TECTONICS, VOL. 11, NO. 2, PAGES 348-359, APRIL 1992

ROLE OF MELT IN THE FORMATION OF A DEEP-CRUSTAL COMPRESSIVE SHEAR ZONE: THE MACLAREN GLACIER METAMORPHIC BELT, SOUTH CENTRAL ALASKA

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Abstract. The Maclaren Glacier metamorphic belt is an exhumed portion of a deep-crustal shear zone where hot, upper amphibolite facies rocks were emplaced over cooler, lower grade rocks. It is located in south central Alaska within the collision zone between terranes previously accreted to North America and the Wrangellia superterrane. Crosscutting relationships and orientations of thin granitoid sills within the hanging wall show that melt was repeatedly intruded into the shear zone during overall compression. In addition, a 1 km thick tonalite sill (the Valdez Creek tonalite) was emplaced into the shear zone while it was active; this conclusion is based on the presence of a well-developed foliation within the sill which is concordant to the fabric of the shear zone, the alignment and tiling of plagioclase laths, the presence of highly strained mafic enclaves within a matrix which shows no evidence of any subsolidus deformation, and the preservation of a delicate magmatic texture which indicates that a magmatic fluid was still present after deformation in the sill had ceased. A one-dimensional thermal model for sill-shaped plutons of varying thicknesses, which are emplaced into country rocks at different initial temperatures, indicates that melt can be present for long periods of time in the deep crust (>1 m.y.). An important factor controlling the length of time that a sill remains molten is the temperature of the surrounding rocks which, in the lower crust, can be as high as the solidus temperature of granitoid melts. If the amount of time for melt to crystallize is sufficiently long, potentially large amounts of strain will be accommodated by zones containing melt due to the low strength of melt compared to rock. In the Maclaren Glacier metamorphic belt, the amount of time needed for the Valdez Creek tonalite to fully crystallize is calculated to be about 90,000 years. This is a minimum value since the original thickness of the sill is not known. Nevertheless, using the 90,000 year value and assuming that the bulk of convergence between North America and Wrangellia was concentrated within the Valdez Creek tonalite, a displacement of at least 10 km could have been accommodated across the sill while melt was present.

INTRODUCTION

The close association of foliated plutonic rocks with regionally metamorphosed and deformed supracrustal rocks has long been recognized in the Late Cretaceous to Early Tertiary crystalline complex which extends in a south-southeasterly direction from south central Alaska through British Columbia to western Washington and into northern Idaho [Buddington and Chapin, 1929; Forbes et al., 1974; Brew and Ford, 1978; Hutchison, 1982; Crawford et al., 1987; Hyndman et al., 1988]. In southeastern Alaska and western Canada, this complex is known as the Coast Plutonic Complex (Figure 1a). Most workers agree that the igneous rocks have been intruded into relatively deep levels of the crust, of the order of 20 km or deeper [Crawford and Hollister, 1982; Crawford et al., 1987; Zen and Hammarstrom, 1984; Stowell, 1989; Brew et al., 1989]. In a plate tectonic framework, the structural position of this plutonic complex is extremely important. Most of southern Alaska and western Canada is assumed to have been formed throughout the Mesozoic by a process of collision of continental fragments with the North American craton [Coney et al., 1980; Monger et al., 1982; Jones et al., 1982]; the Coast Plutonic complex occurs within the suture zone between one of these fragments, the Wrangellia superterrane (as defined by Saleeby [1983]), and the previously accreted Yukon-Tanana terrane in the north and Stikine terrane in the south. The Maclaren Glacier metamorphic belt occurs in southcentral Alaska between a fragment of the Wrangellia superterrane (herein Wrangellia) and the Yukon-Tanana terrane [Nokleberg et al., 1985]. Regional deformation, metamorphism, and intrusion of the igneous rocks of the plutonic complex are likely to be associated with the accretion of Wrangellia [Monger et al., 1982].

Hollister and Crawford [1986], using examples from southeastern Alaska and northwestern British Columbia, emphasized the importance of the presence of melt during regional deformation and metamorphism. According to Hollister and Crawford, melt may intrude into the base of an orogen that is being actively deformed by crustal shortening (that is, during overall compression). As a result, the lower crust is weakened, and strain will concentrate within melt-bearing shear zones.

In order for this process to affect major displacements in the lower crust, melt would be required to remain in the crust for substantial periods of time. In this paper, first we document the occurrence of melt having been intruded into an active deep-crustal thrust shear zone, informally referred to as the Valdez Creek shear zone of the Maclaren Glacier metamorphic belt; second, we evaluate the potential for (and the implications of) melt being present in the lower crust for long periods of time; and, finally, we calculate the minimum length of time needed for a sill to crystallize within the Valdez Creek shear zone and estimate the amount of displacement the sill could have accommodated while partially molten.

GEOLOGIC SETTING

The Maclaren Glacier metamorphic belt, southcentral Alaska, was first described in detail and mapped at a 1:63,360 scale by Smith [1981]. Forbes et al. [1974], Eisbacher [1976], Nokleberg et al. [1985], and others have interpreted it to be a portion of the Coast Plutonic Complex of southeast Alaska and British Columbia which has been offset by dextral motion along the Denali fault (Figure 1). To the south, rocks of the Maclaren Glacier metamorphic belt are separated from the volcanic rocks of Wrangellia by the Broxson Gulch thrust fault [Nokleberg et al., 1985]. The Broxson Gulch thrust is truncated by the Denali fault to the east and merges with the Talkeetna thrust fault [Csejtey et al., 1978] to the west (Figure 1).

The Maclaren Glacier metamorphic belt is composed of variably metamorphosed pelitic rocks interbedded with minor amounts of volcaniclastic rocks and intruded by granitic plutons that display a range of deformation fabrics.

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Paper number 91TC02907.
0278-7407/92/91TC-02907$10.00

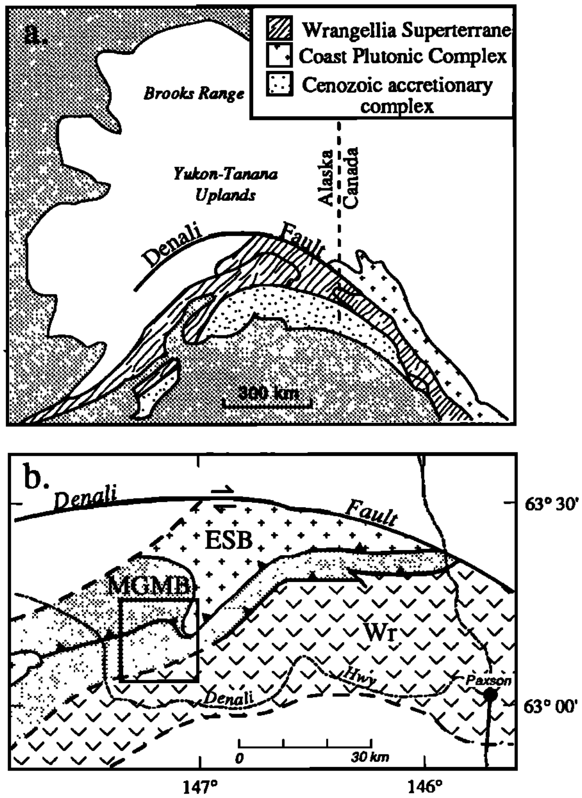


Fig. 1. (a) Distribution of tectonostratigraphic terranes in southern Alaska and northwestern Canada (adapted from Saleeby [1983]). Box encloses the area of Figure 1b. (b) Simplified map from south central Alaska showing the major geologic divisions. Abbreviations are MGMB, Maclaren Glacier metamorphic belt; ESB, East Susitna batholith; Wr, Wrangellia; BGT, Broxson Gulch Thrust; and TTF, Talkeetna thrust fault. Box encloses the area of Figure 2.

