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Emplacement-related microstructures in the margin of a deformed pluton: the San José tonalite, Baja California, México

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Abstract

In the San José tonalite pluton, Baja California, México, interior magmatic fabrics give way progressively to subparallel, solid-state fabrics with S–C structures near the margins. The outer rim ('marginal northern unit'), approximately 1 km thick, is inferred to have been deformed during the emplacement of the inner parts of the magma chamber. The rocks show a transition from strongly deformed at the outer edge to progressively less deformed towards the inner edge of the marginal northern unit. A small amount of melt appears to have been present during the deformation, as indicated by (1) apparent contact melting between some grains of plagioclase, and (2) coarse-grained aggregates and single-grains of quartz, K-feldspar and/or more sodic plagioclase in fracture-controlled openings in primary plagioclase crystals. The more sodic plagioclase typically grew as rims on the walls of the openings in twin continuity with the primary plagioclase. The deformed rocks have a prominent foliation marked mainly by aggregates of biotite, with contributions by hornblende and titanite grains, as well as apparently recrystallized aggregates of quartz and/or feldspar, at least some of which may have originated as zones of fragmentation. Many primary grains of plagioclase, hornblende, biotite and titanite show evidence of brittle deformation and fragmentation. The shapes of aggregates in the folia, relationships between incipient folia and residual primary grains, and occurrences of many folia in narrow zones transecting strong minerals (plagioclase and hornblende), suggest that brittle processes may have controlled at least some of the foliation development. However, evidence of strain-induced grain-boundary migration and bulge nucleation of new grains of quartz and plagioclase indicates that dislocation creep also played a part in the deformation. Recrystallization of plagioclase, together with intrusion of the pluton at greenschist facies ambient temperatures, confirms high-temperature deformation, with heat derived from the cooling magma.

Keywords: Brittle deformation; Foliation development; Intrusion-related deformation; Pluton; Recrystallization; Submagmatic flow; Subsolidus flow

1. Introduction

This paper describes grain-scale processes that we suggest occurred during the relatively high-temperature deformation of the marginal zone of a tonalite pluton (the San José tonalite, Baja California, México) during its intrusion. The multiple-pulse pluton appears to have expanded outward during its intrusion history (Johnson et al., 2003), forming an asymmetrical nested pluton (Gastil et al., 1991). Our aim is to investigate the deformation mechanisms in the outermost rim of the pluton that we infer have facilitated this expansion. We suggest that small amounts of residual melt were present during the deformation, which may have been relatively rapid and partly brittle. However, we emphasize that many of our interpretations are tentative, and are presented to stimulate further research.

The origin of margin-parallel, solid-state foliations within the marginal zones of granitoid plutons has been a topic of considerable debate. Some workers have suggested that pluton margins begin to crystallize during emplacement, and that continued addition of magma from below causes the pluton to expand (e.g. Sylvester et al., 1978; Holder, 1979; Bateman, 1985; Courrioux, 1987; Ramsay, 1989). This expansion is inferred to cause stretching and ductile deformation of the solid pluton margin, as well as the adjacent wall rocks. The above-cited examples call on

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diapiric ascent, but similar processes and problems would be involved if expansion occurred during dike-fed pluton growth (Clemens and Mawer, 1992; Petford, 1996; Clemens et al., 1997).

Several workers have suggested that the amount of solid-state deformation of a solid rind of crystallized plutonic rock that can be attributed to magma expansion is limited (Paterson, 1988; Paterson et al., 1991; Vernon and Paterson, 1993; Paterson and Vernon, 1995; Vernon, 2000), and the very thin marginal northern unit of the San José pluton is consistent with this conclusion. The process would be less difficult if the largely crystallized rind of the pluton contained a small amount of melt when the later pulse of magma was intruded, enabling 'submagmatic flow' (Paterson et al., 1989). This is because of the marked weakening effect, determined experimentally, of even small amounts of melt (as little as 5–10%), owing to extensive penetration of melt along grain boundaries (van der Molen and Paterson, 1979; Rosenberg and Handy, 2000). Melt-assisted deformation may occur, involving hydraulic fracturing (Carney et al., 1991, p. 468), melt-enhanced embrittlement (Davidson et al., 1994; Connolly et al., 1997; Holyoke and Rushmer, 2002), melt-assisted grain-boundary sliding, grain-boundary migration assisted by contact-melting (Park and Means, 1996), strain partitioning into melt-rich zones (Vigneresse and Tikoff, 1999), and transfer of melt to sites of low mean stress (Paterson et al., 1998). Unfortunately, evidence of some of these processes may not be evident in the resulting microstructure (Park and Means, 1996).

An important aim of this paper is to distinguish melt-present deformation microstructures from those formed at subsolidus conditions, in order to determine whether solid-state deformation occurred in the presence of a small amount of melt. We describe the development of emplacement-related microstructures by thoroughly examining the microstructural evolution of the rocks across the preserved gradient in solid-state deformation in the pluton's margin. We use our observations to address: (1) evidence of brittle deformation, (2) evidence of melt-present deformation, (3) evidence for recrystallization involving modification of fragments produced by brittle deformation, as opposed to recrystallization induced by ductile strain, (4) evidence of contact melting during submagmatic deformation (Park and Means, 1996), and (5) whether the spatial deformation gradient in the pluton margin reflects the deformation history experienced by the most deformed rocks.

In a companion paper (Johnson et al., 2004), we combine numerical modeling with microstructural observations from the same rocks to evaluate: (1) the role of mica in initiating and localizing microshear zones, (2) the progressive mechanical evolution from initially brittle to ductile deformation leading to pervasive foliation, and (3) the range of strain rates at which the emplacement-related deformation may have occurred.

2. The San José pluton

The Cretaceous San José pluton (Figs. 1 and 2) is part of the Peninsular Ranges batholith of northern Baja California, México. It was mapped in detail by Murray (1978, 1979), and remapped as part of a regional mapping program by Johnson et al. (1999, 2003).

More than 95% of the pluton is composed of coarse-grained (>3.0 mm) hornblende–biotite tonalite with abundant micrograniotoid enclaves (~1%) and rare wall-rock xenoliths. The remaining 4–5% is composed of largely synplutonic, hornblende-bearing sheets, dikes and elliptical intrusions of intermediate to mafic composition, some of which are localized along internal contacts within the pluton (Fig. 1). The pluton was emplaced at a depth of approximately 7–12 km (S.E. Johnson, unpublished aluminum-in-hornblende data).

On the basis of general outcrop appearance, Murray (1978) divided the pluton into three tonalite units with gradational contacts (Figs. 1 and 2), namely the northern, central and southern units. In the outer part of the northern unit he recognized a gneissose border-phase tonalite (referred to herein as the *marginal northern unit*), characterized by solid-state deformation fabrics that overprint pre-existing, subparallel magmatic fabrics.

The history and mechanism of intrusion of the San José pluton have been discussed by Johnson et al. (2003). The following lines of evidence suggest that the deformation of the marginal northern unit may reflect pluton emplacement, rather than regional deformation: (1) refolding of isoclinal regional folds about axial surfaces parallel to the pluton margin in the northwestern triple point (Fig. 1); (2) contact-parallel foliation in wall rocks of the northwestern triple point that overprints all regional deformation fabrics; (3) lack of evidence at any scale for deformation fabrics overprinting the contact-parallel foliation in the pluton or wall rocks where they lie at low angles to the NE–SW regional shortening direction; (4) sharp truncation of tight regional folds, bedding and regional foliation along the southwestern margin of the pluton (Fig. 1); (5) displacement vectors for S–C surfaces approximately parallel to the solid-state lineation in the pluton and wall rocks adjacent to the pluton margin, their attitudes being consistent with northward pluton expansion; and (6) intrusion of the pluton into relatively cool (greenschist facies) country rocks, so that heat required for high-temperature effects such as plagioclase recrystallization is likely to be magmatic, as opposed to regional, in origin. The absence of evidence for post-pluton, penetrative, regional deformation implies that the marginal deformation accompanied the emplacement of the inner parts of the magma chamber (Johnson et al., 2003), though the possibility that regional stresses (without any evidence of associated structures) might lead to deformation of hot wall-rocks and pluton margins cannot be eliminated. Brittle/ductile displacement on the nearby Rosarito fault

