shear-stress drop from an initial value σ_1 to a final value σ_2, Δu is the mean slip, and A is the fault area. Meanwhile, the change in internal strain energy, ΔU_e, is written as

$$\Delta U_e = -1/2\ (\sigma_1 + \sigma_2)\Delta uA \tag{4}$$

The mechanical work of fault motion, W_f, is equivalent to ΔU_e, with the sign changed. When combining equations (3) and (4), the seismic efficiency η is given as

$$\eta = E_s / W_f = \Delta\sigma/(\sigma_1 + \sigma_2) \tag{5}$$

It is observed in nature, as well as in laboratory experiments, that shear stresses (i.e., σ_1 and σ_2) are in the 100 MPa range, whereas $\Delta\sigma$ is in the range of 5 – 10 MPa (Scholz, 2002). This implies that E_s is smaller than W_f by an order of magnitude or more.

Therefore, the mechanical work of fault motion, W_f, is primarily expended by generation of frictional heat, Q. The generation of heat Q on the fault plane is thus approximately described by

$$\tau v = q \tag{6}$$

where τ is the mean shear stress acting on a fault sliding at velocity v and q is the heat flow generated by the fault motion (Scholz, 2002). The thermomechanical behavior of faulting is classified into two regimes: (a) primarily under brittle-deforming conditions, a transient heat pulse is generated by rapid coseismic fault slip, with v = ~10 – 100 cm/s, and (b) under ductile conditions, steady-state heat generation occurs from a long-term fault motion averaged over geological time, with v = ~1 – 10 cm/y. The former type of heating generally produces highly-localized thermal signatures in and around the fault zone, with very high maximum temperatures being reached, because of the high velocity and low thermal conductivity of rocks. In certain tectonic settings, fault-zone rocks are occasionally molten to form glassy vein-shaped rocks, called pseudotachylytes (Sibson, 1975; Fig. 4). In contrast, the latter type of heating is expected to form a broader, regional thermal anomaly across the fault zone due to the low velocity coupled with constant long-term heat generation. Some regional metamorphic aureoles in convergent plate boundaries (or transcurrent shear zones) may be attributable to such long-term heating at depths (Scholz, 1980). Time and magnitude of heating can be assessed quantitatively for both types of thermomechanical regimes by modern thermochronological methods, as will be described later in detail.

Finally, a note should be added concerning the estimate of the energy budget of earthquakes using thermochronological methods. As given in equation (4), the mechanical work of fault motion, W_f ($= -\Delta U_e$), is a function of shear stresses, σ_1 and σ_2, which are difficult to measure directly on the fault plane. The observation of seismic waves allows to estimate the total amount of shear stress drop, $\Delta\sigma$, by using equation (3), but does not give any information about σ_1 and σ_2. Thus, an available approach to constrain W_f is to measure the generated heat Q (or q) during an earthquake. As the in-situ, real-time measurement is difficult, following alternative schemes have been proposed: (a) detection of the temperature anomaly across the fault zone by drilling into the fault quickly after the earthquake (Brodsky et al., 2010 and references therein), and (b) geothermometric analyses of fault-zone rocks which record past frictional heating by using vitrinite reflectance and/or fission-track (FT) data (e.g., O'Hara, 2004). This issue will be highlighted in the subsequent sections.

2.4. Mineral vein formation due to fluid advection

A variety of mineral veins, such as calcite, quartz and ore deposits, widely occurs in and around fault zones. Those veins are formed as fracture fillings under an extensional stress regime, as a result of fluid advection associated with in-situ chemical precipitation. They are observed regardless of the modes and depths of fault motion (e.g., Cox, 2010; Fig. 5).

At temperatures above 200° – 300°C, high permeability is probably not maintained longer than the conventional lifetime of hydrothermal systems due to compaction, intergranular-pore cementation, and healing and sealing of fractures to produce veins (Cox, 2005). Therefore, the formation of extensional vein arrays in the seismogenic middle to upper crustal depths likely represents the time of brittle failure and permeability enhancement that induce fluid flow. During the seismic cycle of a fault system, it is thus inferred that a significant slip event produces large ruptures in the fault zone, reduces the fault strength, and enhances fluid flow to form veins (e.g., Sibson, 1992). This episode is followed by an interseismic period when the fault strength shows a progressive recovery as healing and sealing of ruptures progress. Such cyclic features are called fault-valve behaviour (Sibson, 1992).

The temporal variation of fluid advection into a fault zone can thus be constrained by dating mineral veins. U-Th disequilibrium analysis of carbonate veins has been applied successfully to date active fault systems (Flotte et al., 2001; Boles et al., 2004; Verhaert et al., 2004; Watanabe et al., 2008). These attempts will throw a new light about the long-term evolution of seismogenic fault zones. In addition, the thermal anomaly formed by hot fluid advection should be recorded within adjacent wall rocks, and thus such thermal signatures can also be decoded by low-temperature thermochronology of host minerals, such as FT analysis on apatite and zircon separated from fault zones.

3. Thermal regime of fault zones

3.1. Regional geothermal structure and background thermal history

This section provides a brief overview about the thermal regime of fault zones in time and space, as a key issue to govern the applied thermochronological systems, particularly in terms of the thermal retentivity/ diffusivity of accumulated daughter nuclides in the target mineral.

The geothermal structure of the solid Earth can be described approximately by a one-dimensional temperature profile against depth from the Earth's surface. The temperature shows more or less a monotonous increase towards the center of the Earth by reflecting primarily the outward transportation of accumulated internal heat energy that is originally derived from the long-term decay of radioactive nuclides. As a result, geothermal gradients of upper continental crust are basically kept around the average of ~30°C/km all over the globe (e.g., Ehlers, 2005). Tectonic movements, however, tend to perturb the first-order geothermal structure. The thermal regime of fault zones thus crucially depends on three parameters (Fig. 6): (a) regional geothermal structure across the fault zone and background thermal history of studied province bounded by fault systems, (b) frictional heating of wall rocks by fault motions, and (c) heating of host rocks by hot fluid advection in and around the fault zone. These will be separately examined below.

Where a fault motion has some vertical components, a couple of crustal blocks across the fault show differential uplift/subsidence movements with respect to each other. If uplift is accompanied by exhumation, the rocks within the uplifted block effectively cool due to the downward motion of geotherms, as a result of adjustment to the new state of geothermal equilibrium. Conversely, if subsidence is accompanied by sediment deposition on the surface, the rocks within the subsided block are effectively heated due to the upward motion of geotherms. Note that such upward or downward adjustment of geotherms has some delay after the uplift or subsidence by reflecting the time constant for resultant heat conduction (e.g., Mancktelow and Grasemann, 1997).

Fig. 6 (a) illustrates an example, in which the hanging wall uplifts and cools while the footwall subsides and is heated. As the fault motion is repeated in the same direction, the amount of fault slips is accumulated and thus the difference of the thermal signature progressively increases between rocks that were once juxtaposed each other across the fault boundary. The difference eventually may be resolved by the application of thermochronological methods, placing some constraints on the timing and magnitude of vertical components of the fault motions. Low-temperature thermochronology, such as (U-Th)/He, FT and K-Ar ($^{40}Ar/^{39}Ar$) technique, is particularly useful for reconstructing such regional thermal histories, within the spatial range of ~1 – 1000 km from the fault.

In a regional scale, the background thermal history can vary depending on factors, such as surface topography, spatial variation of geothermal structure, tectonic tilting, ductile deformation at depth, etc. For more details on thermochronological applications to regional tectonics, see other comprehensive reviews (e.g., Reiners and Ehlers, 2005).

3.2. Frictional heating of wall rocks by fault motions

Frictional heating is characterized by episodic temperature increase up to an order of 1000°C (i.e., occasionally above the melting temperature of wall rocks), within a typical time period of several seconds and a spatial range of several mm from the fault (in case of brittle deformation).

Based on the heat conduction models of Carslaw and Jaeger (1959), Lachenbruch (1986) quantified the production of frictional heat and its conductive transfer into wall rocks (Fig. 6 (b)). When a fault slips across a fault zone of width $2a$ during a time interval $0 < t < t^*$, the temperature elevation ΔT within a uniform fault zone ($x < a$) and beyond it ($x > a$) is given, respectively, by

$$\Delta T = \frac{\tau}{\rho c}\frac{v}{a}\left\{t\left[1 - 2\iint erfc\frac{a-x}{\sqrt{4\alpha t}} - 2\iint erfc\frac{a+x}{\sqrt{4\alpha t}}\right]\right\} \qquad 0 < x < a,\ 0 < t < t^* \quad (7)$$

$$\Delta T = \frac{\tau}{\rho c}\frac{v}{a}\left\{t\left[1 - 2\iint erfc\frac{a-x}{\sqrt{4\alpha t}} - 2\iint erfc\frac{a+x}{\sqrt{4\alpha t}}\right] - (t-t^*)\left[1 - 2\iint erfc\frac{a-x}{\sqrt{4\alpha(t-t^*)}} - 2\iint erfc\frac{a+x}{\sqrt{4\alpha(t-t^*)}}\right]\right\}$$

$$0 < x < a,\ t^* < t \qquad (8)$$

$$\Delta T = \frac{2\tau}{\rho c}\frac{v}{a}\left\{t\left[\iint erfc\frac{x-a}{\sqrt{4\alpha t}} - \iint erfc\frac{x+a}{\sqrt{4\alpha t}}\right]\right\} \qquad a < x,\ 0 < t < t^* \quad (9)$$

$$\Delta T = \frac{2\tau}{\rho c}\frac{v}{a}\left\{t\left[\iint erfc\frac{x-a}{\sqrt{4\alpha t}} - \iint erfc\frac{x+a}{\sqrt{4\alpha t}}\right] - (t-t^*)\left[\iint erfc\frac{x-a}{\sqrt{4\alpha(t-t^*)}} - \iint erfc\frac{x+a}{\sqrt{4\alpha(t-t^*)}}\right]\right\}$$

$$a < x,\ t^* < t \qquad (10)$$

where ρ is the density, c is the specific heat and α is the thermal diffusivity (Fig. 7). The mean shear stress τ can be estimated from geothermometric analyses (O'Hara, 2004), such as zircon FT data, by using the equations (7) – (10) (see section 7 for details).