Metamorphic grade within the pelites increases from lower greenschist grade rocks in the south to amphibolite grade rocks in the north (Figure 2). The metamorphism is inverted; that is, the highest-grade rocks are found in the structurally highest position (Figure 2b). The inverted nature of the metamorphism was accomplished by thrusting of amphibolite grade rocks across the 4-5 km thick Valdez Creek shear zone (Figure 2b), and over the low-grade rocks to the south. A 1000 m thick tonalite sill, informally referred to as the Valdez Creek tonalite, occurs within the shear zone and has a well-developed foliation parallel to that of the wall rocks.

Deformation, metamorphism, and intrusion of the igneous rocks within the Maclaren Glacier metamorphic belt are approximately coeval and occurred in the Late Cretaceous and Early Tertiary [Smith, 1981] (C. Davidson, manuscript in preparation, 1992). Protolith ages for the metamorphic rocks are unknown; however, Smith [1981] correlates the protoliths of the metamorphic rocks to the unmetamorphosed rocks in the extreme southern portion of the study area (Figure 2b) on the basis of the proximity of the unmetamorphosed rocks and similarities in their trace element geochemistry. The age of these rocks is poorly constrained due to the lack of fossils, but they may be as old as Upper Triassic [Smith, 1981]. Eisbacher [1976] and many others have correlated these rocks with the Gravina-Nutzotin belt of Berg et al. [1972]; this belt is Jurassic to Middle Cretaceous in age.

THE VALDEZ CREEK SHEAR ZONE

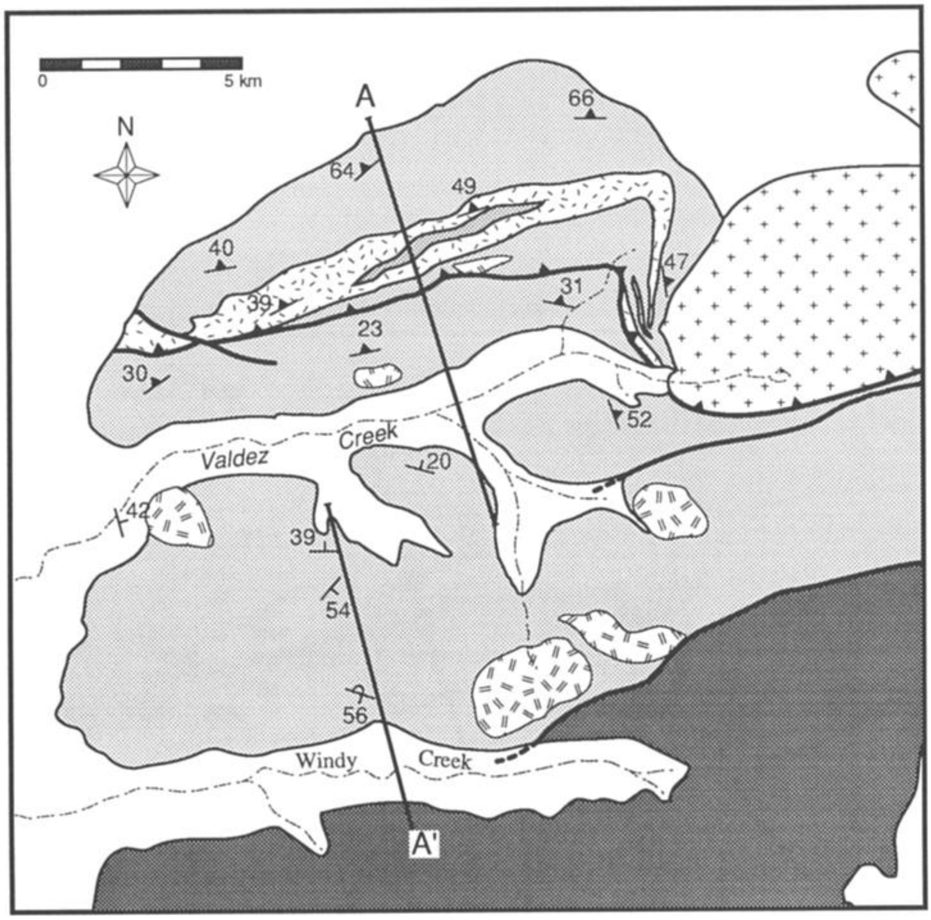
The most striking feature of the Maclaren Glacier metamorphic belt is the occurrence of high-grade metamorphic rocks thrust over lower-grade rocks; thrust motion was accommodated by the Valdez Creek shear zone. Within the shear zone is a boundary (shown by the heavy solid line with teeth in Figure 2a) that marks a discontinuity in texture, mineralogy, and deformational history. The region to the north of the discontinuity is referred to as the hanging wall and the region to the south as the footwall. Most of the strain for the exhumation of the hanging wall was accommodated by distributed shearing within the shear zone; however, the textural discontinuity probably represents a plane of discrete offset in order to juxtapose rocks of different texture.

The Hanging Wall

Two deformation events (D1 and D2), responsible for the formation of two distinct fabrics (s1 and s2), are recorded in the hanging wall rocks (Figure 2). Fabric s1 is a north dipping, bedding parallel foliation. L1 mineral stretching lineations are present in some quartz rich layers and have a downdip orientation. Overgrowing s1 and L1 in random orientations are kyanite + garnet pseudomorphs of staurolite. Approaching the Valdez Creek shear zone from the north, a second foliation (s2) overprints s1 near the upper contact of the Valdez Creek tonalite (Figure 2b). Within a narrow transition zone from s1 to s2, s2 is defined by an axial planar cleavage of crenulation folds (Figure 3). Further south, the s2 fabric rapidly becomes more intense, completely transposing the earlier s1 fabric. Metamorphic minerals such as kyanite have been deformed and rotated into the direction of s2, defining a new downdip L2 lineation (Figure 4); near the upper contact of the Valdez Creek tonalite, muscovite pseudomorphs of kyanite have been strung out parallel to L2.

Sense-of-shear indicators. A number of criteria were used to determine a consistent top-to-the-south sense of shear during both D1 and D2. In the northernmost region, which was unaffected by D2, reliable sense-of-shear indicators are difficult to find. This is due in part to the growth of metamorphic minerals after the fabric had already formed. The macroscopic indicators, however, are unambiguous: Quartz veins and thin granitoid sills have been deformed and boudinaged into asymmetric boudins that consistently give a top-to-the-south shear sense. Within the region affected by D2, macroscopic top-to-the-south sense-of-shear indicators include asymmetric boudins and asymmetric folds; microscopic sense-of-shear indicators include the rotation and recrystallization of kyanite crystals into the s2 foliation plane and asymmetric pressure shadows around garnet.