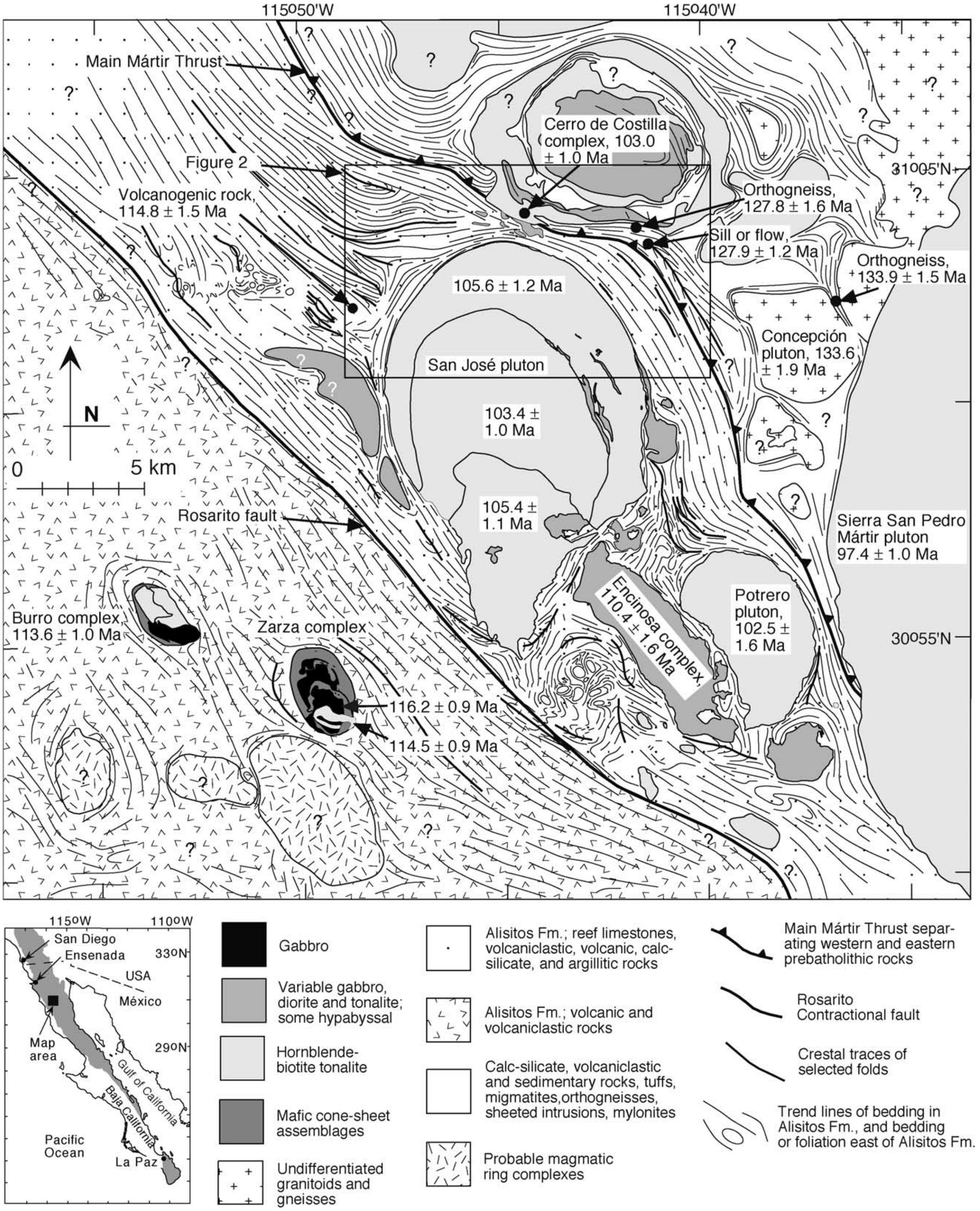

Fig. 1. Geologic map of the San Jose pluton and surrounding region. Map constructed from detailed ground mapping and air photo interpretation. Question marks indicate regions that have not been ground checked. The San José pluton is located in the map center. Ages of the northern, central and southern units of the pluton shown. All ages are from SHRIMP U–Pb zircon data reported in Johnson et al. (1999, 2003). See legend for geologic information. Inset map at lower left shows the Baja California peninsula; the Peninsular Ranges batholith and adjacent pre-batholithic country rocks are shaded.

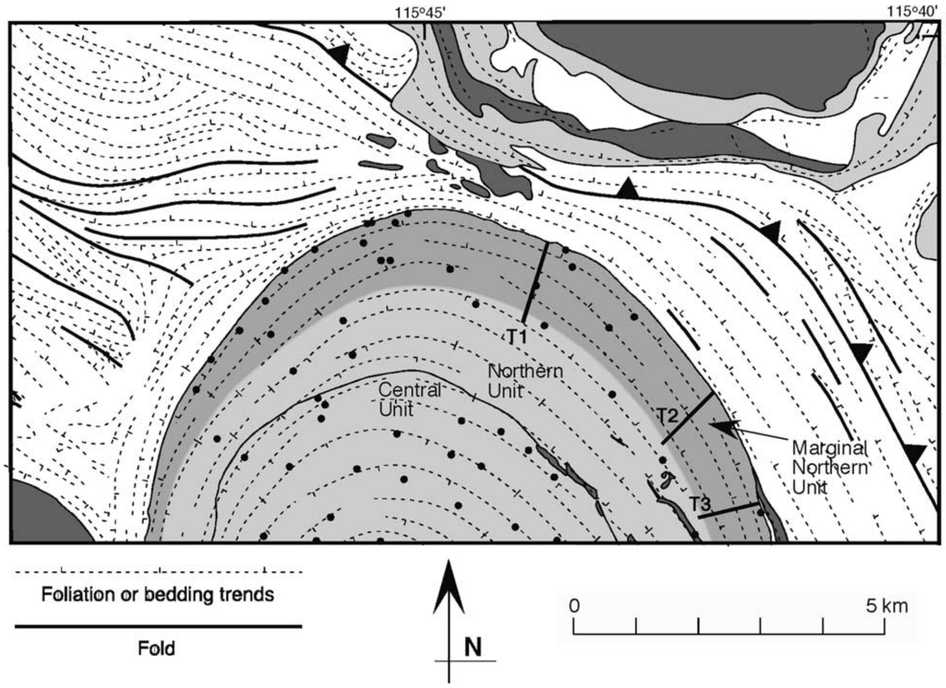

Fig. 2. Map showing bedding and foliation trends within and surrounding the northern third of the San José pluton, using our data and those of Murray (1978). See Fig. 1 for geology legend. Marginal northern unit shown in darker shading within the northern unit. Contact between central and northern units shown. Ages for central and northern units are shown in Fig. 1. T1–T3 are transect lines along which microstructures of the northern marginal unit were investigated. Additional samples available in this study are shown at filled dots; see Johnson et al. (2003) for complete sample distribution in the pluton.

(Schmidt and Paterson, 2002) may be responsible for the late fracturing and alteration in the pluton.

Though all three pulses in the pluton are of the same age, within error, the first pulse apparently had time to extensively crystallize before the next pulse arrived (Johnson et al., 2003). If so, the marginal rocks may have been largely crystalline, with small amount of residual melt, so that the marginal northern unit could provide an excellent opportunity to trace microstructural changes occurring during deformation across the transition from submagmatic to subsolidus conditions.

3. Deformation in the marginal northern unit

A foliation is present throughout most of the pluton (Fig. 2), and is largely magmatic in origin, based on the criteria of Paterson et al. (1989) and Vernon (2000). In the marginal northern unit, this foliation has been overprinted by a subparallel solid-state foliation. This solid-state foliation is best developed in the northwestern, northern and eastern parts of the marginal northern unit, and its intensity diminishes from the wall-rock contact towards the pluton interior. S–C structures or shear bands occur within at least 200 m of the wall-rock contact, where C surfaces form discrete lenticular layers up to a few millimeters thick and several centimeters long.

Most or all of the solid-state deformation in the marginal northern unit occurred in the primary magmatic minerals, namely plagioclase, quartz, biotite and hornblende, with minor K-feldspar and titanite. Secondary epidote, chlorite, white mica and titanite are widespread in small amounts, suggesting deuteric hydrothermal activity. Generally these minerals occur randomly inside the primary grains, although some also occur in fractures and folia. No clear evidence links them to the main deformation, but they may have been involved in the later stages, for example, where epidote and white mica occur in biotite folia.

The tonalite in the marginal northern unit typically has a strong preferred alignment of primary plagioclase grains, delineating a magmatic flow fabric parallel to the margin of the pluton (Johnson et al., 2003). Microgranitoid enclaves are also elongate parallel to the magmatic foliation, with a strong alignment of their plagioclase crystals, reflecting magmatic strain of the enclave magma (e.g. Vernon et al., 1988; Vernon and Paterson, 1993; Paterson et al., 2004). The other deformation features of the microgranitoid enclaves are identical to those described below for the tonalite host rocks, except that evidence for melt-present deformation is rare to absent. This may be due to slightly higher solidi for the enclaves, implying that they were solid by the time submagmatic flow occurred in the host tonalite, but could also be due to the small number of enclave samples investigated.

Oriented samples were collected along three transects across the marginal northern unit (Fig. 2), and more than 100 additional samples in the pluton were also examined

(Fig. 2). Deformation microstructures for each mineral of the tonalite are described in the following section.

4. Deformation microstructures and foliation development

4.1. Biotite deformation

In deformed tonalite without a solid-state foliation, biotite shows evidence of highly variable amounts of solid-state deformation, and tends to be more deformed than the other minerals. Some grains appear to be undeformed, whereas others are slightly bent and/or locally kinked. However, kinking in biotite is extremely rare in all the deformed rocks, suggesting that slip on cleavages was able to accommodate most of the strain. Evidently the strain of biotite was not inhibited sufficiently to necessitate the development of kink bands as alternative slip systems. This apparent ease of deformation could have been assisted by the inferred presence of melt along interfaces between biotite and other minerals, as discussed later.