If the slip-duration t^* is negligibly small relative to post-seismic observation time $(t - t^*)$, and also if our observation time t is large relative to the time constant λ of the shear zone ($\lambda = a^2/4\alpha$), the equations (7) – (10) are simplified as follows (Lachenbruch, 1986):

$$\Delta T = (\tau u / \rho c)(\pi \alpha t)^{-1/2} \exp(-x^2 / 4 \alpha t) \qquad \text{all } x,\ t >> t^*,\ t >> \lambda \quad (11)$$

where u is the slip distance ($u = vt^*$). By substituting individual appropriate values to ρ, c and α, and by measuring ΔT and u, the mean shear stress τ can be calculated for certain x and t conditions. Equation (11) is employed for the case of temperature anomaly measurement across the fault zone by drilling into the fault soon after the earthquake (Brodsky et al., 2010).

3.3. Hot fluid advection in and around the fault zone

Fluid advection within a fault zone can be inferred from the occurrence of mineral veins formed by in-situ chemical precipitation (see section 2.4). The spatial range of the effective advection is primarily at an order of 1 ~ 100 meters by judging from the natural occurrences of mineral veins (e.g., Boles et al., 2004; Watanabe et al., 2008; Cox, 2010). The key factor that governs fluid flows is the permeability structure of crustal rocks that can vary considerably in time and space. The surface of continents is widely covered by Quaternary sediments that contain units of high permeability, such as sandstone and conglomerate layers, for which fluid advection is dominated by porous flows. In contrast, the permeability is generally lower for the underlying basement bedrocks (cf., Ingebritsen and Manning, 2010), except for fault zones where a group of open cracks (either faults or fractures under an extensional stress regime) can behave as effective pathways for seepage flows.

The generalized permeability structure of a fault zone usually consists of three segments: (a) fault core, which comprises fault gouge and breccia, both characterized by low permeability, (b) damage zone, which consists of fractured rocks and has high permeability and (c) bedrock protolith with low permeability (Evans et al., 1997; Seront et al., 1998). The mean permeability of a fault zone is inferred to decrease with time, as a result of narrowing/closure of pathways due to the continued fluid flow and chemical precipitation to form veins. The permeability likely recovers if the fault zone experiences new seismic activity and resultant reopening of the pathways. This temporal model was first tested by the Nojima Fault Zone Probe Project (see section 7).

If the seepage flows in a fault zone is dominated by upward components, the fault zone is heated by flows from deeper crustal levels, and thus hotter than the environmental temperature (Fig. 6 (c)). In fact, hot springs are often found near active fault systems, and some of them likely have deep origins as indicated by their geochemical signatures (e.g., Fujimoto et al., 2007). Some of the large faults continue from the surface to the middle to lower crust, i.e., >10km depths (Scholz, 2002), and the temperature of upcoming fluid may exceed 300°C by assuming a subnormal geothermal gradient and neglecting heat loss during the seepage flow.

4. Thermochronological methods applied in fault zones

4.1. K-Ar and $^{40}Ar/^{39}Ar$ methods

K-Ar dating method utilizes the electron capture decay of ^{40}K to ^{40}Ar, which has a decay constant of 0.581×10^{-10}/yr and thus is suited to age determination at geologic times. K is one of major elements that constitute the Earth's crust, and this qualifies the K-Ar method, in principle, as one of most useful dating tools for a variety of crustal rocks formed at different tectonic settings. Ar is one of rare (noble) gas elements that do not react with other chemical constituents of rocks and minerals, and this results in greater mobility in the crystal lattice and hence relatively low initial abundance of radiogenic daughters, compared to other dating schemes, such as Rb-Sr and U-Pb, etc. In many rocks formed at or near the earth's surface, the initial ^{40}Ar contents is thus negligibly small and, as a result, the non-radiogenic ^{40}Ar contamination can be corrected for using the atmospheric Ar abundance (^{40}Ar/^{36}Ar = 295.5; Steiger and Jager, 1977). This allows to measure a K-Ar age by analyzing a single specimen, without the need of the isochron approach.

In the ^{40}Ar/^{39}Ar dating method, the transformation of ^{39}K into ^{39}Ar using the (n, p) reaction enables to quantify both parent and daughter nuclides on the same element, namely, ^{39}Ar and ^{40}Ar. Thus, the age is obtained by merely measuring the ratio of ^{40}Ar/^{39}Ar on an aliquot of the sample. This advantages the ^{40}Ar/^{39}Ar method to the conventional K-Ar, especially due to the ability of dating small samples by laser ablation. Another notable feature of the technique is that step-heating and extraction of individual gas aliquots generate ^{40}Ar/^{39}Ar gas release spectra, which reveals a variety of thermal history information.

One potential problem is raised, however, when the $^{40}Ar/^{39}Ar$ technique is applied to samples that are entirely fine grained because ^{39}Ar moves significantly from its original locality during the (n, p) reaction. As a result of this recoil motion, the ^{39}Ar travels on average for ~ 0.1 μm in ordinary rock-forming minerals. This movement may cause the loss of produced ^{39}Ar from the target mineral. The expected travel-distance is negligibly small for samples of common grain size (i.e., >10 μm), but probably gives profound influences on fine-grained samples (i.e., ~<1 μm), such as clay minerals and some volcanic groundmass. To resolve this issue, the encapsulation technique was developed and applied occasionally, in which the sample is put into vacuumed glass capsule prior to irradiation so that the escaped ^{40}Ar can be preserved to obtain the total gas age (Foland et al., 1992). Alternatively, the K-Ar technique is still widely employed for such samples. See other literatures for more details of the K-Ar and $^{40}Ar/^{39}Ar$ methods (e.g., McDougall and Harrison, 1999).

In dating clay minerals formed by in-situ mineralization (Fig. 2, 3), the first key issue is how effectively the essential minerals (e.g., authigenic illites) can be separated from protolith clays. Thus, K-Ar ($^{40}Ar/^{39}Ar$) ages are measured on individual size fractions of each sample. If the sample is collected from igneous provinces where no detrital illites are expected, the age(s) of finest fraction(s) is usually interpreted as that of authigenic illite (Zwingmann and Mancktelow, 2004). If the sample is from sedimentary provinces where components of detrital illites are expected, however, the ages from different fractions are plotted against the detrital illite contents based on polytype quantification (e.g., Ylagan et al., 2002). Then, the younger age intercept is usually adopted as the age of 100% authigenic illite (see Fig. 14; Pevear, 1999).

The second point is how reliably the obtained age of authigenic illite (i.e., the age of finest fraction or that of younger intercept, as mentioned in the last paragraph) can be interpreted in terms of the fault-zone development, particularly within the framework of thermo-tectonic evolution. The closure temperature of illite K-Ar is empirically estimated as 260°C for an ordinary grain size of ~2 μm (Hunziker et al., 1986; see section 5 and Table 1), whereas the temperature of in-situ illite formation is estimated as ~150ºC (Zwingmann and Mancktelow, 2004). Hence, the K-Ar age of authigenic illite records the time elapsed since its in-situ mineralization within the fault zone, unless any secondary heating episode causes the opening of the illite K-Ar system. See other articles for more details (e.g., Haines and van der Pluijm, 2008; Zwingmann and Mancktelow, 2004).

4.2. Fission-track analysis

Fission tracks are narrow damage trails that form as a result of spontaneous nuclear fission decay of ^{238}U, and are thus effectively accumulated through time in uranium-rich mineral grains (e.g., apatite, zircon, titanite) and natural glasses. The time elapsed since fission tracks began to accumulate in a material can be determined by measuring (a) the number of ^{238}U (i.e., parent) nuclides per unit volume, ^{238}N, (b) the number of spontaneous fission tracks (equivalent to daughter nuclides) per unit volume, N_S, and (c) the decay constant for spontaneous nuclear fission, λ_F. FT age equation needs to take into account additional theoretical and experimental factors. To measure ^{238}N, the nuclear fission reaction of ^{235}U that is artificially induced by thermal neutron irradiation is utilized, based on the number of induced fission tracks per unit volume, N_I. In addition, only tracks intersecting the etched surface are observable under an optical microscope. Thus, the N_S and N_I are measured as the surface density of etched spontaneous fission tracks, ρ_S, and that of etched induced fission tracks, ρ_I, respectively. Note that the temperatures of apatite, zircon and titanite formations are substantially higher than individual closure temperatures, and hence their FT data record thermal history after mineralization. See Tagami and O'Sullivan (2005) for more details.