Granitoid sills of the hanging wall. A notable feature of the hanging wall is the occurrence of thin, foliation parallel trondhjemitic sills which range in thickness from a few millimeters up to tens of centimeters. The 1 km thick Valdez Creek tonalite sill is described separately below. Among the thinner sills within the hanging wall, several generations have been recognized. The sills which occur in the northernmost region are unaffected by D2 and are parallel to s1. Within the kyanite schists overprinted by D2, at least two generations of sills are observed. The first generation sills are folded by crenulation folds of the incipient s2 fabric (Figure 3) and are cut by a second generation of sills (Figure 5). Within the region where s2 has totally transposed the s1 fabric, all the sills are oriented parallel to s2. A striking feature of both generations of these sills is the lack of any significant fabric within them. In thin section, both quartz and feldspar display an igneous texture with little evidence for any subsolidus deformation (Figure 5b).



EXPLANATION

Metamorphic and Sedimentary Rocks. Upper amphibolite facies pelitic schist and gneiss in the north (locally migmatitic); amphibolite to greenschist facies pelitic schist in the central portion; sub-greenschist to unmetamorphosed flysh in the south.

E. Susitna batholith. Granodiorite to quartz diorite; massive, moderate foliation at margins.

Tonalite sill. Tonalite to quartz diorite, well-developed foliation.

Small stocks. Variable intermediate to mafic compositions.

Wrangellia volcanics. Mixed subareal to subaqueous flows of tholeiitic basalt.

Structural and textural discontinuity. Marks the boundary between the hanging wall (to the North) and the footwall (to the South). The area affected by the Valdez Creek shear zone is shown in section A-A'.

40 Foliation

37 Bedding

Fig. 2a. Simplified map of the Maclaren Glacier metamorphic belt (modified from Smith [1981]).

The Footwall

Only one deformation event is recorded in the rocks of the footwall within the Valdez Creek shear zone; these rocks are characterized by a pervasive foliation (s2) defined by the preferred orientation of micas. Metamorphic amphibole commonly forms a mineral lineation which has a downdip orientation (see stereonet in Figure 2b). The highest-grade metamorphic rocks within the footwall (amphibolite facies) are found in the highest structural position. Metamorphic grade rapidly decreases to the south, reaching biotite grade in about 2 km. Metamorphic minerals are easily identified in hand sample, and the inverted nature of the metamorphism is readily apparent in the field. The isograds shown in Figure 2b are mapped on the basis of first appearance of the isograd phase and are not necessarily reaction isograds.

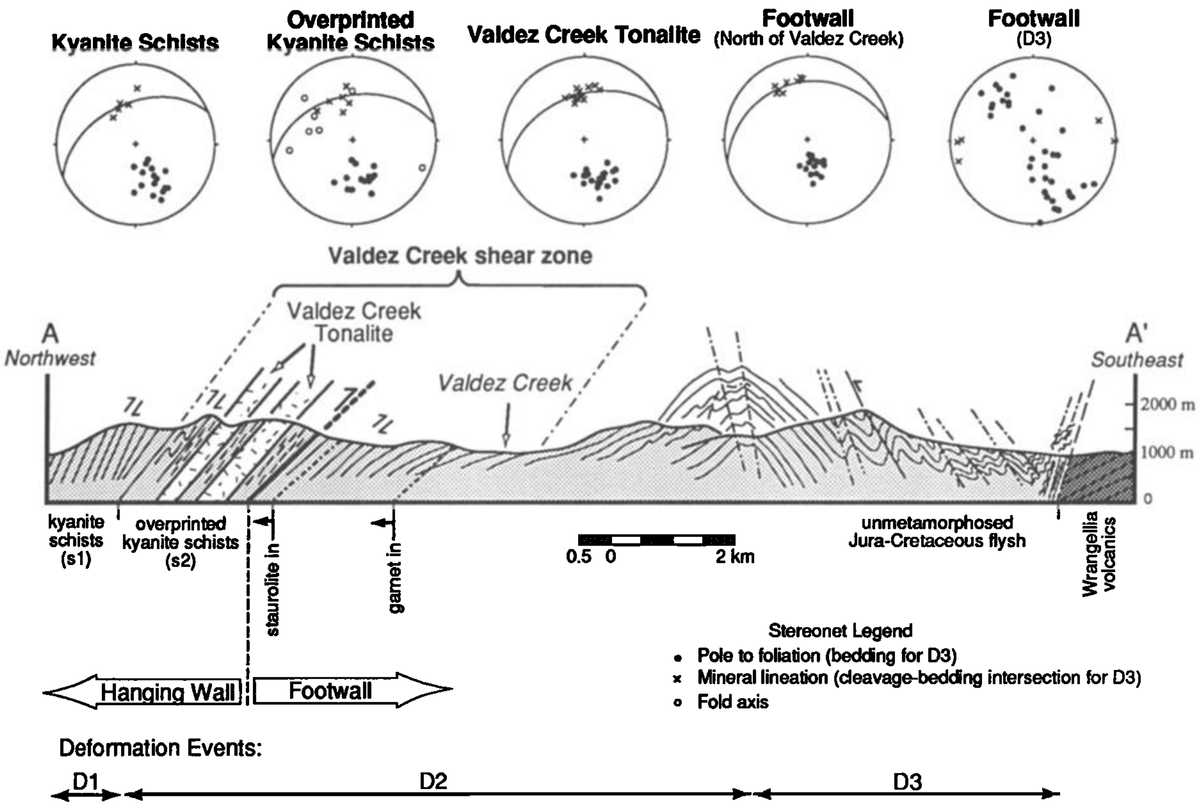


Fig. 2b. Cross section A-A' from Figure 2a. Stereonets are equal-area lower hemisphere projections; great circles are the traces of the average foliation plane. The arrows along the bottom of the figure mark the spatial distribution of three deformation events (D1-D3) recorded within the rocks of the Maclaren Glacier metamorphic belt. D1 and D2 are discussed in the text; D3 is characterized by tight to open north facing folds related to the obduction of Wrangellia (C. Davidson, manuscript in preparation, 1992).

Sense-of-shear indicators. Deformation was coincident with the growth of metamorphic minerals throughout much of the footwall. Staurolite, garnet, and amphibole commonly have s-shaped inclusion trails, and many garnets have asymmetric pressure shadows (Figure 6). A wealth of microscopic sense-of-shear indicators such as those shown in Figure 6 exist in the footwall; by using the criteria outlined by Simpson and Schmid [1983], a reliable top-to-the-south sense of shear is obtained for the deformation in the footwall. The coincidence of sense of shear and fabric orientation in both the hanging wall and footwall supports the conclusion that deformation in the footwall was coincident with the formation of s2 in the

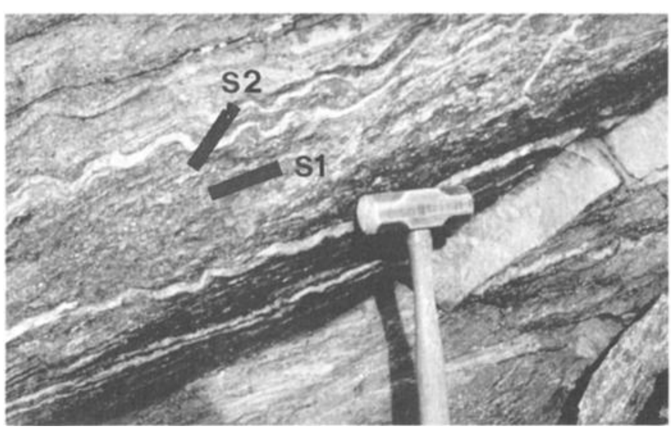


Fig. 3. Crenulation folding (D2) of thin trondhjemitic sills in the hanging wall intruded parallel to s1. North is to the left in the photograph.