In more deformed tonalite, biotite shows evidence of internal deformation and/or partial to complete conversion to aggregates. Some of the aggregates retain relatively equant shapes, whereas others merge into folia (Fig. 3), indicating that the folia developed from formerly large primary grains.

Some biotite grains show evidence of limited slip along cleavage planes, resulting in the curved, 'overlapping-hinged' microstructure shown in Fig. 3, which is similar to microstructures described by Vernon (1977). Some biotite grains show ragged internal fractures, and have been converted to mosaics of slightly misoriented segments, especially at the edges (Fig. 4). Ragged edges suggest slip on the cleavage, and resemble microstructures produced in experiments designed to maximize slip on the basal plane (Kronenberg et al., 1990).

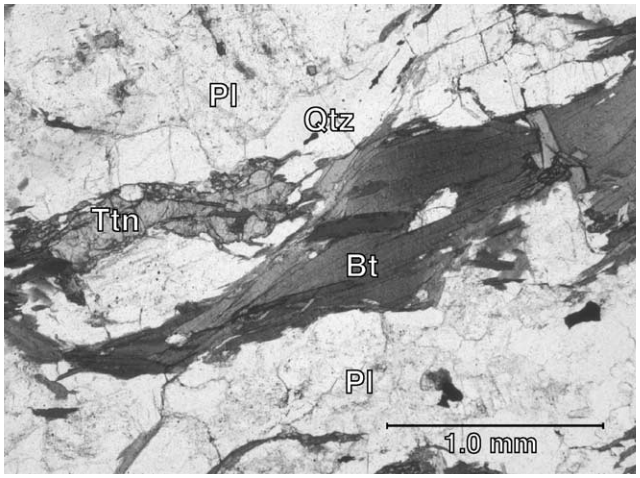

Fig. 3. Biotite grain that has been thinned down at its end by cleavage slip, forming 'overlapping-hinged' microstructure, leading into a biotite folium. Plane-polarized light.

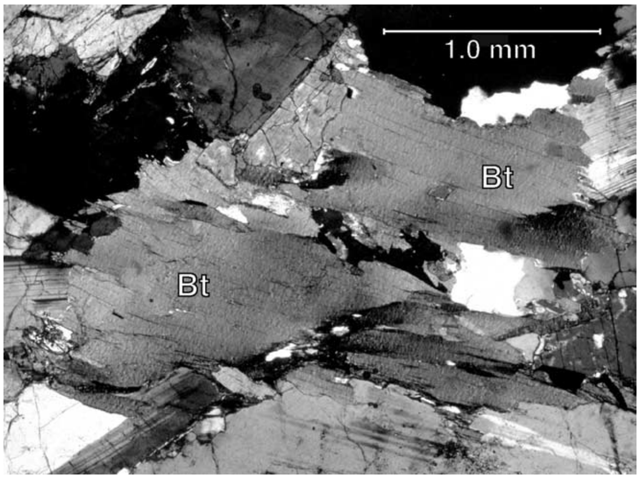

Fig. 4. Fracturing and incipient fragmentation of primary biotite grains, producing misoriented segments, especially at the edges, and ragged internal boundaries. Ragged edges against plagioclase may be due to slip into opening fractures in the plagioclase. Crossed polars.

Markedly misoriented new grains of biotite have crystallized in some fractured segments, and short internal fractures and misoriented shreds may be produced. Small new grains occur in some examples (Fig. 5), and these may have grown from misoriented shreds and short kink-like features, in the manner suggested for the formation of new grains by Bell (1978, 1979).

Locally, biotite grains have been bent around and crushed against strong plagioclase grains, resulting in aggregates of slightly misoriented cleavage fragments, together with growth of a few new grains that are more misoriented (Fig. 6). Some strongly deformed biotite aggregates appear to have been 'shredded' into aggregates of cleavage fragments (Fig. 7).

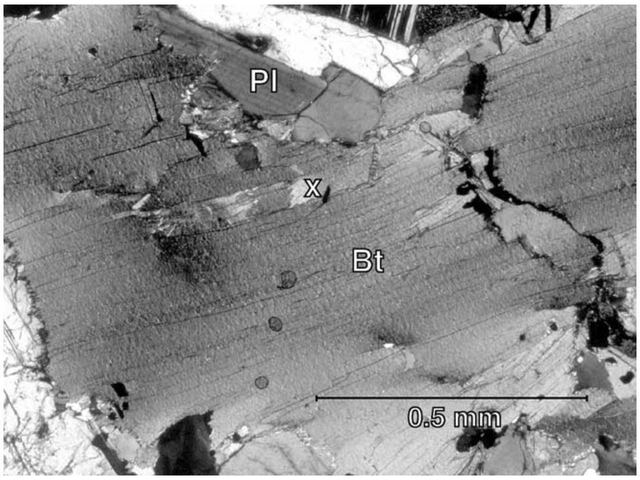

Fig. 5. Short internal fractures and misoriented shreds in a primary biotite grain. Small new grains may have grown from misoriented shreds (X) and short kink-like features in the manner suggested for the formation of new grains by Bell (1978, 1979). Crossed polars.

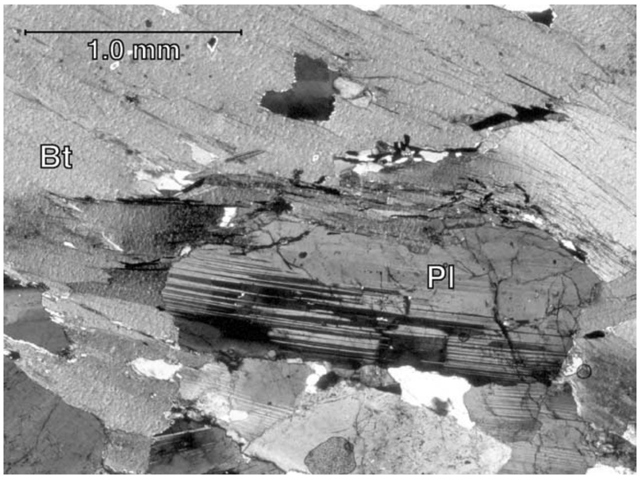

Fig. 6. Biotite grain distorted and heavily damaged against a plagioclase crystal, showing fracturing and slightly misoriented segments. Crossed polars.

Incipient stages in the development of biotite-rich folia are indicated by small, en-échelon grains that appear to be cleavage slices of biotite squeezed into a developing foliation (Fig. 8). Some biotite flakes appear to have been opened along the cleavage to form 'splays' projecting into adjacent grains or grain boundaries of quartz or plagioclase. In addition, biotite folia appear to have been produced by slip of cleavage fragments (Figs. 3 and 9). Biotite aggregates occurring along intergranular and intragranular fractures in the plagioclase, hornblende and quartz may have been emplaced mechanically or deposited from solution, though we have no clear evidence for either process.

Biotite folia are progressively better-developed in the more strongly deformed tonalite, and tend to drape and anastomose around the plagioclase (Fig. 10) and hornblende grains. Several examples of biotite folia crossing hornblende grains have been observed, some involving boudinage of the hornblende. Some biotite-rich folia drape around plagioclase crystals that locally appear to have had their corners

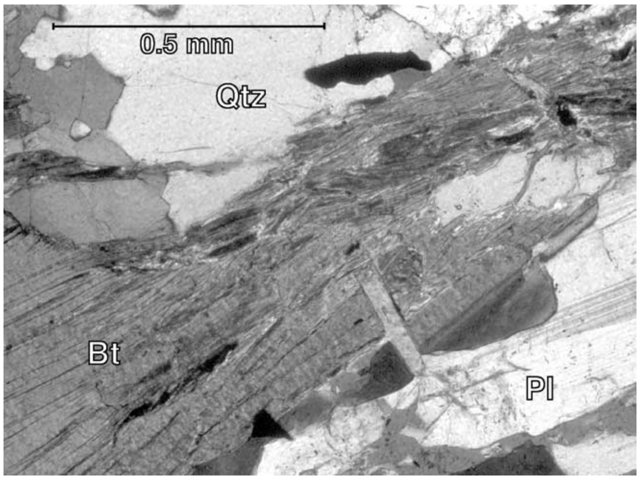

Fig. 7. Strongly deformed ('shredded') biotite aggregate that appears to consist of cleavage fragments. Plane-polarized light.