4.3. U-Th method

U-Th dating (also referred to as ^{230}Th dating) is the most important of the U-series disequilibrium dating schemes, and utilizes the radiometric decay of ^{234}U to ^{230}Th comprising a part of the following ^{238}U decay chain. That is, ^{238}U decays by α emission to ^{234}Th (half-life = $4.4683 \pm 0.0048 \times 10^9$ years), which in turn decays by β^- emission to ^{234m}Pa (half-life = 24.1 days), which decays primarily by β^- emission to ^{234}U (half-life = 1.2 minutes), which decays by α emission to ^{230}Th (half-life = $2.4525 \pm 0.0049 \times 10^5$ years), which decays by α emission to ^{226}Ra (half-life = $7.569 \pm 0.023 \times 10^4$ years), which further decays through a series of intermediate daughter nuclides to stable ^{206}Pb. As a consequence of long-term system closure in terms of all the relevant nuclides, the ^{238}U decay chain reaches a state of secular equilibrium where the activities of all the nuclides are equal. When there is an event that produces the chemical fractionation, the secular equilibrium is once broken and then the ^{238}U decay chain starts again to return to the equilibrium. If the chemical fractionation can be quantified, the subsequent gradual return to the equilibrium is a function of time, allowing to measure the age of the fractionation event.

The U-Th method has been successfully applied to date a variety of upper Pleistocene carbonates, such as corals, mollusks shells, speleothems, etc. In general, U is soluble as uranyl ion into natural water under oxidizing conditions, whereas Th has extremely low solubility in water. As a result, natural carbonates precipitated from such water are relatively rich in U and extremely poor in Th. For example, the seawater at its surface possesses $^{230}Th/^{238}U$ ratios ~10^5 times lower than that in the secular equilibrium. In the decay scheme of ^{234}U to ^{230}Th, therefore, the initial concentration of daughter nuclide ^{230}Th can often be regarded as negligible when analyzing carbonates. Subsequent to the precipitation and system closure, ^{230}Th is gradually accumulated in the carbonate as a consequence of the decay of ^{234}U contained. Thus it enables to determine the age of carbonates, with a possible age range from several to ~5 x 10^5 years.

Two key factors to be taken into account for reliable dating are: (1)

whether or not the initial abundance of ^{230}Th is negligibly small compared to ^{234}U (or ^{238}U), and (2) how well the assumption of closed system holds since precipitation of the carbonate. In addition, in the particular case of dating faults, carbonate veins frequently occur as narrow bands (<1 cm) and thus it is not necessarily possible to collect a sufficient amount of the target carbonate without contamination. For more details, see other review papers (e.g., Edwards, et al., 2003; Richards and Dorale, 2003).

5. Thermal stability of thermochronological systems

5.1. General kinetic background

A series of fault motions may involve multiple thermo-mechanical processes (cf. sections 2 and 3). Individual thermochronological systems (i.e., the combinations of method and mineral, such as illite K-Ar, zircon fission-track, etc.) exhibit substantially different responses against each of those processes relevant to faulting. Hence, it depends both on the fault-zone material used and the thermochronological systems applied whether or not a specific faulting episode can be decoded.

In general, a reset of a thermochronological clock is achieved by the loss of accumulated daughter nuclides (or fission tracks) from the mineral employed for analysis. Daughters are secondary products that accumulate after the solidification of the mineral, and thus tend to have higher mobility compared to their parent nuclides that were originally incorporated into the mineral structure. The most important environmental factor with respect to faulting is the temperature increase and resultant melting and/or recrystallization of fault-zone rocks. The thermal retentivity of radiogenic daughters shows a large variation between individual elements (rigorously, nuclides) utilized as well as between minerals.

It is widely recognized that thermal diffusion is the fundamental process of the loss of daughter nuclides from the host mineral. In general, the concentration distribution $C(x,y,z)$ within a solid is described by the following diffusion equation, as a function of the rectangular space coordinates and time t:

$$\frac{\partial C}{\partial t} = D\left(\frac{\partial^2 C}{\partial x^2} + \frac{\partial^2 C}{\partial y^2} + \frac{\partial^2 C}{\partial z^2}\right) \tag{12}$$

where D is the diffusion coefficient. In the case of spherical diffusion geometry of radius r, with initially uniform concentration C_0 and boundary condition of infinite reservoir having zero concentration, the equation (12) can be solved for the radial concentration distribution using the conversion to spherical coordinates (Crank, 1975; see also McDougall and Harrison, 1999):

$$C = \frac{C_0 2r}{\pi R} \sum_{n=1}^{\infty} \frac{(-1)^n}{n} \sin\frac{n\pi R}{r} \times \exp\left(\frac{n^2\pi^2 Dt}{r^2}\right) \tag{13}$$

where R is the distance form the origin. In terms of the atomistic mechanisms of diffusion, there are four possible schemes of migration via a series of random jumps from the current lattice position to one of adjacent sites; namely, exchange, vacancy transfer, interstitial movement and interstitialcy displacement (McDougall and Harrison, 1999). The net flux of the random migration is controlled by the concentration gradient in and around a solid, such as given by equation (13). As the environmental temperature increases, the probability to yield a random jump increases exponentially as a result of increased atomic oscillation. Hence, the rate of diffusion follows the Arrhenius equation:

$$D = D_0 \exp\left(-\frac{E}{RT}\right) \qquad (14)$$

where D_0 is the frequency factor (D at infinitely high temperature), R is the gas constant, T is absolute temperature, and E is the activation energy.

On the basis of these fundamental equations, a variety of equations are derived that describe followings. (a) The episodic fractional loss of a nuclide from a solid, which is useful in analyzing noble gasses (such as Ar and He) because the equation (13) is of limited use in its form due to the lack of analytical resolution to directly image isotope distribution in site (e.g., Harrison and Zeitler, 2005). (b) ^{40}Ar/^{39}Ar age spectra, given as a theoretical release pattern of relative flux of ^{40}Ar with respect to ^{39}Ar (Turner, 1968). (c) An Arrhenius plot that, combined with the fractional loss formulas, offers a basis of diffusion experiments to determine E and D_0. See other comprehensive summaries for details (e.g., McDougall and Harrison, 1999; Harrison and Zeitler, 2005).

Although the fission tracks are equivalent to daughter nuclides in terms of the principle of radiometric age calculation, the kinetic formulation of FT annealing is different from the one based on volume diffusion (i.e., Equations (12) – (14)). If a host rock is subjected to temperature increase, fission tracks that have been accumulated are shortened progressively and eventually erased by thermal recovery (i.e., annealing). Because thermal diffusion basically governs the annealing process, the reduction of FT lengths is a function of heating time and temperature. Moreover, fission tracks are partially annealed over different temperature intervals within different minerals. The FT age reduction results from the decrease in lengths of etched tracks, which is basically controlled by the diffusion of disordered atoms and was described by first–order kinetics in early studies (e.g., Mark et al., 1973). Later studies demonstrated that the FT annealing is better described by higher-order kinetics (e.g., Green et al., 1988). In addition, it was recognized that (a) FT annealing is more precisely quantified by using the reduction of etched track length compared to that of etched track density, and (b) the shape of track length distribution is indicative of a rock's thermal history. Accordingly, horizontal confined track lengths are routinely analyzed to determine the annealing kinetics of fission tracks in apatite and zircon. An example equation is given below:

$$m = 11.35\left[1 - \exp\left\{-6.502 + 0.1431\frac{(\ln t + 23.515)}{(1000/T - 0.4459)}\right\}\right] \qquad (15)$$

where l_0 is the mean FT length in zircon, after annealing at T Kelvin for t hours (Tagami et al., 1998). The FT kinetic models tend to be fan-shaped if the most appropriate models are chosen empirically so as to best fit to the laboratory heating data. See other comprehensive reviews for more details (e.g., Donelick et al., 2005; Tagami, 2005; Tagami and O'Sullivan, 2005).

5.2. The closure temperature concept

Dodson (1973) conducted a landmark study that introduced the closure temperature (T_C) concept for a geological cooling system, and eventually led to the rise of thermochronology . He gave an analytical solution of T_C by assuming that T changes linearly with $1/t$ for a relevant part (at least, over the closure interval), as generally expected for cooling crustal rocks. Based primarily on the equation (14) and the fractional-loss derived from equation (13), T_C is given by

$$\frac{E}{RT_C} = \ln\left(\frac{ART_C{}^2 D_0/r^2}{E\, dT/dt}\right) \qquad (16)$$

where A is the geometric constant (55 for sphere, 27 for cylinder and 8.7 for plane sheet; Dodson, 1973). T_C for the FT system was approximately given, based on the first-order kinetics, by

$$B\exp\left(\frac{E_{50}}{RT_C}\right) = -\frac{RT_C{}^2}{E_{50}\, dT/dt} \qquad (17)$$

where B is a constant and E_{50} is the activation energy for 50% track loss under isothermal conditions (Dodson, 1979). Note that T_C is incorporated on both sides of each equation and hence needs to be solved by iteration, although the calculated T_C values converge rapidly, usually in two iterations (McDougall and Harrison, 1999).