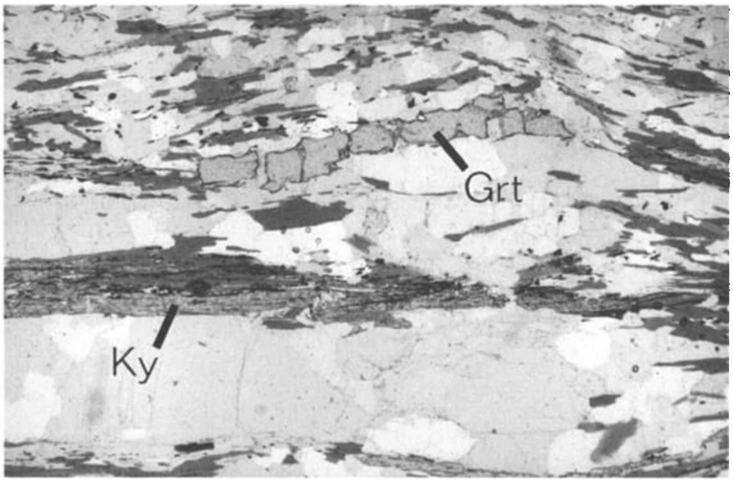


Fig. 4. Photomicrograph (5 mm across) of deformed kyanite (Ky) and garnet (Grt) from the southern portion of the hanging wall affected by the Valdez Creek shear zone.

a

b

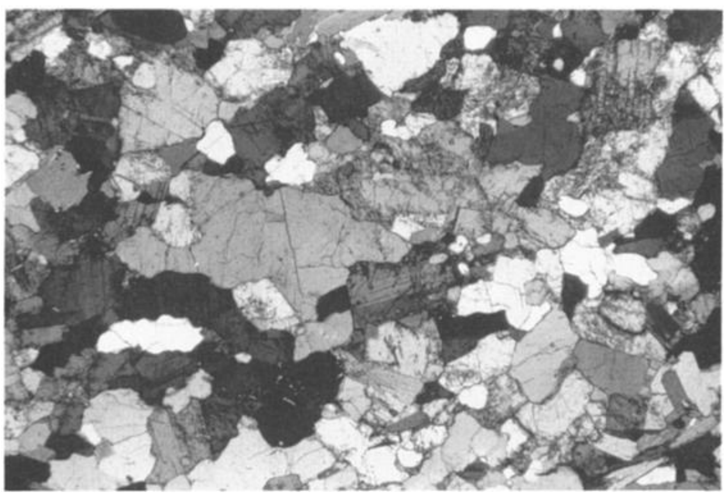


Fig. 5. Granitoid sills of the hanging wall. (a) First generation sills parallel to s1 cut by a second generation sill. (b) Photomicrograph (5 mm across, crossed nicols) displaying the typical igneous texture of the granitoid sills.

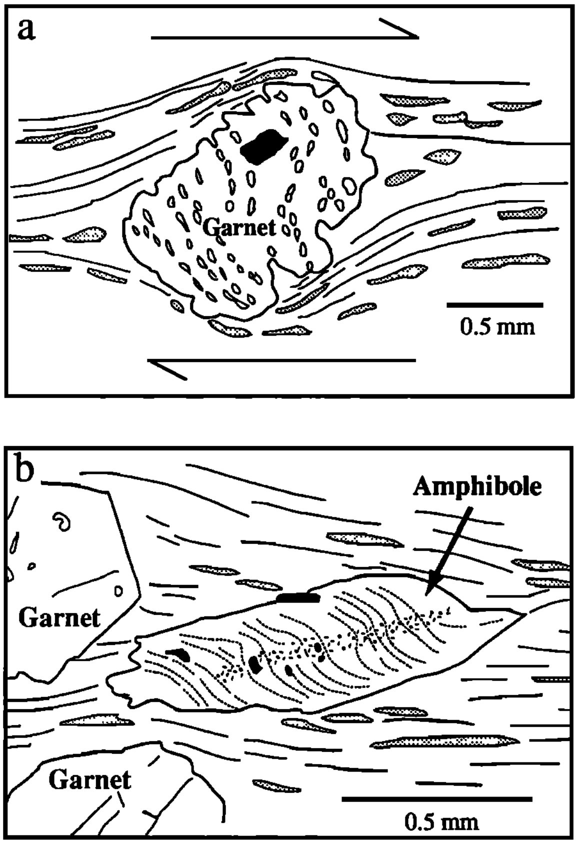


Fig. 6. Line drawings of thin sections demonstrating the synkinematic growth of porphyroblasts in the footwall. (a) S-shaped inclusion trails of quartz in garnet. (b) S-shaped inclusion trails of graphitic material in amphibole. The mineral with the stippled pattern is biotite. North is to the left in both drawings.

hanging wall. Therefore s2 of the hanging wall and the fabric of the footwall are grouped under the same deformation event (D2).

Granitoid sills of the footwall. Thin granitoid sills are found in the highest structural levels of the footwall. They all lie within the foliation plane defined by the metamorphic rocks, and they commonly have been deformed into asymmetric boudins which give the same sense of shear as the microscopic sense-of-shear indicators discussed above. These sills, like those in the hanging wall, retain an igneous texture.

Metamorphism. The most notable feature of the metamorphism in the footwall is the presence of a pronounced inverted metamorphic field gradient. The highest-grade rocks contain the pelitic mineral assemblage staurolite + garnet + biotite + quartz + plagioclase + ilmenite + rutile + graphite ± kyanite ± muscovite ± pyrite. Structurally below these rocks but north of the garnet isograd, the stable pelitic assemblage is garnet + biotite + quartz + plagioclase + ilmenite + graphite ± chlorite ± muscovite ± rutile; in more calcic rock types, this assemblage includes Ca-amphibole ± epidote ± calcite. South of the garnet isograd, the pelitic assemblage is biotite + quartz + plagioclase + graphite ± chlorite ± muscovite ± ilmenite ± rutile. These pelitic assemblages, when considered together, suggest an increase in metamorphic temperatures from south to north (From ~400°C to ~600°C) at moderate to high pressure (>5 kbar) [Winkler, 1979].

THE VALDEZ CREEK TONALITE

The composition of the Valdez Creek tonalite is relatively consistent, containing plagioclase (48%), quartz (20%), biotite + hornblende (25%), clinopyroxene (5%), and K-feldspar (<2%) as the major phases; accessory phases include sphene and apatite. The sill possesses a well-developed foliation defined by the preferred orientation of biotite (Figure 7), which is parallel to the fabric of the shear zone. As shown in Figure 7, plagioclase laths also have a preferred orientation and are commonly tiled suggesting they have been rotated while in a partially molten matrix [Blumenfeld and Bouchez, 1988; Hutton, 1988; Paterson et al., 1989]. This alignment of plagioclase laths defines a weakly developed lineation; measured foliations and lineations of the sill are plotted in Figure 2b.

Mafic enclaves found within the sill have been deformed into elongate ellipsoids (Figure 8) with X/Z aspect ratios ranging from 8 to 18 and X/Y from 2 to 8. The high finite strain recorded by the enclaves is not consistent with the microscopic fabric of the sill. The Valdez Creek tonalite retains an igneous texture, and there is no evidence that appreciable subsolidus deformation occurred (Figure 7). In addition, K-feldspar of the sill is completely anhedral and occurs along grain boundaries filling the interstices between the phases (Figure 7c); in order for this texture to have been preserved, deformation within the sill must have ceased prior to the crystallization of the K-feldspar.