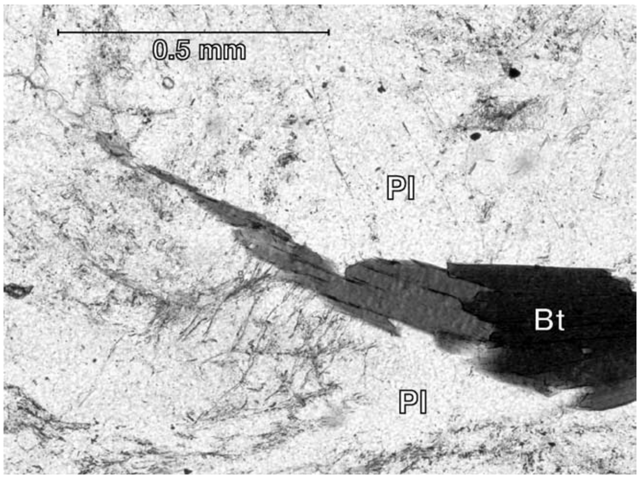

Fig. 8. Folium of en échelon biotite grains in an inferred former fracture between plagioclase grains. We infer that the biotite grains were initially part of a single grain, and that the folium was formed by slip of biotite cleavage fragments. Plane-polarized light.

removed, as indicated by truncated zoning (Fig. 11), suggesting possible operation of the mechanism observed in experimentally deformed octachloropropane by Bons and Jessell (1999). In the more strongly deformed rocks, biotite folia link together along plagioclase and hornblende grain boundaries, across grains of plagioclase or hornblende, or through recrystallized quartz aggregates (Fig. 10).

Though many of the biotite folia give the impression of cleavage fragments that may have slipped and been oriented mechanically (Figs. 3 and 8), the individual grains generally are well-formed and show no evidence of internal deformation. Moreover, many of the foliation-forming aggregates contain well-formed crystals oblique to the foliation, and deformed parts of the original primary biotite grains commonly are encased or transgressed by aggregates of new grains (Fig. 3). Some biotite grains and folia have been converted to decussate aggregates (Fig. 12), even in

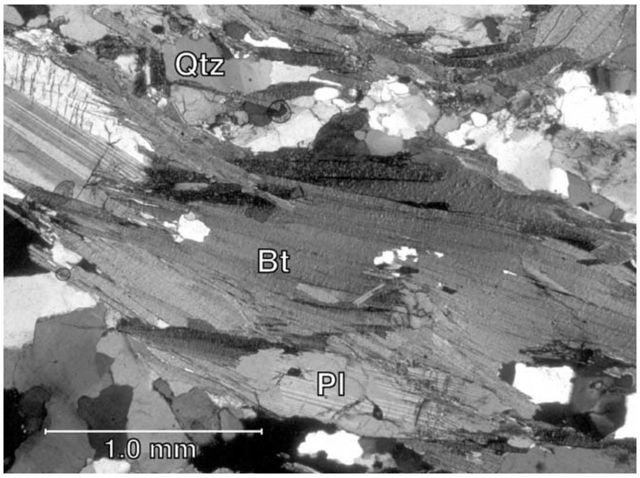

Fig. 9. Strongly deformed biotite grain with misoriented cleavage segments and frayed ends, passing into a biotite folium (top-left). Plane-polarized light.

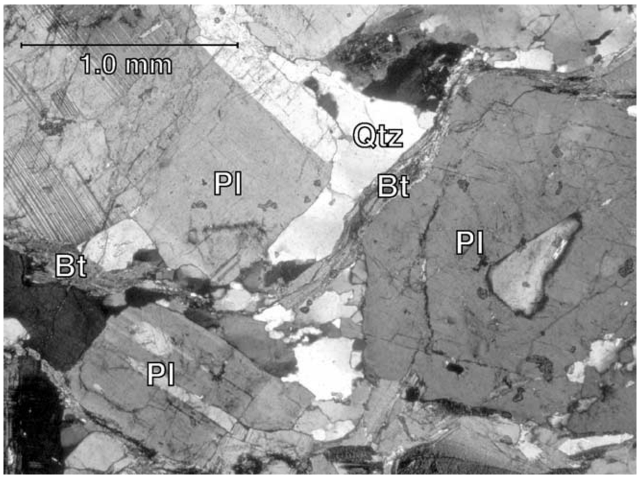

Fig. 10. Biotite folium draped around plagioclase crystals, suggesting that relatively weak biotite-rich folia are deflected around large grains of strong minerals. Crossed polars.

weakly deformed tonalite. Decussate aggregates are generally inferred to be formed by classical recrystallization or neocrystallization, in response to dislocation-controlled strain accumulation (e.g. Etheridge and Hobbs, 1974; Vernon, 1977; Bell, 1978, 1979). However, in view of the evidence for brittle deformation of biotite, the possibility should be entertained that the grains in the decussate aggregates may have been formed by growth of fragments, in order to minimize interfacial free energy by the development of low-energy (001) faces (Kretz, 1966; Vernon, 1968). Though deformation-induced, this recrystallization would be driven by high interfacial free energies caused by fragmentation, rather than by crystal–plastic strain. This is supported by the observation that, though evidence of deformation is commonly preserved in the primary relics (showing that it was not removed by the recrystallization), it is absent from the new grains.

Many folia in these rocks consist entirely of biotite.

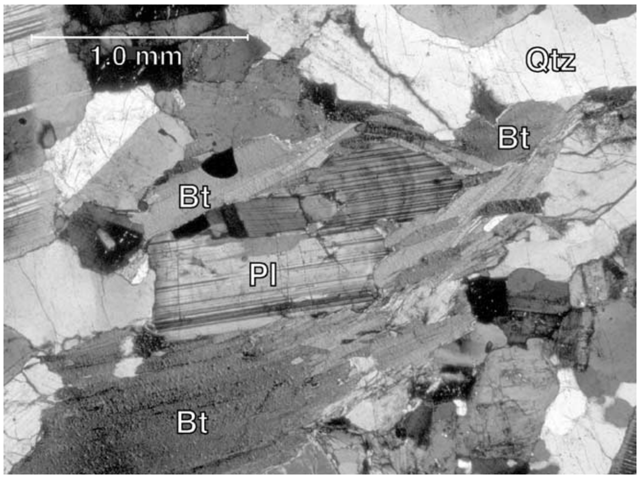

Fig. 11. Truncation of plagioclase zoning by several biotite folia, suggesting removal of part of the plagioclase grain, possibly by fracturing accompanied and/or followed by development of the folia. Crossed polars.

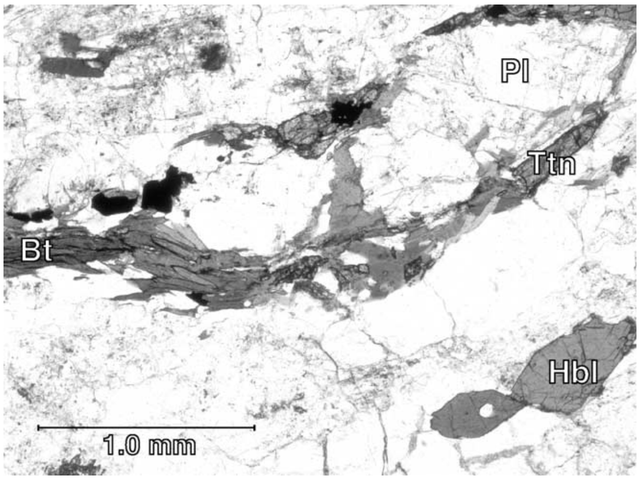

Fig. 12. Irregular and angular grains, resembling fragments, of titanite distributed along a biotite folium. Plane-polarized light.

However, many others consist of biotite with other minerals, especially quartz and plagioclase, less commonly white mica, opaque minerals, hornblende, epidote or titanite (Fig. 12). Some of the hornblende, epidote and titanite grains appear to be fragments dislodged from larger grains and mechanically incorporated into the biotite folia (Fig. 12; e.g. Mancktelow, 1985). Other biotite-rich folia, especially those involving quartz, plagioclase and white mica, may have involved chemical reaction and consequent neocrystallization. This applies also to rocks in which some of the biotite grains have different colors (greenish brown versus reddish brown).

4.2. Hornblende deformation

In the less foliated tonalite, hornblende shows little or no optical evidence of deformation, having sharp or very slightly undulose extinction. Locally it has been partly replaced by epidote and/or chlorite, owing to subsolidus hydration. With increasing deformation, some hornblende shows marginal replacement by biotite. In slightly more deformed rocks, the hornblende commonly shows a weak internal substructure, with small misorientations, though a few grains show much larger misorientations across cleavages that probably acted as fracture surfaces (Fig. 13).

Some hornblende grains show major fractures along cleavage planes, forming jagged offsets, along which biotite (Fig. 13) or quartz have crystallized locally, suggesting crystallization from solutions (possibly melt) penetrating along the fractures. Biotite may be dispersed along both sets of hornblende cleavage planes. In some fractured grains, cleavage rhombs of hornblende have been markedly rotated out of registry with the adjacent part of the grain.