Figure 8 illustrates the schematic outline of the closure temperature and time model. The cooling curve of a monotonically cooling rock may be approximated by a hyperbolic function, in which T changes linearly with $1/t$

(Fig. 8, top). If the rock is analyzed by a specific thermochronological system, an apparent age called closure time (t_C) is calculated by implementing the observed amounts of parent and daughter nuclides (P and D, respectively) into the conventional age equation:

$$t_C = \frac{1}{\lambda} \ln\left(\frac{D}{P} + 1\right) \qquad (18)$$

where λ is the decay constant. On the D/P vs t diagram (Fig. 8, bottom), the t_C is graphically given as an intercept on the horizontal t-axis by extrapolating the growth curve based on the D/P value observed at present. Then, on the cooling curve (Fig. 8, top), a T_C value can uniquely be defined that corresponds to the given t_C value. An important remark here is that the 'switching' between the closure and opening of the system is actually transitional because the daughter's loss from the mineral is primarily governed by the thermal diffusion, in which diffusion coefficient changes gradually for a range of temperature (see equation 14). The transitional feature is graphically expressed on the Fig. 8 (bottom).

The classical way to reconstruct thermal histories via T_C concept is to measure a set of (t_C , T_C) values using multiple thermochronological systems to provide markers on the cooling curve: t_C is given by the radiometric dating of individual samples, whereas T_C is commonly determined by equation (16) or (17) based on previous diffusion or annealing experiments.

Table 1 shows a list of closure temperatures for geological cooling systems (modified from Reiners et al., 2005). Key issues are briefly summarized below:

- Cooling rate: T_C depends on the cooling rate of the system around each closure interval, as expressed in equation (16) or (17). The T_C values given in the Table 1 were calculated for ordinary geological cooling rates (i.e. 1-100°C/m.y.) typical for most tectonic, magmatic and metamorphic processes within the middle to upper crust. Those values are, however, not necessarily useful for geological processes that involve much faster heating and cooling of rocks, such as volcanism, faulting, impact cratering, etc. In such cases, the effective T_C values may increase substantially, and thus need to be recalculated for individual thermal processes.

- Grain size (i.e. radius r): T_C can vary with r (see equation (16)) and thus care should be taken when thermochronological analysis is conducted on samples with unusual grain sizes, e.g. fine-grained fault gouges. As r decreases, T_C also shows substantial reduction and hence the quoted values in Table 1 should be recalculated. Note this does not refer to FT analysis (see equation (17)).
- Excess daughter products: the T_C model (equation (16)) is based upon the assumption that the target mineral is surrounded by an infinite reservoir having zero concentration of daughters (see last subsection). Hence, care should be taken if the model is applied to thermal history analyses of igneous or metamorphic rocks, particularly with K-Ar and $^{40}Ar/^{39}Ar$ techniques (i.e., the issue known as excess ^{40}Ar; see e.g., McDougall and Harrison, 1999;

Kelly, 2002).

- Radiation damage: since the He diffusivity of apatite is negatively correlated with the accumulated radiation damage (Shuster et al., 2006; Shuster and Farley, 2009), the T_C of the apatite (U-Th)/He system may increase with radiation damage, ranging from ~50 to 115 °C for a cooling rate of 10 °C/m.y. (Shuster et al., 2006; Table 1). This relationship is consistent with the observed discrepancy between the FT and (U-Th)/He ages of apatites from old cratonic basements, yielding anomalously old (U-Th)/He ages (e.g., Green and Duddy, 2006). New kinetic models have been explored by incorporating the effects of accumulation and annealing of radiation damages (Flowers et al., 2009; Gautheron et al., 2009).
- Chemical composition: the T_C of apatite FT system depends on the chemical composition. Fission tracks in Cl-rich apatites anneal slower than those in OH- and F-rich apatites (e.g., Green et al, 1986; Barbarand et al., 2003). As a result, the T_C assigned to ~90°

– 120 °C for 'ordinary' apatites (Table 1) needs to be significantly raised for Cl-rich apatites (see e.g., Ketcham et al., 1999).

- Absence of recystallization: a (re)crystallization of minerals below T_C violates the prerequisite of the closure model, and hence thermochronological ages cannot be considered as an immediate time record. This may be the case for low-grade metamorphism or hot-fluid alteration, including in-situ clay mineralization within fault gouge zones argued in section 2.

5.3. The Partial Retention (Annealing) Zone

The closure temperature proved a useful concept to delineate fast and simple cooling events, but the model does not account for complex, non-monotonous thermal histories (cf. Fig. 8, top). Thermal processes of fault zones often represent such cases, where reheating episodes of wall rocks are likely involved. This is particularly critical if the maximum temperature reached during the reheating is close to the T_C value of the adopted thermochronological system and, accordingly, the radiometric clock is not completely reset (see the last subsection). In this case, the observed age (referring to the closure time t_C) will be younger than the initial cooling of wall rocks and older than the reheating, and thus lacks any specific geological relevance.

Instead, the transitional character of age reset is well represented by the partial retention zone (PRZ) model (Fig. 9). The PRZ is defined as a temperature range in which the daughter nuclide is partially retentive within the mineral, and intervenes between a higher-temperature zone of ~100% loss and a lower-temperature zone of ~100% retention. The transitional age reduction that characterizes the PRZ in nature has been observed using less retentive systems, such as FT and (U-Th)/He, in ultradeep boreholes (e.g., Gleadow and Duddy, 1981; Warnock and Zeitler, 1998), in exhumed crustal sections (e.g., Wagner and Reimer, 1972; Gleadow and Fitzgerald, 1987; Stockli and Farley, 2004) and at contact metamorphic aureoles (e.g., Calk and Naeser, 1973; Tagami and Shimada, 1996). The transition was first recognized for apatite fission tracks (Partial Annealing Zone/ PAZ; Wagner and Reimer, 1972; Gleadow and Fitzgerald, 1987,), followed by the expansion of the concept to incorporate noble-gas diffusive systems (PRZ: e.g., Wolf et al., 1998). The nature of the PAZ is slightly different from that of the PRZ because the governing equations depend on higher-order kinetics vs. volume diffusion, respectively.

For a homogenous geological body with the subnormal geothermal regime, the modern PRZ can be recognized in vertical crustal sections (e.g., boreholes) as a characteristic profile of systematic, downward decrease of observed ages (Fig 9). In tectonically stable regions, where exhumation rates have been relatively low for an extended period of time, the shape of PRZ is primarily governed by the vertical temperature profile (Fig. 9 (b)). In tectonically active regions, however, the shape of PRZ is also sensitive to the rate of exhumation (Fig. 9 (a)). If the extended period of tectonic stability is followed by a recent, rapidexhumation (or unroofing) event, the characteristic profile of age decrease can be preserved as a fossil PRZ (Fig. 9 (c)). Conversely, a vertical age profile allows to deduce a variety of thermo-tectonic information, such as exhumation rate, timing and amount of unroofing, paleogeothermal gradient, etc (see e.g., Brown et al., 1994, for more comprehensive studies).

Thermochronological transects from a fault zone into the host rocks often show a similar trend of age decrease towards the fault, and thus provide constraints upon timing and amount of heat generation and /or transfer as well as the resultant anomalous geothermal structure. When the PRZ model is applied to fault zone analysis, however, the effective heating time requires particular attention. Because the retentivity of daughter nuclides within the mineral is governed by thermal diffusion, the temperature range corresponding to the PRZ can vary with the duration of heating (see equation (13), in which the concentration of daughter, C, depends on heating duration, t). Figure 10 illustrates the dependence of the PRZ upon the heating time in an Arrhenius plot, using the zircon FT annealing kinetics as an example

(Tagami et al., 1998).

5.4. Thermal history modeling

Quantitative analysis of thermal history has been archived in the past two decades by thermal modeling techniques that incorporate kinetic formulation of individual thermochronological systems. Forward and inverse models were developed for reconstructing (relevant ranges of) temperature-time pathways from thermochronological data. Several sophisticated modeling packages became available and used widely, such as Monte Trax (Gallagher, 1995; Fig. 11), HeFTy (Ketcham 2005), etc. The approach was particularly successful for FT data by using the track length distribution as a diagnostic, sensitive tool for low-temperature thermal history (e.g., Laslett et al., 1987; Green et al., 1989). Subsequently, (U-Th)/He data is also widely modeled to cover even lower temperature ranges (e.g., Wolf et al., 1998).

The basic scheme of a modeling procedure is summarized as follows:

(1) Thermal diffusion of daughter nuclides, or thermal annealing of fission tracks, in the mineral is formulated as a kinetic equation with temperature and time as variables (e.g., equation (15)). The kinetic equation is conventionally derived from laboratory heating experiments, and then its geological applicability is verified using well-controlled geological samples, such as the deep borehole rocks of a sedimentary basin that has well-known burial history (e.g., the Otway Basin; Green et al., 1989; Ketcham et al., 1999).

(2) The kinetic equation is incorporated into a thermal-modeling algorithmic package to construct a forward model that describes the evolution of kinetic parameters (e.g., FT length-distribution and age in apatite) as a function of time and temperature (e.g., Green et al., 1989; Ketcham 2005). A variety of time-temperature pathways generated randomly can accordingly produce characteristic evolutions of kinetic parameters.