The lack of subsolidus deformation features, the presence of deformed enclaves, the well-developed foliation, the tiling of plagioclase laths, and the preservation of the intricate K-feldspar texture show that the sill was being deformed while still in a partially molten state. In addition, the correspondence of fabric orientations within the sill, and D2 features of the hanging wall and footwall, implies that the sill was emplaced while the Valdez Creek shear zone was active.

THERMAL MODELING

Wells [1980], Lux et al. [1986], DeYoreo et al. [1989a], and others have discussed the importance of heat supplied by magma in the thermal structure of evolving mountain belts. These studies have predominantly dealt with the relationship between metamorphism and magmatism, with particular attention paid by some to the role of magmatism in low-pressure/high-temperature metamorphic belts [e.g., DeYoreo et al., 1989a]. Implicit in such studies is the assumption that the length of time for a pluton to crystallize is short relative to the amount of time needed for large amounts of deformation to take place. However, melt intruded into the hot lower crust will take considerably longer to crystallize than melt intruded into the cold upper crust. The possibility of melt being present for long periods of time in the deep crust has important implications for the deformational history of active mountain belts. If melt is intruded into an actively deforming region of the crust, then strain will be concentrated where melt is present due to the low strength of melt compared to rock. If the length of time for crystallization of the melt is sufficiently long, then large displacements could occur within melt-bearing shear zones under natural strain rates.

Below we introduce a numerical thermal model which we use to calculate crystallization times of melts under different physical conditions. The model is then applied to the Valdez Creek shear zone in order to calculate crystallization times of the Valdez Creek tonalite and to calculate thermal profiles within the footwall.

a

b

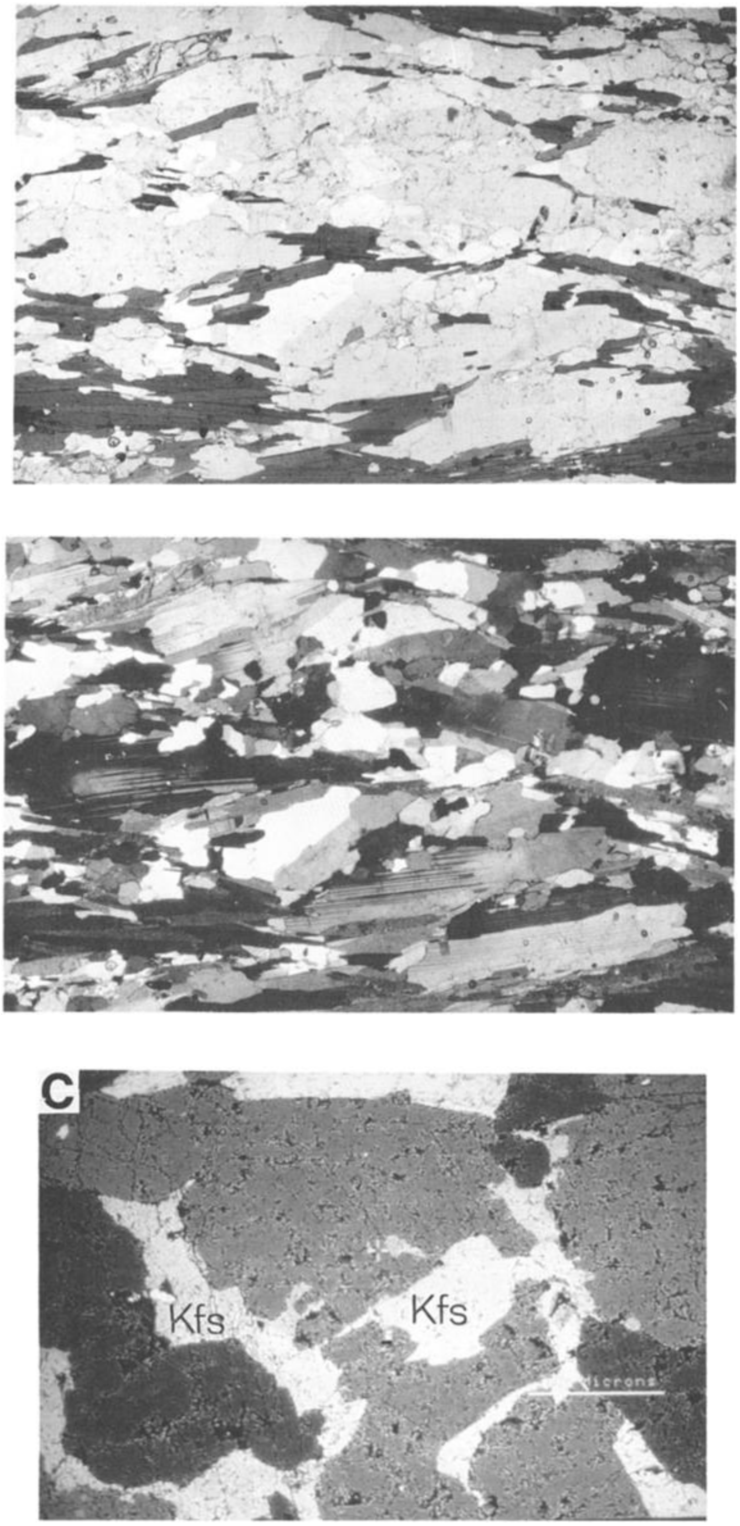


Fig. 7. Microscopic features of the tonalite sill. (a) Photomicrograph (5 mm across) showing the well-developed foliation of the tonalite sill. Note the tiling of plagioclase laths (lower right) and lack of subsolidus deformation features. Thin section is cut parallel to the mineral lineation and perpendicular to foliation. North is to the left. (b) Same as Figure 7a, but with crossed nicols. (c) Backscattered electron image (scale bar is 300 μm); note the delicate texture formed by the K-feldspar (Kfs) indicating that it crystalized after deformation ceased.

Fig. 8. Deformed mafic enclaves of the tonalite sill. Photograph is looking normal to the XZ principle plane of the strain ellipse as defined by the enclaves; north is to the left.

Numerical Model

The numerical model used in this study assumes that heat transfer is by conduction only and is in one dimension (that is, only in a vertical direction). The one-dimensional equation for conductive heat transfer in a moving medium containing a heat source is

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} - v \frac{\partial T}{\partial z} + \frac{A}{\rho c} \quad (1)$$

where T is the temperature, t is the time, κ is the thermal diffusivity, z is the distance, v is the velocity with respect to the surface (z = 0), A is the volumetric heat production, ρ is the density, and c is the heat capacity of the medium. Values of the physical parameters used in this study are given in Table 1. These values are taken from DeYoreo et al. [1989a, b], are typical of values quoted in the literature [e.g., England and Thompson, 1984], and produce a steady state surface heat flow of 60 mW/m^2 and a temperature of 450°C at 30 km.

Thermal diffusivity (κ) is defined by

$$\kappa = \frac{K}{\rho c} \quad (2)$$

where K is the thermal conductivity. Following the treatment of DeYoreo et al. [1989b] the heat capacity is modified in the region where melt is present in order to account for the latent heat of fusion of the melt. In this particular application, this is done by assuming that crystallization occurs uniformly over the crystallization range; therefore the modified heat capacity term (c_m) is given by

$$c_m = c + \frac{L}{T_l - T_s} \quad (3)$$

where L is the latent heat of fusion, and T_l and T_s are the liquidus and solidus temperatures of the melt, respectively.