In the more foliated tonalite, some hornblende crystals have been strongly stretched, and consequently have undergone boudinage, fragmentation into misoriented segments (Fig. 14), and fragmentation into what appear to be elongate splinters, especially at their ends (Fig. 15). Some

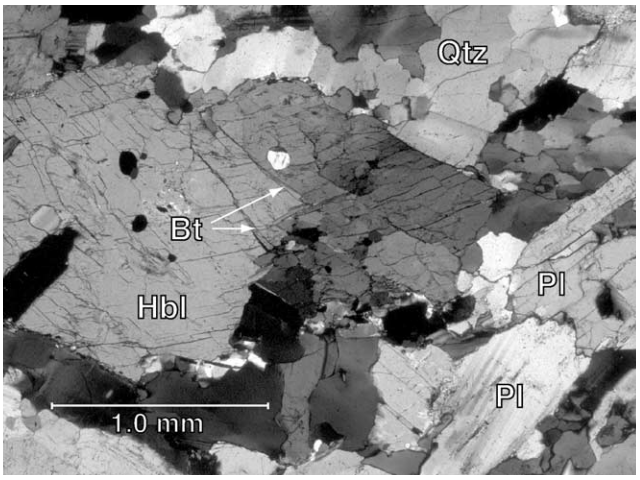

Fig. 13. Subgrain microstructure in hornblende related to cleavage-controlled fracturing. Biotite occupies some of the fracture-controlled openings. Plane-polarized light.

hornblende grains also thin down at their ends (Fig. 16) and break into apparently cleavage-controlled fragments that become part of a developing foliation (Figs. 15–17). Some of these inferred fragments have lozenge-shapes, with 'tails' that disperse into minute grains. Rarely hornblende forms short folia of recrystallized aggregates (Fig. 18). The hornblende also occurs uncommonly with quartz and/or plagioclase in recrystallized aggregates in folia. Hornblende folia tend to be more continuous in the most deformed rocks (Fig. 18).

The microstructural evidence is consistent with predominantly brittle deformation processes for hornblende, as observed in other naturally and experimentally deformed rocks. For example, Allison and La Tour (1977) described fracturing, imbrication, boudinage and detachment of fragments of hornblende in an amphibolite facies mylonitic rock. Nyman et al. (1992) observed {110} cleavage-controlled subgrains in amphibole porphyroclasts from a deformed amphibolite, and inferred cataclastic deformation.

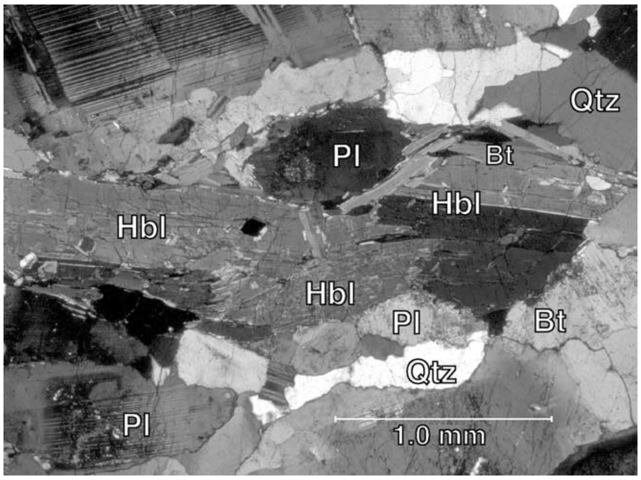

Fig. 14. Hornblende crystal that has been fractured and broken into three sections, the central one of which has been rotated. Crossed polars.

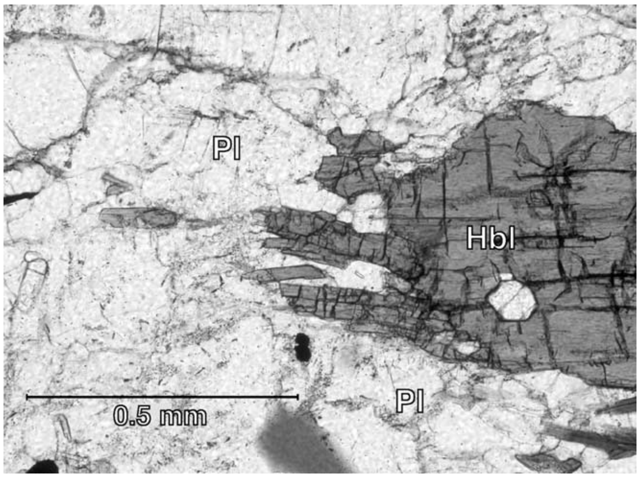

Fig. 15. Hornblende grain that appears to have been fractured at its end, forming what we suggest are splintery fragments that may have slid along former fractures in the adjacent plagioclase. Plane-polarized light.

Ross and Wilks (1996) observed that in experimentally sheared orthopyroxene granulite, amphibole grains changed their shapes by fragmentation of original grains along microfaults, both across and parallel to cleavage planes, with the result that foliation development involved mechanical rotation of amphibole fragments, without optical or TEM evidence of recrystallization. Nyman et al. (1992) observed that many amphibole fragments have an 'undulose mosaic' extinction pattern, which probably was formed by incipient cataclasis, as well as microfractures that offset grain boundaries, owing to slip along {110} cleavages, as we have observed for the hornblende in the deformed rocks of the San José pluton. Evidence of ductile deformation in amphibole was found in fault zones produced experimentally at subsolidus conditions at 650–950 °C by Hacker and Christie (1990), but cataclastic deformation occurred in their melt-present experiments.

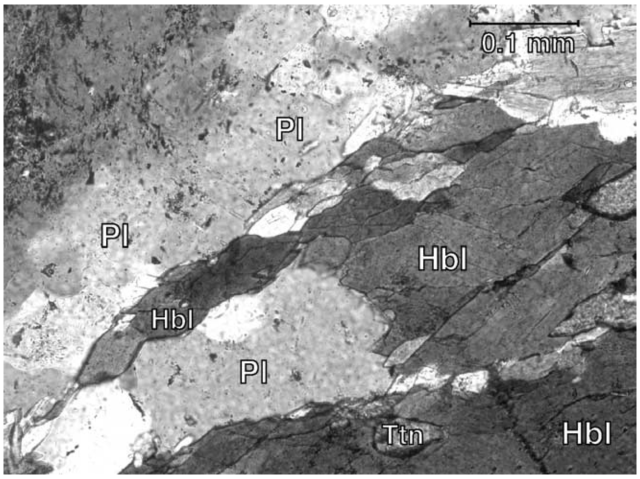

Fig. 16. Apparent cleavage-controlled fragments of hornblende in a folium leading away from a deformed grain of primary hornblende. Plane-polarized light.

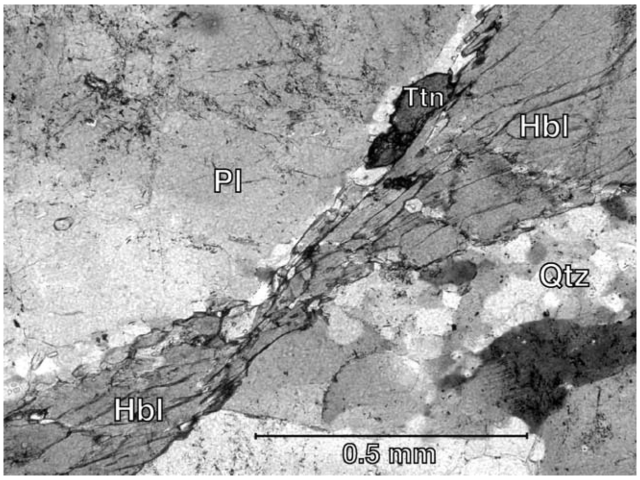

Fig. 17. Aggregate of hornblende thinned down to an incipient folium, apparently by slip on cleavage fragments. Plane-polarized light.

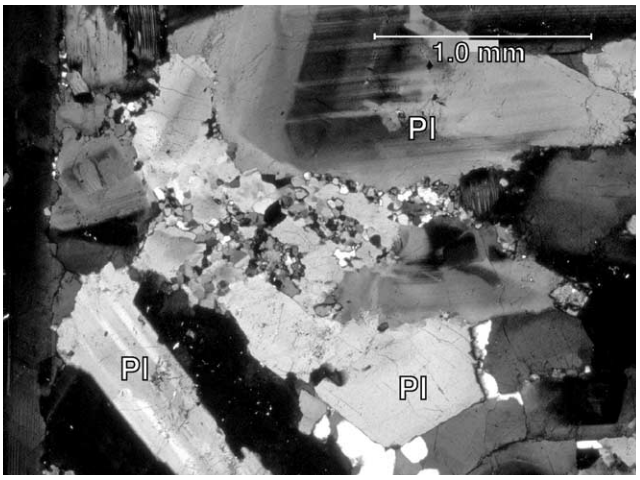

Fig. 19. Recrystallized area in deformed plagioclase. The relatively large grainsize variation in the recrystallized aggregate may suggest the presence of fragments or former fragments Crossed polars.

4.3. Plagioclase deformation

Plagioclase deformation is mostly restricted to local bending, accompanied by deformation twinning, in relatively weakly deformed tonalite. A few crystals are locally intensely bent, with the formation of kink bands. However, some crystals show apparent recrystallization along internal deformation zones, and in the form of small intergranular aggregates of polygonal plagioclase grains (Fig. 19). Some of these occur in equant patches, whereas others occur along plagioclase grain boundaries, with or without quartz, less commonly with minor biotite or hornblende (Fig. 20). The extent of the intergranular recrystallization of plagioclase increases markedly in the more strongly deformed rocks (Fig. 21). The initiation of plagioclase recrystallization is generally thought to occur at temperatures of approximately 500 °C (Tullis et al., 2000), which we consider to be the maximum ambient temperature of the country rocks at the time of intrusion (Johnson et al., 2003). The microstructural evidence for abundant recrystallization of plagioclase in the higher-strain parts of the marginal northern unit of the pluton, together with the microstructural evidence for melt-present deformation, are consistent with temperatures near the wet tonalite solidus during deformation ($T \sim 680$ °C at 300 MPa; Wyllie, 1977; Schmidt and Thompson, 1996).