(3) Given a measured kinetic parameter, the (statistical range of) time-temperature pathway is estimated by inverse modeling approaches, in which optimal fits of the kinetic parameter are searched computationally between the forward-model calculation and measured data. A number of computational solutions and modeling packages have been developed (e.g., Gallagher, 1995; Ketcham et al., 2000).

For more details, see other review articles (e.g., Ketcham, 2005; Dunai, 2005).

5.5. Hydrothermal and flash heating experiments

The thermochronological analysis of fault zones requires special consideration of hydrothermal and flash heating. Fault-zone rocks may have been subjected to hot fluid advection during fault development (see sections 2 and 3) and suffered from effective thermal overprints at hydrothermally-pressurized conditions. Various laboratory heating experiments were conducted (Brix et al., 2002; Yamada et al., 2003) to evaluate the effect of such heating on track annealing characteristics in zircon. Yamada et al. (2003) found that the observed FT length reduction is indistinguishable between the atmospheric and hydrothermal conditions using the same zircon sample and analytical procedure. This finding validates the application of conventional annealing kinetics to hydrothermal heating in nature, such as fault zones and sedimentary basins.

Frictional heating along a fault is a short-term geological phenomenon with effective heating duration in an order of seconds (see section 2 and 3), which is significantly shorter than conventional laboratory heating of ~10^{-1} to 10^{4} hours. Thus, high-temperature and short-term annealing experiments were specially designed and conducted using a graphite furnace coupled with infrared radiation thermometers (Murakami et al., 2006b; Fig. 12). The observed track length reduction by 3.6 – 10 sec heating at 599 – 912°C is, overall, slightly more advanced than that predicted by the FT annealing kinetics based upon the heating for ~10^{-1} to 10^{4} hours at ~350 –

750 °C (Yamada et al., 1995; Tagami et al., 1998). Spontaneous tracks in zircon are totally annealed at 850 ± 50°C for ~4 seconds, suggesting that the zircon FT system can be completely reset during the pseudotachylyte formation in nature (e.g. Otsuki et al., 2003, for estimated heating conditions of pseudotachylyte of the Nojima fault).

6. Geological applications and key studies

6.1. Fault gouges

The illite K-Ar and $^{40}Ar/^{39}Ar$ dating techniques have been applied to fault gouge samples collected from a variety of tectonic settings. Lyons and Snellenburg (1971) presented a pioneering work using gouges from normal faults in basement terranes of western New Hampshire, USA. They employed <2 μm fractions that contain only clay mineral and quartz, without 1M or 2M muscovites, and obtained internally-consistent K-Ar ages of 160 ±

4 Ma, demonstrating the potential for determining times of brittle faulting.

Subsequently, the technique was applied to fault gouges from other settings and protolith types, using even finer size fractions (e.g., Kralik et al., 1987, and references therein). Kralik et al. (1987) reported K-Ar and Rb-Sr ages of four gouges from two fault zones in the Eastern Alps, and of the undeformed host rocks. The K-Ar ages are concordant with individual Rb-Sr isochorn ages in three gouge samples, and are consistently younger than the ages of individual host rocks. In particular, a gouge sample consisting of pure illites shows concordant K-Ar (22 ± 2 and 30 ± 2 Ma) and Rb-Sr ages (23 and 32 Ma) for <0.5 and 0.5 – 0.9 μm fractions, respectively, which ages are considerably younger than those of the undeformed host rocks (i.e., K-Ar ages of 133 ± 10 and 165 ± 9 Ma and Rb-Sr ages of 97 and 97 Ma for <0.5 and 0.5 – 0.9 μm fractions, respectively).

Occasionally illite K-Ar ages overestimate the time of faulting in sedimentary host rocks, probably as a consequence of gross contamination of detrital clay materials even in very fine size fractions (e.g., Grathoff et al., 2001). To overcome this problem, two approaches have been attempted so far. Tanaka et al. (1995) studied the K-Ar systematics of cataclasites and fault gouges from the Akaishi Tectonic Line, central Japan. They showed that an increasing degree of deformation and alteration corresponds to the crystallinity index (Kübler index; Kübler, 1968) of the micaceous minerals contained. In this study, the apparent K-Ar ages and Kubler index yielded a negative correlation that can be fitted by a hyperbolic curve converging to ~15 Ma (Fig. 13). This age was interpreted as the time of hydrothermal alteration associated with fault motion.

van der Pluijm and his collaborators utilized quantitative X-ray analysis of clay grain size populations to quantify the ratio of authigenic and detrital micas for individual clay size fractions (e.g., van der Pluijm et al., 2001, 2006; Solum et al., 2005; Haines and van der Pluijm, 2008). A correlation of apparent $^{40}Ar/^{39}Ar$ ages (i.e., total-gas ages of the encapsulation technique) and detrital illite contents allows to estimate the ages of authigenic and detrital illite populations by age trend extrapolating at the 0% and 100 %

intercepts (Haines and van der Pluijm, 2008; Fig. 14). Using this approach, Haines and van der Pluijm (2008) estimated the age of fault motion of the Sierra Mazatan detachment, Sonora, Mexico, as 14.9 Ma, whereas the detrital illite was formed at 18.5 Ma.

The applicability of illite K-Ar dating was further demonstrated on Alpine fault gouges by Zwingmann and collaborators (Zwingmann and Manckeltow, 2004; Zwingmann et al., 2010a). They studied brittle fault zones in protoliths without illites, so that all dated illites were formed by necrystallization within the gouge. The studied localities include two tunnel sites in which any influence of weathering can be excluded (Zwingmann et al., 2010a). The measured K-Ar ages are consistent internally, and checked independently by field evidence and other thermochronological systems, primarily apatite and zircon FT data. Moreover, the tunnel studies also revealed that potential contamination of fine-grained cataclastic protoliths does not significantly influence the authigenic illite ages.

6.2. Pseudotachylytes

Several attempts have been made to date the glassy matrix of pseudotachylytes: e.g., ^{40}Ar-^{39}Ar thermochronology of the Vredefort dome in South Africa (Reimold et al., 1990), the North Cascade Mountains in the western United States (Magloughlin et al., 2001), the Alpine Fault in New Zealand (Warr et al., 2003), the More-Trondelag Fault, Central Norway (Sherlock et al., 2004); glass fission-track thermochronology of the Alpine Fault Zone in New Zealand (Seward and Sibson, 1985); and Rb-Sr geochronology of the Quetico and Rainy Lake-Seine River fault in the western Superior province of the Canadian Shield (Peterman and Day, 1989).

Warr et al. (2003) conducted laser ^{40}Ar-^{39}Ar step heating analysis on six segments of a pseudotachylyte and its wall rock, coupled with the observation of mineral assemblage and fabric, particularly the biotite-glass microstructure and chemistry. They found that the ^{40}Ar-^{39}Ar total gas ages become progressively older from ~1.1 Ma at the center of pseudotachylyte to ~7.2 – 13.3 Ma at the wall rock. Because the regional background age is ~1 – 5 Ma in the studied area (published mica K-Ar ages of the Alpine Schist), the ~7.2 – 13.3 Ma ages likely suffered from excess ^{40}Ar, as also suggested by the ^{40}Ar/ ^{40}Ar step-heating spectra. They interpreted that the time of frictional melting is given by the youngest age of ~1.1 Ma from the center, where the wall-rock biotite and other K-rich minerals are absent.

Sherlock et al. (2004) carried out the laser-probe ^{40}Ar-^{39}Ar spot analysis on polished thick sections of pseudotachylyte that have less host-rock clasts within the matrix, in comparison to the mylonite host rock. Eleven spot ages of pseudotachylyte range from 268 ± 5 to 311 ± 26Ma, significantly younger than the muscovite ages of the mylonite that range from 392 ± 6 to 431 ± 15 Ma. The significant variation of pseudotachylyte ages, as also observed in other cases (e.g., Kohut and Sherlock, 2003), may be due to heterogeneously distributed inherited argon. The presence of inherited argon may result from the failure of one of the key assumptions of dating pseudotachylyte, i.e., the host-rock argon is lost to an infinite reservoir during near-instantaneous frictional melting (Sherlock et al., 2004).

As highlighted above, the ages from glassy matrix of pseudotachylyte may be difficult to interpret depending on two potential issues: (1) whether or not frictional heating associated with pseudotachylyte formation reset the age, i.e., the issue of diffusion kinetics of the radiogenic isotope during the flash heating; (2) whether or not the isotopic system has been affected by later thermal perturbations, such as syn- or post-formational fluidization, hot fluid migration, i.e., the issue of diffusion kinetics of the radiogenic isotope under hydro-thermal heating conditions. In addition, glassy matrix of pseudotachylytes is often found to be devitrified in nature, making proper interpretation of data further difficult.

As an alternative approach, Murakami and Tagami (2004) conducted FT analysis on zircons separated from nine segments of pseudotachylyte and host rocks from the Nojima fault zone, southwest Japan (Fig. 15). The pseudotachylyte age was determined as 56 ± 4 Ma, which is significantly younger than the age of background regional cooling (74 ± 3 Ma). Four gouge samples from the hanging wall (i.e., NT-UG 1-4) and two gouge samples from the footwall yielded ages that range from 65 to 76 Ma, with progressive reduction toward the pseudotachylyte layer. Both FT ages and track lengths systematically vary with the distance to the pseudotachylyte, suggesting that the zircon FT system was totally reset and subsequently cooled at ~56 Ma. This interpretation is supported by a pseudotachylyte formation temperature of ~750 – 1280°C as estimated on the base of feldspar melting textures and laboratory heating experiments (cf. subsection 4.2., see section 7 for tectonic implications.)