Equation (1) is solved numerically using the Crank and Nicolson implicit finite difference method [Smith, 1965], with a distance step of 100 m and a variable time step. Small time steps (starting at 10 years) are used just after the sill is emplaced and are increased geometrically to 10,000 years by 0.5 m.y. This is done in order to minimize errors caused by rapidly changing temperatures in the vicinity of the sill just after emplacement.

Initial conditions, boundary conditions, and geometry. Two models are presented below. Model 1 is used to calculate crystallization times of sill-like bodies emplaced into the lower crust; model 2 simulates the geometry of the Valdez Creek shear zone. The geometries of the models are shown in Figure 9. In model 1, the temperature at the upper boundary (z = 0) is fixed at 0°C; the lower boundary condition is placed at a distance of 100 km from the surface and is defined by a constant heat flux (Q_m). Radioactive heat production (A) is confined to the upper crust and has a constant value down to depth b; below depth b, A = 0. A sill of thickness d at

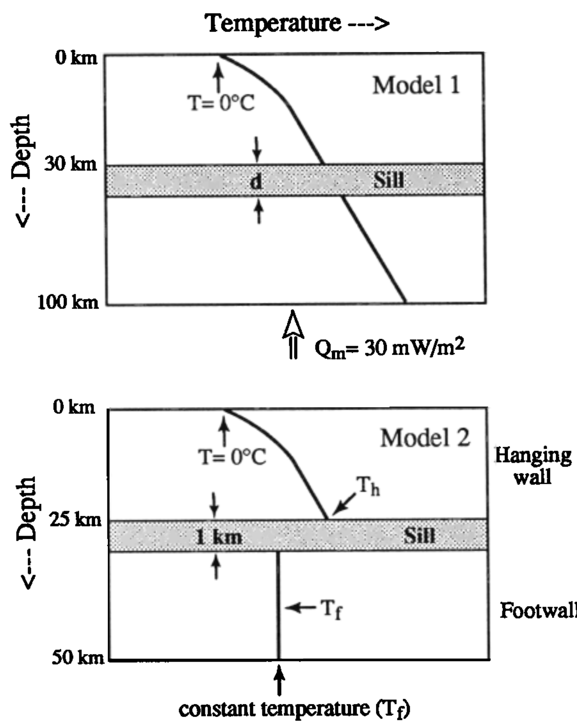


Fig. 9. Geometry of the thermal models. Initial temperature profiles (t = 0) are shown by the heavy solid lines (scale is arbitrary). T_f is temperature of the footwall; T_h is temperature of the hanging wall; d is thickness of sill.

temperature T_m and depth of 30 km is emplaced into the model instantaneously at t = 0. The erosion rate was set equal to zero (v = 0) in all of the examples of model 1.

The thermal structure of the crust at t = 0 is generally different from the steady state thermal structure (t = ∞) in model 1 (Figure 10). This is because the steady state geotherm is perturbed in order to raise the temperature in the lower crust to the desired initial temperature of the country rock (T_c) at the time of emplacement of the sill. This is accomplished by artificially increasing the radiogenic heat production in the upper crust; after the initial conditions are set, heat production is set back to the normal value (Table 1). (Increasing A is arbitrary and is done only to establish the initial temperature profile; alternatively, if the initial temperature profile of the crust is somehow known at the time of emplacement of the sill, then this could be used as the initial condition.) The geological process (or processes) responsible for raising temperatures in the lower crust is not addressed in this study; however, a number of possible processes such as crustal thickening [e.g., England and Thompson, 1984] and the underplating of magma [e.g., Wells, 1980] have been discussed in the literature.

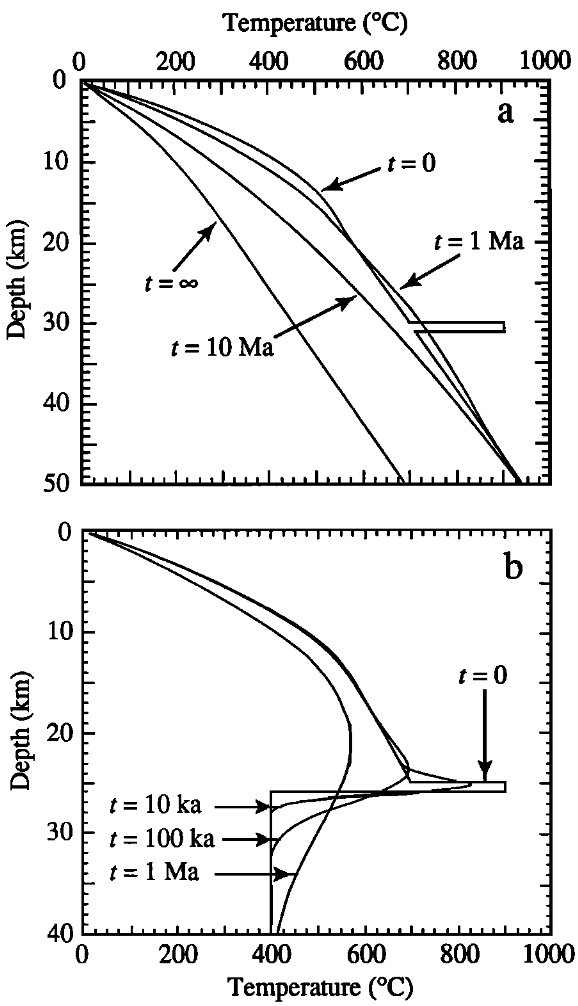


Fig. 10. Temperature versus depth profiles at different times (t). (a) Model 1; top of sill (d = 1 km) at 30 km. (b) Model 2 (v = 0); top of sill (d = 1 km) at 25 km; note that the sill has passed below its solidus temperature (700°C) by 100,000 years.

TABLE 1. Physical Parameters Used in This Study

Parameter	Value
Thermal conductivity (K)	2.5 W/m°K
Thermal diffusivity (κ)	30 km^2/m.y.
Heat capacity (c)	1.0 J/g°K
Radiogenic heat production (A)	2.0 μW/m^3
Depth of heat production (b)	15 km
Mantle heat flux (Q_m)	30 mW/m^2
Latent heat (L)	320 J/g
Liquidus temperature (T_l)	900°C
Solidus temperature (T_s)	700°C

In model 2, the upper boundary condition is the same as in model 1, and the lower boundary condition is fixed at a constant temperature, T_f, the initial temperature of the footwall. T_f is uniform throughout the footwall at t = 0. This geometry is chosen in order to simulate the underplating of "cold" rocks, or rocks which have been transported from the near surface to great depths such as might take place in a subduction zone. The motivation for this is based on tectonic reconstructions for the formation of the Valdez Creek shear zone (C. Davidson, manuscript in preparation, 1992). Similar to model 1, the initial geotherm in model 2 is calculated on the basis of the initial temperature of the hanging wall, T_h, just above the sill. Values chosen for T_h and T_f are discussed below. Erosion rate (v) was varied, and the thickness of the sill (d) was set equal to 1 km in all cases.