The intergranular plagioclase aggregates tend to occur preferentially on plagioclase–plagioclase boundaries approximately parallel to the foliation, suggesting possibly an attempt of the primary plagioclase grains to slide past each other. The aggregates shown in Figs. 19, 21 and 22 suggest that at least some fracturing may have been involved, because of the marked grainsize variation and/or rectangular grain shapes. Locally, areas of apparently recrystallized plagioclase pass into damaged zones of healed fractures and fragmentation, suggesting a close relationship between the apparently recrystallized aggregates and fracturing. Moreover, Fig. 23 shows zones of apparent recovery and incipient recrystallization involving relatively rectangular subgrains and grains, along fractures

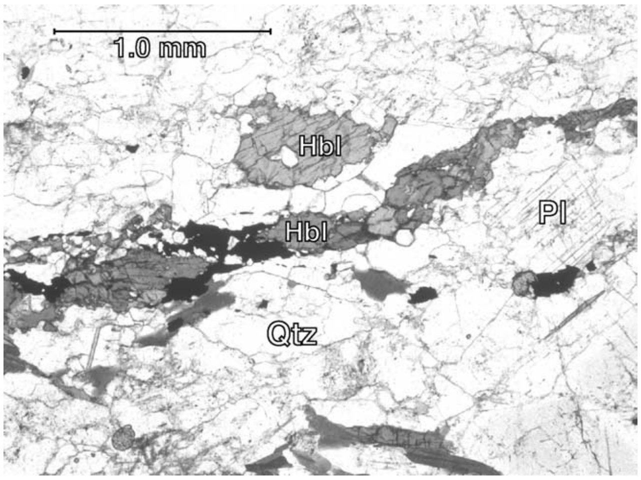

Fig. 18. Folium of recrystallized hornblende in a strongly deformed rock of the northern marginal zone. Plane-polarized light.

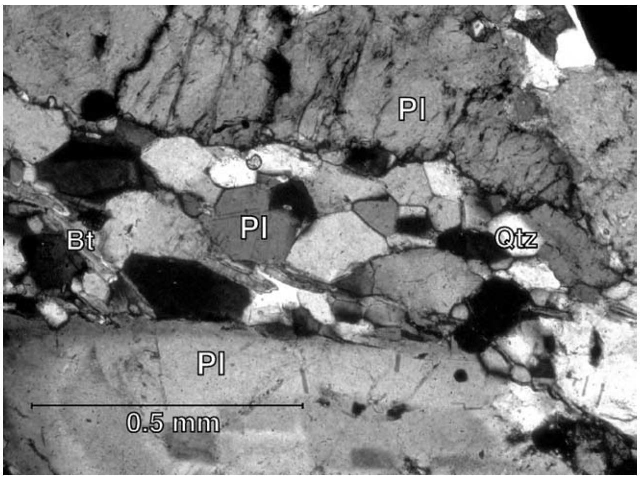

Fig. 20. Recrystallized aggregate of plagioclase, with some quartz and biotite, along a hornblende–plagioclase boundary. Crossed polars.

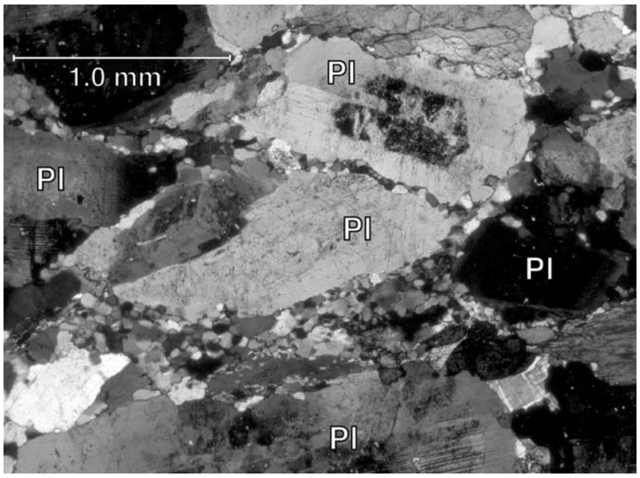

Fig. 21. Extensive intergranular recrystallization of plagioclase, the new grains showing a relatively large variation in size, which suggests a possible origin as fragments. Crossed polars.

and possible healed fractures in plagioclase, suggesting that the recrystallized aggregates may have originated as small fragments. However, the aggregate shown in Fig. 20 has low-energy polygonal grains of quartz, and plagioclase, as well as low-energy biotite crystal faces (Voll, 1960; Kretz, 1966; Vernon, 1968), without evidence of fragmentation.

Though recrystallization microstructures are generally inferred to form by dislocation creep and recovery, it may be difficult to distinguish optically the products of this process from those formed by fracturing (cataclasis) along microfractures to produce small fragments, followed by growth and grain-boundary migration causing the fragments to look like dislocation-controlled recrystallized aggregates. The possible development of polygonal aggregates from fragments in plagioclase has been pointed out by Vernon (1975) and Stünitz (1998) and, as noted above, a fracture mechanism for the development of recrystallized aggregates in amphibole has been proposed by Nyman et al. (1992).

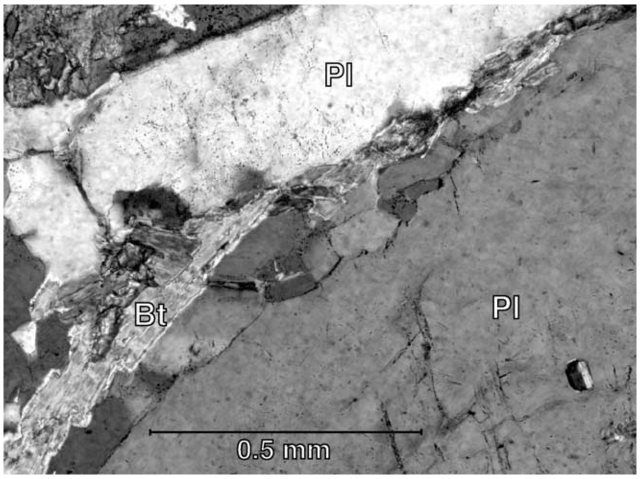

Fig. 22. New grains of plagioclase along a former grain boundary, with roughly rectangular shapes that are consistent with an origin by fragmentation. Crossed polars.

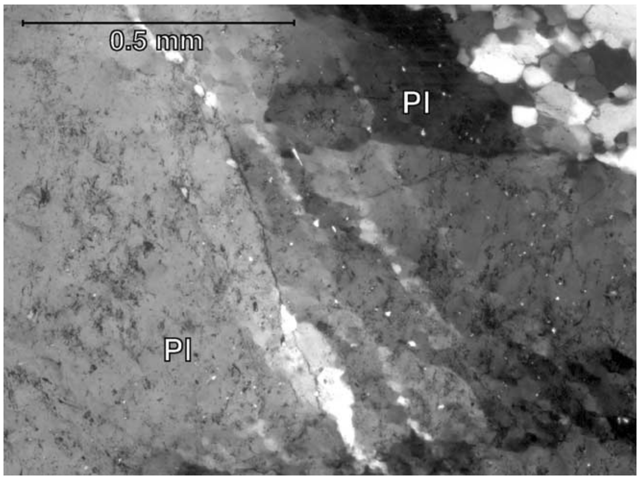

Fig. 23. Zones of recrystallization involving relatively rectangular grains along fractures in plagioclase, suggesting that the recrystallized aggregates may have originated as small fragments. Crossed polars.

Furthermore, Means (1989, fig. 7g, p. 173) observed thymol deforming by brittle processes to produce a 'cataclasite', after which the material transformed by plastic deformation of the fragments, dynamic recrystallization, and void-healing into an aggregate resembling a quartz mylonite.

Plagioclase in some of the least deformed rocks shows very fine-scale, strain-induced grain-boundary migration and apparent incipient recrystallization. More broadly sutured plagioclase/plagioclase, plagioclase/quartz and quartz/quartz boundaries are common in the more deformed rocks, and these also may have been formed by strain-induced grain-boundary migration. In places, hornblende–plagioclase boundaries are also sutured.