Subsequently, the zircon FT analysis has been applied to pseudotachylytes formed in different geologic settings: Asuke shear zone, central Japan (Murakami et al., 2006a), Tsergo Ri Landslide, Langtang Himal, Nepal (Takagi et al., 2007), Median Tectonic Line, southwest Japan (Takagi et al., 2010). Murakami et al. (2006a) performed FT and U-Pb analyses on zircons extracted from four segments of a pseudotachylyte and its granitic host rocks, collected from an ancient shear zone. The FT age of the pseudotachylyte is 53 ± 9 Ma, significantly younger than the age of host rock, 73 ± 7 Ma, which represents background regional cooling. Together with the track length information, the zircon FT system of pseudotachylyte was interpreted to have been totally reset and subsequently cooled at ~53 Ma. In addition, U-Pb ages fall in a range of ~67 – 76 Ma, suggesting that the formation ages of host rocks are indistinguishable from each other throughout the section.

6.3. Mylonites

Mica K-Ar (and $^{40}Ar/^{39}Ar$) and other thermochronological analyses have been applied to mylonitic rocks from a variety of tectonic settings (e.g., Mulch et al., 2002; Sherlock et al., 2004; Rolland et al., 2007; and references therein). Mulch et al. (2002) investigated muscovites and biotites from a crustal-scale mylonite zone of the Ivrea Verbano Zone, Italy, using furnace step-heating and in-situ UV-laser ablation $^{40}Ar/^{39}Ar$ thermochronology, in order to reveal the relationship between mica deformation and obtained $^{40}Ar/^{39}Ar$ ages. Two types of strongly-deformed muscovite have weighted mean $^{40}Ar/^{39}Ar$ spot ages of 147.7 ± 5.1 and 147.3 ± 6.5 Ma with a range of ~65 m.y., while the strongly-deformed biotite age is 123.3 ± 7.2 Ma with a range of 55 m.y. These are substantially younger than undeformed muscovite porphyroclasts that yield $^{40}Ar/^{39}Ar$ plateau ages of 182.0 ± 1.6 Ma. A correlation with field and microstructural observations suggested that the range of $^{40}Ar/^{39}Ar$ spot ages results from protracted cooling around argon T_C (Table 1) following mylonitization, by which micas were deformed and variably segmented at intra-grain scale. Microstructural segmentation should have dramatically reduced the effective length scale for argon diffusion and hence lowered the effective T_C by ~50 - 100 °C, causing diffusional argon loss.

Rolland et al. (2007) examined $^{40}Ar/^{39}Ar$ laser step-heating data of synkinematic phengite within low-grade Alpine shear zones, with a particular focus on microstructures of deformed and crystallized minerals. Four $^{40}Ar/^{39}Ar$ plateau ages obtained on phengite pressure shadow aggregates of feldspar porphyroclasts are 15.8 – 16 ± 0.2 Ma, indistinguishable between different shear zones of the "Mont Blanc back thrust". The ages are significantly younger than the biotite $^{40}Ar/^{39}Ar$ plateau ages of 20.0 - 63.7 Ma from undeformed granites in the region (Leloup et al., 2005). Pressure-temperature calculations of the shear zones indicate 5 ± 0.5 kbar and 400 ± 25 °C, and imply phengite growth close to argon T_C conditions (Table 1). Therefore, the phengite $^{40}Ar/^{39}Ar$ ages likely provide a close time estimate (or at least a minimum age) of the ductile deformation.

Thermochronological analyses also have been applied to ductile fault zones, in order to detect steady-state heat generation that likely occurs from a long-term fault motion averaged over geological time, with v = ~1 – 10 cm/y (Fig. 1; see also section 2). This type of heating phenomena is expected to form a broader, regional thermal anomaly across the fault zone compared to brittle regimes. Some regional metamorphic aureoles across convergent plate boundaries (or transcurrent shear zones) were attributable to such long-term heating at depths (Scholz, 1980), a typical example of which is the Alpine fault, New Zealand, where the continental collision is going on. A series of K-Ar dating unraveled a systematic decrease in age towards the fault, which was interpreted to represent an argon depletion aureole formed by frictional heating within a ductile regime (Scholz et al., 1979; and references therein; Fig. 16). This interpretation was based on the assumption that the pronounced uplift in the past 5 m.y. has been constant throughout the studied area, which was preferred by the geological structures of pre-5 Ma lamprophyre dikes and metamorphic folds.

However, a later apatite and zircon FT study demonstrated that the total amount of late Cenozoic exhumation shows an exponential increase toward the Alpine fault. Kamp et al. (1989) showed that apatite and zircon FT ages, which have greater thermal sensitivities (i.e., lower T_C) compared to

K-Ar systems (Table 1), exhibit a systematic trend to become younger toward the Alpine fault, consistent with previously-reported spatial distribution of K-Ar ages. The deduced differential, asymmetric uplift pattern is responsible for the exposure of a 13-25 km wide regional metamorphic belt immediately east of the Alpine fault known as the Alpine schist.

A similar age decrease toward a fault was reported for the Median Tectonic Line, southwest Japan, which was initially interpreted as a consequence of fault shear heating (Tagami et al., 1988). However, a recent FT and (U-Th)/He study (Sueoka et al., 2011) revealed a pattern of regional differential exhumation bounded by fault systems. Thus, the previously observed age variation near the Median Tectonic Line needs to be readdressed and may also be attributable to differential uplift.

6.4. Mineral veins

Flotte et al. (2001) conducted the first attempt of U-Th dating on fault-related calcites from the Corinth-Patras Rift, Greece. They studied the Xylokastro-Loutro fault and the Valimi fault, for both of which most of the deformation occurred on the major fault plane. A 3 cm thick crystalline calcite mat across the fault plane was collected from a karstic conduit of the Xylokastro-Loutro fault (Fig. 17, left). Flotte et al. (2001) dated a couple of subsamples, S1-a and S1-b: the former is located close to the conduit surface and thus should closely postdate the fault lock, whereas the latter is located higher and hence should be younger than the former. The obtained U-Th ages are 112.4 ± 0.4 (2 sigma) ka and 108.2 ± 1.0 ka for S1-a and S1-b, respectively, consistent with the stratigraphy. The Valimi fault was considered to be inactive and older than the Xylokastro-Loutro fault, on the basis of geological occurrences. The dated calcite was sampled from a fault breccias (Fig. 17, right; S2) and yielded an age of 382 ± 31 ka, substantially older than the age of Xylokastro-Loutro fault.

Boles et al. (2004) reported U-Th ages of calcite fault cement of the Refugio-Carneros fault, which marks a north-bounding structure to the

Santa Barbara basin. The calcite cement has resulted from leakage of fluids and hydrocarbons into the fault. It is considered to be synchronous with fault movement, as evidenced by pervasive crystal twinning and brecciation. In addition, the calcite fills fault-related openings and, hence, postdates the initiation of faulting. Five of eleven showed U-Th ages ranging from 107 to 420 ka, whereas other six yielded ages > 500 ka, i.e., the upper age limit of the technique. Because the texture of calcites indicate multiple stages of fluid flow and cementation, those ages probably reflect the final stage of crystallization and thus give minimum ages of the fluid flow. It therefore suggests that the Refugio-Carneros fault was active for at least several hundred thousand years in the Pleistocene. The paleotemperature was estimated as ~80 - 95°C on the basis of the homogenization temperature of fluid inclusions, implying that hot water leaked from depths of ~2 to 3 km along faults on the basin flank.

7. The Nojima Fault: an example for multiple application of

thermochronological methods in fault zones

The Nojima fault is a high-angle reverse fault dipping 83°SE, and runs along the northwestern coast of the Awaji Island, Hyogo Prefecture, Japan (Fig. 18). As a result of the 1995 Kobe earthquake (Hyogoken-Nanbu earthquake; M7.2), a >10 km long surface rupture was formed along the preexisting Nojima Fault, and has been investigated by means of various geological and geophysical disciplines (Oshiman et al., 2001; Shimamoto et al., 2001; Tanaka et al., 2007). The Nojima Fault Zone Probe Project was initiated by a series of drilling to penetrate the fault zone shortly after the 1995 Kobe earthquake. Several boreholes were drilled into an active fault system, including the Geological Survey of Japan 750 m (GSJ-750) borehole at the Hirabayashi site and University Group 500 m (UG-500) borehole at the Toshima site (Fig. 18): These two boreholes clearly penetrated the Nojima fault at 625.27 m and 389.4 m depths, respectively.

First in situ measurements of post-seismic permeability decay were conducted as a part of the Nojima Fault Zone Probe Project. The temporal change of permeability (Kitagawa et al., 2007), along with that of strain (Mukai and Fujimori, 2007) and electro-kinetic and hydraulic properties (Murakami et al., 2007) were measured by water injection experiments, and showed that the permeability was reduced for ~40 – 70 % over 8 years after the earthquake. This observation is consistent with the post-seismic recovery of fault strength, as expected from the fault-valve behaviour (Sibson, 1992; see sections 2.4. and 3.3).