Model 2 was constructed in order to simulate the geometry of the Valdez Creek shear zone. Presently, the shear zone and Valdez Creek tonalite dip steeply to the north; if this was the configuration at the time of emplacement of the sill, then a one-dimensional thermal model would not be valid for long time periods (or great distances away from the sill). However, at the time of emplacement (and shortly thereafter), the maximum temperature gradient (and principal direction of heat flow) would be normal to the sill. Therefore assuming one-dimensional heat flow normal to the sill will be valid for short time periods after emplacement (about the length of time for the sill to crystallize). In the application below, this model is used to calculate the length of time for the Valdez Creek tonalite to crystallize and the maximum temperatures which are generated in the footwall immediately adjacent to the sill; maximum temperatures are achieved within the first 3 km of the footwall adjacent to the sill during crystallization.

Crystallization Times of Melt in the Deep Crust

Model 1 is used to calculate the length of time for melt to crystallize in the deep crust under different conditions. In all of the examples $T_l = T_m$ = 900°C, T_s = 700°C, and L = 320 J/g; these values are similar to those used elsewhere for granitoid melts [DeYoreo et al., 1989b; Wells, 1980]. The only variables we will explore here are the thickness of the sill (d) and the temperature of the country rock (T_c) into which the sill is emplaced. Figure 10a is a plot of thermal profiles for different times where d = 1 km and T_c = 700°C. The initial temperature profile is shown (t = 0) along with profiles for t = 1 Ma, t = 10 Ma, and t = ∞; the steady state geotherm (t = ∞) is the same for all of the examples used in model 1. Figure 11 shows crystallization times for sills of 100 m, 1 km, and 2 km in thickness emplaced into rocks with initial temperatures of 500°C, 600°C, and 700°C. It is apparent from Figure 11 that the initial temperature of the country rock has a large effect on crystallization times. A 1 km thick sill emplaced into rocks of 500°C crystallizes in about 45,000 years, whereas a sill of the same thickness emplaced into rocks which

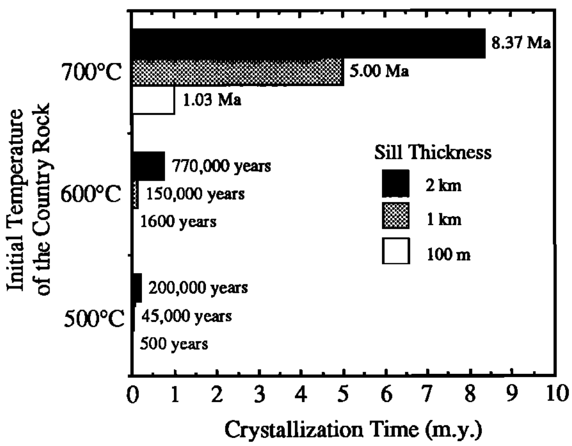


Fig. 11. Crystallization times for sills of different thicknesses intruded into country rocks of different initial temperatures.

were initially at the solidus temperature of the melt (700°C) takes about 5 m.y. to crystallize.

Because melt can remain in a partially molten state for substantial periods of time in the deep crust, it can have a profound impact on the deformational history of a mountain belt. As a simple illustration, consider a region of the deep crust which is being homogeneously deformed by the convergence of two plates. If melt is introduced into this region, strain would be expected to concentrate in the rocks weakened by the presence of melt. If the total convergence of the two plates was accommodated within a single melt-bearing shear zone, then the total amount of displacement across the shear zone will be dependent on the convergence rate of the plates and the length of time that melt is present.

In an attempt to quantify the magnitude of possible "melt-enhanced" displacements during the collision of Wrangellia in southern Alaska, assume the entire convergence in the Late Cretaceous between the North American plate and the Kula plate (to which Wrangellia is assumed to be attached) was accommodated within a single melt-bearing shear zone; convergence rates between North America and the Kula plate in the Late Cretaceous were approximately 10 cm/yr [Engebretson et al., 1985]. In such a situation, melt that takes 1 m.y. to crystallize would allow up to 100 km of displacement across the melt-bearing shear zone before it would become "locked up" by solid rock. Although this is an extreme example, it illustrates that melt-bearing shear zones can account for large displacements in the deep crust.

Application to the Valdez Creek Shear Zone

Crystallization time of the Valdez Creek tonalite. From the results presented above, it is apparent that the length of time for a sill to become fully crystalline is strongly dependent on the temperature of the country rocks into which it intrudes. In the case of the Valdez Creek shear zone (model 2), a 1 km thick sill is emplaced between a hot hanging wall, T_h = 700°C, and a cooler footwall, T_f = 400°C. The initial hanging wall temperature is based on temperature estimates from mineral assemblages found within the hanging wall far from (and assumed to be unaffected by) the Valdez Creek tonalite [Davidson, 1991]. The initial temperature estimate of the footwall is based on temperatures calculated for common biotite producing reactions [Ferry, 1984]; biotite is common in the footwall and found well south of the garnet-in isograd (Figure 2b), suggesting that the footwall was at about biotite grade at the time the hanging wall was emplaced.

A relatively cold footwall would be expected to act as a heat sink for the tonalite sill and hot hanging wall. In this case, the sill will crystallize from the bottom up as long as the initial hanging wall temperature is similar to the solidus temperature of the sill. This is indeed what is observed in model 2, where the last bit of melt to crystallize is at the margin of the sill in contact with the hanging wall. An example of the evolution of geotherms for model 2 is shown in Figure 10b. The same liquidus, solidus, and intrusion temperatures used in model 1 are used in model 2. The crystallization time for the sill in model 2 is about 90,000 years. Again, the crystallization time of the sill is very

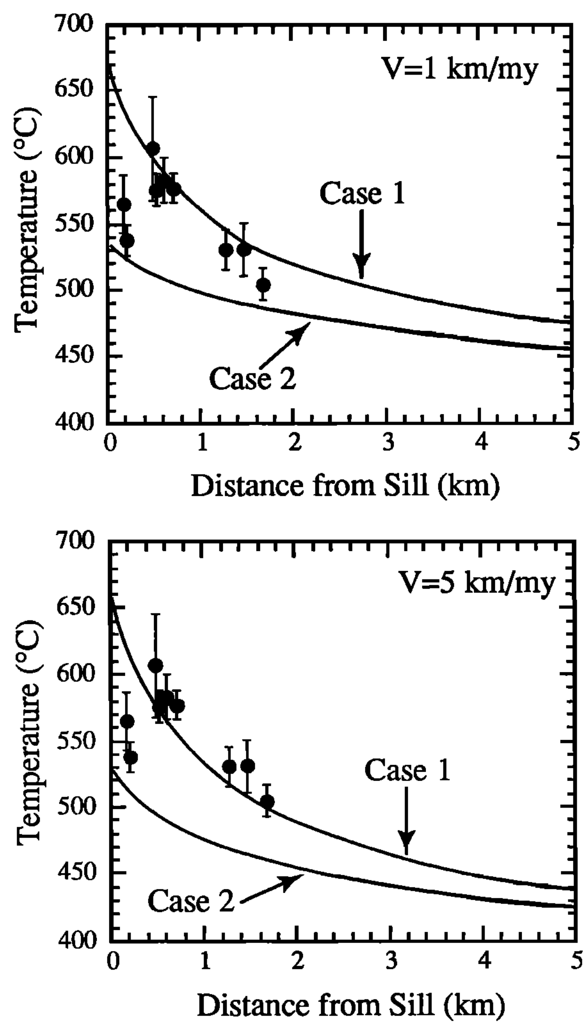


Fig. 12. Model metamorphic field gradients in the footwall for erosion rates of 1 and 5 km/m.y. Solid curves map out the maximum temperatures achieved in the footwall of model 2; lower contact of sill is at 0 km. Case 1: tonalite sill and hanging wall are emplaced above the footwall instantaneously at t = 0; case 2: hanging wall (alone) is emplaced above the footwall at t = 0. Garnet-biotite temperatures [Ferry and Spear, 1978] calculated from rocks across different structural levels from within the footwall are shown as solid circles; error bars are 1σ (the reason for the two anomalously low garnet-biotite temperatures recorded immediately adjacent to the sill is not known).

sensitive to the initial temperatures of the hanging wall and footwall. Increasing the initial temperature of the footwall to 450°C increases the crystallization time to 120,000 years; reducing the hanging wall temperature to 650°C reduces the crystallization time to 70,000 years.