Some plagioclase crystals are slightly misoriented across former fractures, the fractures having been healed by more sodic plagioclase (Fig. 24) with abundant, small secondary fluid inclusions, or by more sodic plagioclase with fine-grained biotite and/or epidote. More sodic plagioclase also appears to have replaced the primary plagioclase adjacent to

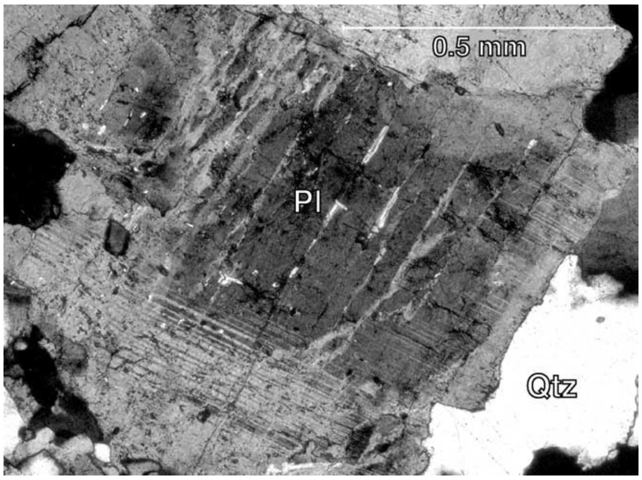

Fig. 24. Fractures in plagioclase marked by thin zones of more sodic plagioclase. Crossed polars.

the fractures. In places, deformation twins appear to have nucleated on fractures, suggesting interplay between brittle and ductile processes (McLaren and Pryer, 2001). In other places, twins have been offset by microfractures in damaged areas against strong hornblende grains. In the more strongly deformed rocks, some plagioclase grains show extensive networks of healed fractures (Figs. 24 and 25), some being filled with epidote, but most with more sodic plagioclase. Some of these damaged areas also show partial replacement by white mica and epidote. Some fracture networks have induced what appears to be incipient fragmentation of the plagioclase, lines of slightly misoriented segments occurring along the former fractures (Fig. 23), as discussed previously. A few plagioclase grains show local kinks passing into discontinuous fractures that are healed by more sodic plagioclase. Rare grains have fractures filled with single grains or decussate aggregates of biotite, some having been later partly pseudomorphed by chlorite. Fracturing and incipient fragmentation of plagioclase crystals are very extensive in the more strongly deformed rocks.

Some fracture-induced openings in plagioclase contain coarse-grained aggregates and single grains of quartz, more sodic plagioclase and/or K-feldspar, as described in Section 5.2. These occurrences are consistent with melt-present deformation. However, the occurrence of epidote and white mica in some fractured zones in plagioclase suggests either that fracturing continued into the subsolidus stage of deformation or that hydrous solutions took advantage of existing fractures.

In the more strongly deformed rocks, many plagioclase grains appear to have indented and partly removed material from adjacent plagioclase grains, as indicated by truncated zoning along irregular boundaries, which may be evidence for contact melting (Park and Means, 1996), as discussed in Section 5.3.

As mentioned previously in connection with biotite deformation, the corners of some plagioclase grains appear to have been truncated by biotite-rich microshears, which may be evidence for the process observed experimentally in octachloropropane by Bons and Jessell (1999). An inferred early stage in this process is shown in Fig. 26.

4.4. K-feldspar deformation

K-feldspar is present in very small amounts in the main part of the San José tonalite, but is more abundant in some of the outermost rocks of the pluton. The K-feldspar shows abundant microcline twinning, which is characteristic of K-feldspar that has been deformed in the solid state (Eggleton, 1979; Eggleton and Buseck, 1980; Bell and Johnson, 1989; Brown and Parsons, 1989). Fitz Gerald and McLaren (1982) found that microcline twinning did not occur in the absence of stress in their experiments. Uncommon flame microperthite is also consistent with solid-state deformation (Pryer and Robin, 1995, 1996; Vernon, 1999). Replacement of the microcline by myrmekite is very common, as is typical of K-feldspar deformed in the solid state (Vernon et al., 1983; Simpson and Wintsch, 1989; Vernon, 1991). Recrystallized and partly recrystallized myrmekitic aggregates are incorporated into fine-grained folia in the most strongly deformed rocks (Fig. 27b), as in the deformed granite described by Vernon et al. (1983).

4.5. Quartz deformation

Primary quartz has been locally recrystallized, forming small equant patches of polygonal grains, even in weakly deformed tonalite. In some rocks, the boundaries of recrystallized grains are sutured, evidently as a result of strain-induced grain-boundary migration. Commonly the sutures have steps, presumably in an attempt to produce lower energy coincidence boundaries (McLaren, 1986; Kruhl, 2001; Kruhl and Peternell, 2001, 2002). Sutured boundaries become more abundant with increasing

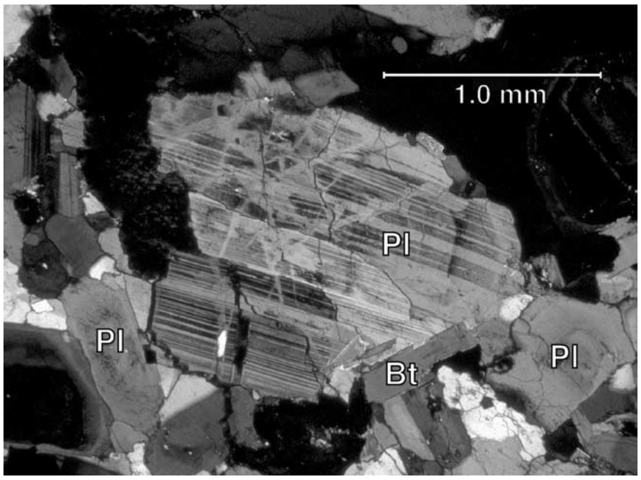

Fig. 25. Angular fracture network in deformed plagioclase, marked by thin zones of more sodic plagioclase. Crossed polars.

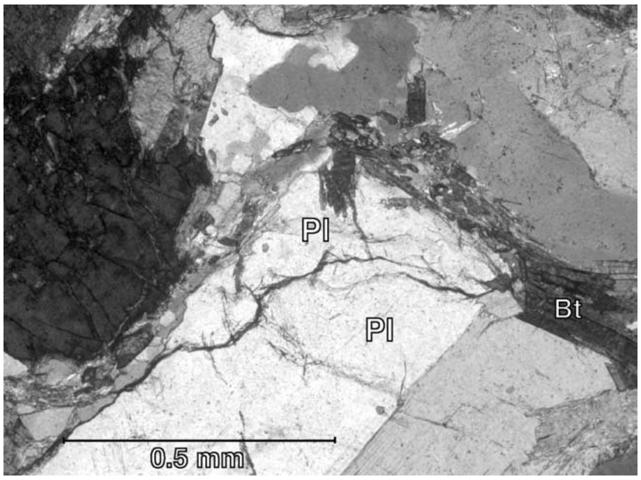

Fig. 26. Irregular fractures that may mark an incipient stage in the removal of a corner of a plagioclase crystal. Plane-polarized light.

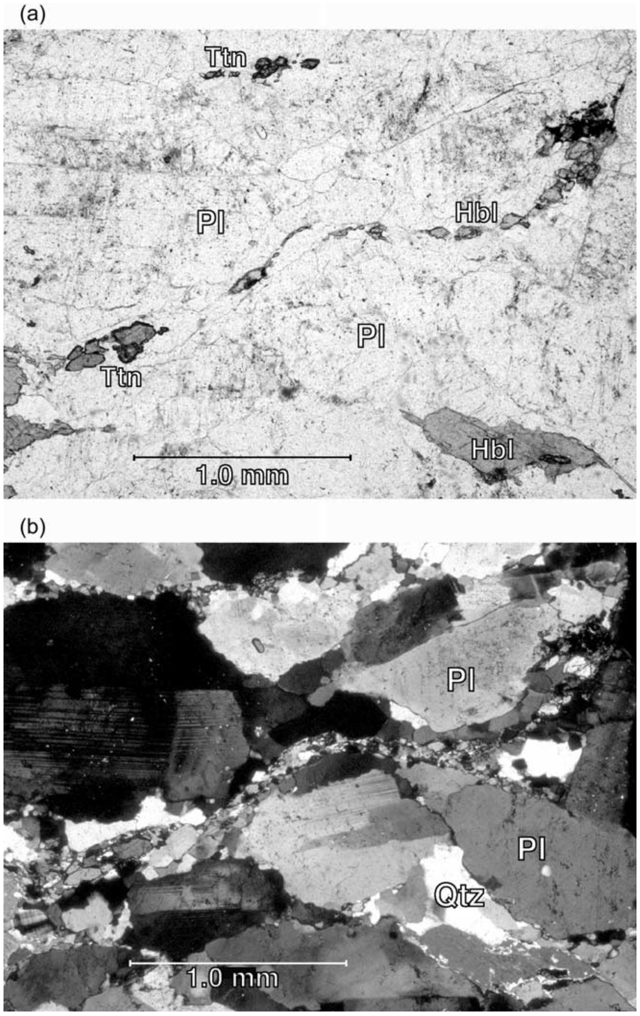

Fig. 27. (a) Folium rich in recrystallized and partly recrystallized myrmekite, with dispersed fragments of hornblende and titanite, in strongly deformed tonalite. Plane-polarized light. (b) Same field of view, showing the myrmekite. Crossed polars.

deformation, and new (recrystallized) grains have formed by bulge nucleation.