The Nojima fault was also trenched at Hirabayashi: The exposed fault rocks consist of granitic cataclasite, a 2-10 mm wide pseudotachylyte layer and siltstone of the Osaka Group, from the hanging wall to the footwall. Otsuki et al. (2003) estimated the temperature of the pseudotachylyte formation as ~750-1280 °C, based primarily on the observation of melting of K-feldspar and plagioclase. For fission-track analysis, a 50-cm-wide gray fault rock consisting of the following four layers was sampled from the footwall toward the hanging wall (Murakami and Tagami, 2004; Fig. 15); (1)

greenish-gray gouge of footwall (NT-LG; ~20 mm wide), (2) pseudotachylyte (NT-Pta; ~2-10 mm wide), (3) gray gouge of hanging wall (NT-UG; ~30 mm wide) and (4) reddish granite (NF-HB1; ~20 mm wide). Note that the fault rupture formed in 1995 is located about 10 cm below the pseudotachylyte (Otsuki et al., 2003). The zircon fission-track data and their interpretation of those rocks were documented briefly in section 6.2.

Zircon FT analysis was also conducted on the boreholes GSJ-750 and UG-500 across the Nojima fault-zone (e.g., Tagami et al., 2001; Murakami et al., 2002; Tagami and Murakami, 2007). The results of GSJ-750 suggest that an ancient heating event into the ZPAZ took place around the fault, within ~25 m in both the footwall and hanging wall: whereas, the rocks of UG-500 record also an ancient heating into the ZPAZ in the hanging wall only, within <3 m from the fault. The age of initiation of last cooling, which marks the cessation of the secondary heating, was estimated by inverse modeling using the program Monte Trax (Gallagher, 1995; Fig. 11). Three partially-annealed samples near the fault in the GSJ-750 borehole

(GSJ-TH093, 099 and 103) yielded ages of initiation of last cooling as 35.0 ± 1.1 (1SE), 38.1 ± 1.7 and 31.3 ± 1.4 Ma; whereas a partially-annealed sample from the UG-500 (NF51-FTG14b) yielded a significantly younger age estimate of 4.4 ± 0.3 (1SE) Ma. Note that the time of the heating event may be determined with a certain confidence (resolution of an order of 1 My) while the maximum temperature heavily depends on the heating duration (see the section 5). The plausible heat source of these thermal events is heat transfer and dispersion via fluids within the fault zone, on the basis of one-dimensional heat conduction modeling as well as the positive correlation between the degree of fission-track annealing and deformation/alteration of borehole rocks. The result of in-situ heat dispersion calculation favors the upward flows of hot fluid along the fault zone from the deep crustal interior.

The authigenic illite K-Ar ages were determined on six outcrop samples of granitic cataclasite collected at Hirabayashi, including three fault gouge samples in close proximity to a pseudotachylyte layer, as well as on five UG-500 core samples at Toshima (Zwingmann et al., 2010b). At

Hirabayashi, six ages of the <2 mm fractions fall in the range of 56.8 ± 1.5 to 42.2 ± 1.0 Ma, whereas the <0.1 and <0.4 mm fractions of the three gouge samples are anomalously younger at 30.3 ± 0.9 to 9.1 ± 1.6 Ma. The former ages are significantly younger than the regional cooling age of granitic protolith (74 ± 3 Ma), and interpreted as the time of brittle faulting. The latter ages on finer fractions, which have lower effective T_C, are likely resulted from the secondary loss of radiogenic ^{40}Ar, as a probable consequence of thermal overprints in the core of fault zone caused by hot fluid circulation. At Toshima, three ages of <0.4 and <2 mm fractions containing less K-feldspar contamination range from 50.7 ± 1.2 to 45.0 ± 0.9 Ma, suggesting the time of brittle faulting similar to that of Hirabayashi.

The age data presented in the preceding paragraphs suggests that the Nojima fault had already been initiated at ~56 Ma as the in-plane fault offset at the crustal interior, much older than the previous estimate of ~1.2 Ma. It is likely, therefore, that the present Nojima fault system was formed by the Middle Quaternary reactivation of an ancient fault, as recognized widespread elsewhere (e.g., Holdsworth et al., 1997). This new reconstruction is consistent with the plausible deep origin (>15 km) of the Nojima pseudotachylyte, as constrained by fluid inclusion data (Boullier et al., 2001). The FT and K-Ar data also imply that the Nojima fault shows a temporal/spatial variation in terms of the thermal anomalies recorded in the fault rocks, likely reflecting the heterogeneous heat transfer via fluids migrated along the fault zone from the deeper crust. See Tagami and Murakami (2007) and Zwingmann et al. (2010b) for more details.

Another crucial aspect of the zircon FT data from the Nojima pseudotachylyte is that the mean shear stress τ can be estimated by combining the equations of frictional heat production, conductive transfer (i.e., equations (7) – (10)) and zircon FT annealing (equation (15)) (Murakami, 2010). The principle of his approach is:

(1) Temperature profiles (Fig. 7) are obtained through time t for a specific τ by assigning or assuming necessary parameters of equations (7) – (10), such as the density ρ, specific heat c, thermal diffusivity α, fault zone

width a, sliding velocity v and slip-duration t^*. Hence, thermal histories can be derived for each locality with a certain distance from the fault center, x.

(2) For the specific τ value, a mean FT length μ is predicted for each locality (i.e., each x value) using equation (15), by integrating the annealing effects for each of the divided intervals of the thermal history. Accordingly, a spatial profile of predicted μ is constructed against x.

(3) For a range of τ values, the spatial profiles of μ are computed and compared with measured FT length data, in order to search for the best fitting τ value.

Murakami (2010) estimated that the net shear work (i.e., $\tau v t^*$) of approximately 50 MPam had been consumed by the faulting event at ~56 Ma.

Watanabe et al. (2008) attempted to constrain the timing of fluid infiltration into the Nojima fault zone by measuring U-Th radioactive disequilibrium on calcite veins collected at 1484 m depth from the

UG-1800m borehole (~1 km southwest of UG500 site at Toshima; Fig. 18). The analyzed samples plot near the equiline in a ($^{230}Th/^{232}Th$) – ($^{234}U/^{232}Th$) diagram, and the U-Th age was accordingly estimated as 486 (+380, -190) ka using the IsoPlot program. The presence of radioactive disequilibrium in $^{234}U/^{238}U$ suggests that the age of calcite precipitation was younger than 1 Ma. These age constraints are consistent with the Middle Quaternary reactivation of the fault.

8. Summary and future perspectives

Fault-zone rocks are the result of long-term, repeated movements of faults reflecting both spatial and temporal variations in tectonic regimes. Age determination of fault movements, hence, places valuable constraints upon a variety of geoscientific issues at various crustal depths, such as the regional tectonic history, seismological frameworks, environmental assessments, hydrogeological scheme, etc. The technical and methodological advancements in thermochronology achieved so far, particularly in the last decade, enable to date fault-zone materials formed during the development of a fault: (a) K-Ar ($^{40}Ar/^{39}Ar$) dating on authigenic illite within a fault gouge, coupled with the evaluation of the influence of detrital contamination, (b) FT thermochronology on zircons from a pseudotachylyte layer, fault gouge and associated deformed rocks, using the annealing kinetics based on the flash- and hydrothermal-heating experiments at laboratory, and (c) $^{40}Ar/^{39}Ar$ laser-probe dating on pseudotachylyte matrix may also work if the samples are well degassed during natural flash-heating, (d) $^{40}Ar/^{39}Ar$ laser-probe dating on recrystallized micas in mylonitic rocks, and (e) U-Th dating of carbonate veins. In addition, systematic drilling into active fault zones, such as the Nojima Fault Zone Probe Project, provided access to fresh borehole rocks across the fault zone and promoted fault-zone thermochronology, as an important part of the material seismology.

Further perspectives with respect to both methodology and application are opened by:

(a) laboratory flash-heating experiments for other low-temperature thermochronological systems, such as apatite FT, apatite and zircon (U-Th)/He systems. This will encourage comprehensive thermochronological analyses of fault zones, by focusing upon a variety of fault-zone rock types that are formed at different depths and temperatures (Fig. 1).

(b) combined use of thermochronological systems with substantially different activation energies, such as zircon FT and (U-Th)/He (Table 1), may allow to estimate duration and temperature of a heating episode, which will help to identify its heat source (see the section 3). The zircon FT system is more retentive than the zircon (U-Th)/He for geological time scale (as expressed in their T_C values; Table 1), whereas zircon (U-Th)/He appears more retentive than zircon FT for much shorter timescale (e.g., an order of seconds). A reliable interpretation of faulting processes from Zr data requires laboratory flash-heating experiments of the zircon (U-Th)/He system and additional case studies on fault rocks (e.g., Yamada et al., 2011).

(c) Systematic analyses of ultra-deep boreholes penetrating active faults at depths >3 km, such as the case of IODP seismogenic-zone drilling at the Nankai Trough. A combination of illite K-Ar, zircon FT and (U-Th)/He, and carbonate U-Th analyses, coupled with other geoscientific constraints, will shed new light on our understanding of the thermo-mechanical processes in the seismogenic-zone faults.