Valdez Creek tonalite as a heat source. The presence of a steep inverted metamorphic gradient within the footwall suggests that the heat for the metamorphism was supplied from above. Two potential heat sources exist within the hanging wall; the first is the hanging wall itself, and the second is the Valdez Creek tonalite. Model 2 is applied to the Valdez Creek shear zone in order to test if heat supplied by the hanging wall alone is sufficient to produce the observed inverted metamorphic gradient or if an additional heat source (such as the Valdez Creek tonalite) is required. As a quantitative measure of the metamorphic field gradient in the footwall, the garnet-biotite geothermometer of Ferry and Spear [1978] was used to calculate temperatures for nine samples from across different structural levels in the footwall. These temperatures are then compared to "model metamorphic gradients" (maximum temperature arrays) generated in the footwall of the model. In Figure 12, model metamorphic gradients for erosion rates of 1 and 5 km/m.y. are calculated with the tonalite sill (case 1) and without the sill (case 2).

As is apparent from Figure 12, the presence of the sill greatly increases the magnitude of the model metamorphic gradient in the footwall for a given erosion rate. For either erosion rate, the maximum temperatures achieved in case 2 do not fit the garnet-biotite temperatures. A better fit is achieved in case 1, where model metamorphic gradients generated by erosion rates of 1 and 5 km/m.y. bracket the garnet-biotite temperatures. These results support the conclusion that the observed inverted metamorphic field gradient within the footwall is due to the emplacement of the hanging wall and simultaneous intrusion of the Valdez Creek tonalite.

DISCUSSION AND CONCLUSIONS

The first deformation event (D1) within the Maclaren Glacier metamorphic belt has been correlated to the initial stages of collision between Wrangellia and North America (C. Davidson, manuscript in preparation, 1992). With continued convergence, the Valdez Creek shear zone was formed by the emplacement of the hanging wall over the footwall; this deformation event is designated D2.

Our observations suggest that melt was present during D1 and D2. Evidence includes the presence of thin granitoid sills parallel to the s1 fabric which have been deformed and boudinaged during the D1 event; these sills have in turn been cut by younger granitoid sills which are parallel to s2. The macroscopic fabric within the Valdez Creek tonalite is parallel to that of the shear zone, with high apparent finite strains recorded by mafic enclaves. Mineral lineations and long axes of the enclaves within the sill are parallel to stretching lineations in the shear zone; this suggests that they formed in the same stress field. The amount of strain recorded by the mafic enclaves appears to be greater than that recorded by the microscopic fabric of the tonalite host. However, if the sill was being deformed while it was still partially molten, then the apparent strain deficit as indicated by the microscopic fabric can be rationalized. Very little intragrain deformation would be expected if much of the strain was taken up by grain boundary sliding in a crystal mush or by melt enhanced diffusion creep [Dell'Angelo and Tullis, 1988]. The mush would have to be sufficiently crystalline to be able to transmit a shear stress to the mafic enclaves which could then be deformed into their present geometry. Therefore the actual amount of strain could be even greater than that recorded by the enclaves. Other evidence from the microscopic features of the Valdez Creek tonalite also implies that the sill was being deformed while melt was still present. We conclude from our observations that melt was repeatedly intruded into the Valdez Creek shear zone while it was in an active compressional regime.

In order for melt to play a significant role in a deforming orogen, the length of time for the melt to crystallize has to be long (>1 m.y.). Two important parameters controlling the length of time for crystallization are the initial temperature of the country rock and the geometry of the igneous body. In this communication, we have addressed only sill-shaped geometries of varying thickness (Figures 9 and 11). Of these two parameters, the initial temperature of the country rock is the most important. Sills of 100 m to 2 km in thickness intruding into rocks with initial temperatures equal to the solidus temperature of the melt (700°C) have crystallization times of over 1 m.y.; the same sills intruded into initial country rock temperatures of 500°C crystallize in 200,000 years or less (Figure 11). These results suggest that melt may be present for long periods of time in the deep crust when temperatures of the surrounding rocks approach the solidus temperatures of granitoid melts. The processes responsible for raising temperatures in the lower crust are not addressed in this communication; however, England and Thompson [1986] and DeYoreo et al. [1989b] show that temperatures in the lower crust can easily reach the solidus temperature of common granitoid melts during crustal thickening.

In the Valdez Creek shear zone, macroscopic and microscopic evidence suggests that the Valdez Creek tonalite was emplaced during deformation; this is further supported by modeling the metamorphic field gradient observed in the footwall. Model metamorphic gradients generated by the simultaneous emplacement of the Valdez Creek tonalite and hanging wall over the footwall fit the observed geometry of the "actual" metamorphic gradient. Because the growth of metamorphic minerals in the footwall was coincident with deformation within the shear zone, the model results support the conclusion that the Valdez Creek tonalite was emplaced into the shear zone while the shear zone was active. The Valdez Creek tonalite is estimated to have crystallized in about 90,000 years. However, crystallization times are a strong function of sill thickness; 90,000 years is probably a minimum value since the original thickness of the sill at the time of emplacement is not known. Following the example presented earlier, and assuming that the bulk of convergence between the North American plate and Wrangellia were concentrated within the Valdez Creek tonalite, we estimate the minimum amount of displacement that could have been accommodated across the sill while melt was present is about 10 km. The uncertainty in this value is difficult to quantify, and it will not be attempted here; however, factors contributing to error include the uncertainty in the convergence rate, uncertainties in the original thickness of the sill, and the uncertainty in the temperature profile at the time of the emplacement of the sill.

Acknowledgments. This work was supported by National Science Foundation grant EAR87-19368, U.S. Geological Survey TACT project, National Geographic Society, Geological Society of America, Sigma Xi, and the Kraus Memorial fund of Princeton University. Preliminary heat flow calculations done with James Connely at Eidgenössische Technische Hochschule, Zürich, in 1988 first demonstrated to us that melt could reside long enough in the deep crust to strongly influence deformation in orogenic belts. We thank

Jim DeYoreo for his help in the thermal modeling. Terry Pavlis (University of New Orleans) is acknowledged for first recognizing the significance of the strained xenoliths in the Valdez Creek tonalite. We also thank Joe Kurtak and the U.S. Bureau of Mines for providing valuable helicopter time. The thoughtful reviews of William McClelland and Jane Selverstone are gratefully appreciated. Portions of this material are based on work supported under a National Science Foundation graduate fellowship to Davidson.

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(Received April 30, 1991;
revised September 30, 1991;
accepted November 13, 1991.)