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Figure captions

Fig. 1. Synoptic model of a suture zone after Scholz (1988), showing the schematic section across fault zone, along with seismic and mechanical parameters against temperature (or depth).

Fig. 2. Two types of mineral transformations within fault gouges to form illites suitable for K-Ar dating (Haines and van der Pluijm, 2008). (A) Illitization of illite-smectite observed in clay-rich fault gouges, which results in higher contents of illite in illite-smectite compared to the wall rocks. (B) Neocrystallization of authigenic discrete $1M_d$ illite in fault gouges.

Fig. 3. Representative photographs of clay particles from the Nojima fault zone (Zwingmann et al., 2010b): secondary and transmission electron images of a sample from the cataclasite zone (A and B). I stands for illite, qz for quartz, C for chlorite; mineralogy was confirmed by EDS.

Fig. 4. Photographs of pseudotachylytes collected from granitic mylonites in the Fuyun fault zone, northwest China (Shimamoto et al., 1992; Lin, 1994). (A) Newly-grown rectangular sanidines and ultra-fine microlites. Field of view is 0.3 mm wide. (B) Banded structure suggesting fluidization. Field of view is ~1.5 mm wide. (C) Newly-grown microcrystalline plagioclases that show radial distribution. Field of view is 4.8 mm wide. See also Fig. 15 for the photograph and sketch of the Nojima pseudotachylyte layer.

Fig. 5. Photographs of vein developments in fault zones (Cox, 2010; see also references therein). (A) Near-optimally oriented reverse fault with associated extension vein arrays, from Wattle Gully gold mine, Australia. Field of view is 1-m wide. (B) Ductile shear zone with associated sigmoidal calcite extension veins, from Columbia Icefields Highway, Canada. Field of view is 2-m wide.

Fig. 6. Three models of thermal processes relevant to the fault zone development: (a) regional geothermal structure across the fault zone and background thermal history of studied province bounded by fault systems, (b) frictional heating of wall rocks by fault motions, and (c) heating of host rocks by hot fluid advection in and around the fault zone.

Fig. 7. (a) Profiles of temperature increase during faulting, and (b) profiles of temperature decay after a faulting event for a shear work of 50 MPam. The temperature increase ΔT can be estimated from the equations of Carslaw and Jaeger (1959) and Lachenbruch (1986), assuming the duration of the local seismic slip is 5 seconds. The temperature was above 1000 °C in a faulting zone 8 mm in width (blue boxes) for a 30-second interval following the onset of faulting. After Murakami (2010); color illustration by M. Murakami.

Fig. 8. A schematic outline of the closure temperature and time model

(Dodson, 1973). T_O, initial temperature of a cooling rock; T_C, closure temperature; t_C, closure time; D/P, daughter/parent ratio. The transitional feature between the closure and opening of the system is graphically expressed on the diagram (bottom) as the intermediate zone between the regions of 100% loss (on the older or higher temperature side; where daughters are diffused so rapidly that they are not effectively accumulated at all in the mineral) and 100% retention (on the younger or lower temperature side; where daughters are not effectively diffused at all and thus are regarded to be completely accumulated within the mineral). See text for explanation.

Fig. 9. Schematic age-elevation diagram across the partial retention zone (PRZ) for a specific thermochronometer. Geothermal gradient of 30°C/km is assumed. (a) An idealized age profile in the upper continental crust where substantial exhumation-related cooling has continued at a constant rate for an extended period of time (i.e., >20 my in the present case). The gradient

above the PRZ approximately gives an average rate of exhumation-related cooling (i.e., 0.3 km/my, or 9°C/my). (b) In tectonically stable regions, the age profile is primarily governed by the vertical temperature profile, forming a characteristic downward age-decrease within the PRZ. The age profile above the PRZ reflects the rate of exhumation-related cooling that precedes the time of tectonic stability (i.e., 0.3 km/my, or 9°C/m before ~80 Ma). (c) If the extended period of tectonic stability is followed by a recent, rapid exhumation (or unroofing) event (i.e., for ~3 km in the past ~3 my), the characteristic profile of age decrease can be preserved as a fossil PRZ.

Fig. 10. Arrhenius plot showing the design points of the laboratory annealing experiments of spontaneous fission tracks in zircon as well as contour lines for the fitted fanning model extrapolated to geological time scale (Tagami et al., 1998; see also Yamada et al., 1995; Galbraith and Laslett, 1997). Three temperature zones can be defined as first-order approximation of fission-track annealing: the total stability zone (TSZ) where tracks are

thermally stable and hence accumulated as time elapsed; the partial annealing zone (PAZ) where tracks are partially stable and slowly annealed and shortened; and the total annealing zone (TAZ) where tracks are unstable and fade soon after their formation. Note that the zircon PAZ is defined here as a zone of mean lengths of ~4 to 10.5 μm and is shown as a white region intervened by TSZ and TAZ. Also shown are the time-temperature envelope of "flash heating" experiments that simulate the shear heating of fault motion (Murakami et al., 2006b); data points of four deep borehole samples subjected to long-term natural annealing, which were used to test the extrapolation of the annealing model (Tagami et al., 1998; Hasebe et al., 2003); and a box that indicates the time-temperature condition estimated for the higher temperature limit of the zircon PAZ on Crete (Brix et al., 2002). After Tagami (2005).

Fig. 11. Examples of thermal history modeling using fission-track length and age data (Tagami and Murakami, 2007). (Top) Modeling results for

representative runs of the two zircon samples (a, b) using the genetic algorithm of the Monte Trax program (Gallagher, 1995). Two grey boxes for each diagram show initial time-temperature constraints given, from where time-temperature points were selected at random when the modeling started. 50 modeled runs at the final stage of iterations are shown by red lines for each sample. The solid black lines represent the average of the modeled thermal histories. The 95% confidence limits for the kink points of the history were shown as open black boxes. (Bottom) The observed track length data plotted as a histogram and predicted track length distribution curve for best data fitting model.

Fig. 12. A representative temperature change vs time during flash heating experiments using the graphite furnace. The signal intensity of an infrared radiation thermometer (IR-II) is plotted against time of annealing. The signal is kept stable until when zircon grains are dropped into the preheated graphite cuvette, but changes significantly as soon as the grains enter the

monitoring area by IR-II. After a scheduled duration, the heating is ended by rapid cooling of the entire graphite cuvette. The annealing effect during the cooling (i.e., gray area) can be assessed by using the kinetic model of Tagami et al. (1998), and is equivalent to an isothermal heating for <0.25 sec at the scheduled temperature. After Murakami et al. (2006b); color illustration by M. Murakami.

Fig. 13. Relationship between crystallinity index (Kubler Index) and K-Ar ages of mica clay minerals from fault gouges of the Akaishi Tectonic Line, central Japan. BOT-N plotted on the uppermost point is a sample from the protolith, i.e., the Sambagawa Metamorphic Rocks. Kubler Index and K-Ar ages show a negative correlation. The data trend can be fitted by a hyperbolic curve that converges to ~15 Ma, which was interpreted as the time of a hydrothermal alteration event associated with fault motion. After Tanaka et al. (1995).

Fig. 14. Illite K-Ar ages of a gouge from the Sierra Mazatan detachment fault, Sonora, Mexico, plotted against detrital illite ($2M_1$) contents (Haines and van der Pluijm, 2008). Age is shown in the form of "$e^{\lambda t}$ -1" (where λ is the total decay constant of ^{40}K and t is the time elapsed), which is linearly related to the detrital illite contents. The ages of authigenic and detrital illite populations, which were mixed in the fault gouge, can be estimated as 14.9 and 18.5 Ma, respectively, by extrapolating the observed age trends to 0% and 100 % intercepts.

Fig. 15. The photograph (A) and sketch (B) of a sampled fault rock section in the Hirabayashi trench, with a plot of mean age, mean length and length distributions of zircon fission tracks. NT-PTb is also plotted (open circle). The blue boxes represent the mean age and length of NF-HB2 (the Ryoke host rock sample). The sample from the pseudotachylyte layer (NT-PTa) has an age significantly younger than that of initial cooling of the Ryoke host rock sample (NF-HB1 and -HB2). Photograph (C) shows a "syringe-shaped"

tracks found in NT-UG1, which shows the evidence of partial annealing. Error bars are ±1 SE. After Murakami and Tagami (2004).

Fig. 16. Biotite and whole-rock K-Ar ages east of the Alpine Fault, New Zealand, between Arthur's Pass and Haast Pass (Scholz et al., 1979; see also the quoted references therein). Overall, the ages show a systematic decrease towards the fault. The age trend was interpreted as an argon depletion aureole, which was formed by the frictional heating at the ductile deformation regime.

Fig. 17. Photographs of the calcite samples dated by the U-Th method. 3a: Post-tectonic calcite of Xylokastro-Loutro fault. S1-b is located higher and thus expected to be younger than S1-a, which was confirmed by the dating. 3b: Syn-tectonic calcite collected from the Valimi fault. It is mixed with brecciated limestones of the footwall. After Flotte et al. (2001).

Fig. 18. A geologic map showing sampling localities of the Nojima fault (after Tagami and Murakami, 2007). MTL, Median Tectonic Line; ISTL, Itoigawa-Shizuoka Tectonic Line.