Recrystallized and non-recrystallized primary quartz grains show undulose extinction, owing to recovery in response to plastic deformation, without evident grain elongation. Mostly the subgrains indicate a-slip, with local evidence of c-slip (Blumenfeld et al., 1986). C-slip is common in granites deformed near their solidus temperatures (e.g. Blumenfeld et al., 1986; Gapais and Barbarin, 1986; Mainprice et al., 1986; Paterson et al., 1989; Vernon, 2000). Some apparent subgrains may be slightly misaligned fragments (Urai et al., 1986; Lloyd and Freeman, 1994; den Brok et al., 1998).

In the more strongly deformed rocks, quartz aggregates are more elongate, and discontinuous layers of coarse-grained quartz may occur, probably owing to elongation and recrystallization of former single quartz grains, although the possibility of quartz veining cannot be excluded (Vernon, 2004).

4.6. Titanite and epidote deformation

Both primary (coarse-grained, euhedral) and secondary (fine-grained, replacing biotite and hornblende) titanite is present. Some of the primary titanite shows evidence of brittle deformation, and in the most strongly deformed rocks, very small, inferred fragments of titanite are distributed along fine-grained biotite, quartz–plagioclase and quartz–myrmekite folia (Figs. 12 and 27a). However, some of these are difficult to distinguish from secondary titanite. In addition, some epidote grains in folia could have originated as fragments, though some could also be due to hydrothermal alteration of plagioclase. These small grains could have contributed to the deformation, in view of the experimentally observed weakening effect of fine-grained aggregates formed by chemical reactions in plagioclase (Stünitz and Tullis, 2001). Therefore, both minerals appear to have been involved in the deformation and foliation development, at least in the later stages.

5. Discussion

5.1. Evidence of brittle deformation

Microstructural evidence of brittle deformation consists of extensive fracturing of all primary minerals, and possibly some of the secondary titanite and epidote. Fracturing is especially noticeable in plagioclase, which locally shows extensive networks of healed fractures, commonly with increased concentrations of fluid inclusions and secondary white mica, epidote and more sodic plagioclase, as described previously. This suggests increased access of hydrous solutions in these damaged areas. In addition, extensive local fracturing occurs in hornblende, and irregular fractures have also been observed in biotite. Apparent fragments of hornblende and titanite in some folia, as well as abundant inferred cleavage slices of biotite, also suggest that brittle processes operated during at least part of the formation of the foliation.

The problem is to determine when the fracturing occurred. Some of it may have occurred when deuteric solutions were active, because some of the fractures in plagioclase are filled with more sodic plagioclase, epidote or white mica and replacement of plagioclase by these minerals is commonly extensive in fractured zones. Alternatively, deuteric solutions could have taken advantage of fractures formed at supersolidus conditions. On the other hand, some fracture-controlled openings in plagioclase have been filled with what we infer to be crystallized melt (see next section), which would require fracturing at supersolidus conditions.

As discussed previously, much evidence of brittle deformation in the fine-grained folia may have been removed during slow cooling of the pluton, although it is typically preserved in the deformed relics of the primary grains, as well as in the least deformed rocks of the marginal northern unit (Johnson et al., 2004). This applies especially to decussate aggregates of biotite, but may also apply to aggregates of plagioclase and quartz showing low-energy grain shapes, with or without biotite.

5.2. Evidence of melt-present deformation

Evidence that melt has moved during solid-state deformation is critical for the inference of submagmatic flow (Paterson et al., 1989; Vernon, 2000). Late magmatic minerals in 'pressure shadows' or in zones between fragmented primary grains (e.g. Gapais and Barbarin, 1986; Paterson et al., 1989) indicate solid-state deformation in the presence of migrating melt. For example, Bouchez et al. (1992) and de Saint Blanquat and Tikoff (1997) suggested that quartz- and/or feldspar-filled fractures in plagioclase indicate submagmatic flow. Bouchez et al. (1992) described and illustrated fragmented plagioclase crystals with intervening spaces filled by quartz and lower temperature feldspar, such as more sodic plagioclase, which may form fringes on the fragments, with twinning continuous from primary crystal to secondary overgrowth. They observed that some spaces are filled with quartz and two feldspars and, in some examples, the quartz is optically continuous with immediately adjacent primary quartz.

In the most strongly foliated tonalite of the marginal northern unit, numerous plagioclase grains have been fractured and mildly boudinaged, the interboudin zones being filled with aggregates or large single grains of quartz and/or more sodic plagioclase, rarely together with minor biotite. In uncommon rocks with K-feldspar, single-grains or coarse-grained aggregates of K-feldspar (converted to microcline by later deformation, as mentioned previously) fill fracture-controlled openings in plagioclase (Fig. 28a and b); some fracture fillings show matching walls. Commonly, more sodic plagioclase has been deposited on the walls of the vein, in twin continuity with the primary plagioclase (Fig. 28a and b).

Rarely, the plagioclase interboudin zones contain quartz, microcline and plagioclase that is more sodic than the primary plagioclase. The more sodic plagioclase typically has crystallized on the walls of the inferred fracture, in twin continuity with the primary plagioclase. These uncommon three-mineral aggregates are our strongest evidence for crystallization of residual melt. However, single-mineral interboudin fillings (quartz, plagioclase or K-feldspar) are also consistent with crystallization from melt, different minerals nucleating in different openings.

For these examples, the question arises as to whether the vein-filling minerals were deposited from a melt or hydrothermal solution, or even whether they were deformed in the solid state. Deposition from hydrothermal solution cannot be ruled out for quartz fillings, and the most deformed rocks have a few examples of irregular fine-grained quartz–feldspar aggregates, with some titanite and hornblende, cutting boudinaged plagioclase; these are consistent with emplacement of the quartz by ductile deformation, especially because they form part of continuous folia extending well outside the primary plagioclase grain.

However, a strong argument in favor of crystallization of melt in a single local cracking event is that the openings extend only to the edge of the primary plagioclase crystal, and that no fractures or ductile deformation features occur in the adjacent grains to accommodate strain associated with slow or episodic boudinage of the plagioclase. Instantaneous penetration of melt along the plagioclase boundaries and into the opening presumably would allow the opening to form without the need for strain accommodation by deformation of the adjacent grains.

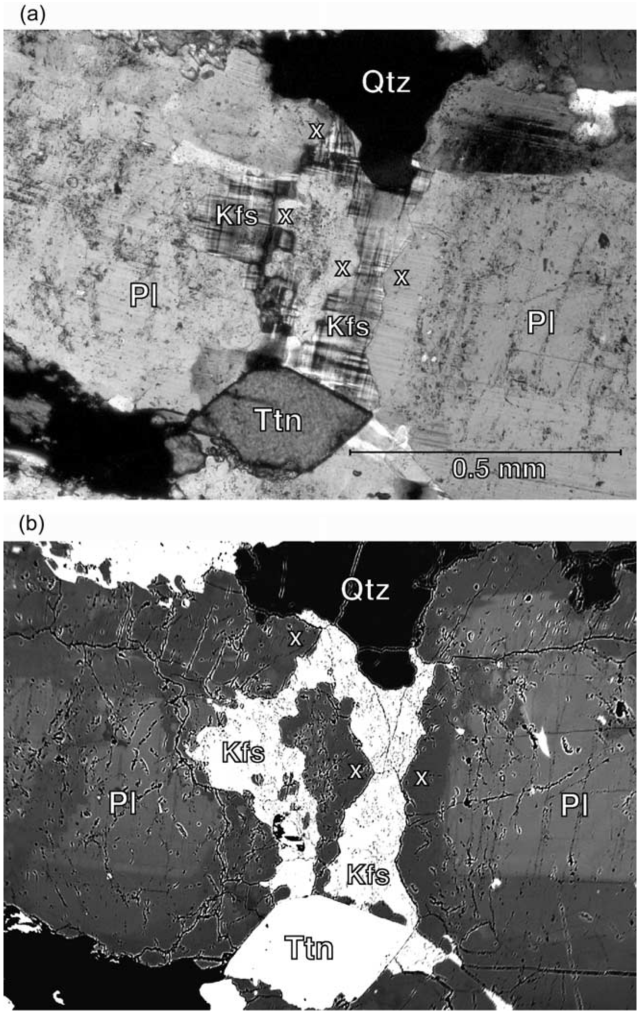

Fig. 28. (a) Single-grain of K-feldspar (now microcline) filling two fracture-controlled openings in plagioclase. Clear (inclusion-poor), more sodic plagioclase (x) has crystallized on the walls of the openings, in twin continuity with the primary plagioclase. These areas typically show compositions of An_{23}, whereas adjacent portions of original grain typically show An_{41}. Some of the indentations on the walls of the openings can be matched on the opposite wall, especially the dovetail indentations in the vein at the left. Crossed polars. (b) Backscatter electron image of area shown in (a). Note darker gray color of more sodic plagioclase. See (a) for scale.

Rare features that also support melt infiltration are openings partly filled with single-crystal microcline making lenticular terminations along the former cracks (Fig. 29). Equally rare occurrences of decussate aggregates and single grains of biotite occupying openings in plagioclase and grains of biotite in cleavage-controlled openings in hornblende conceivably could also represent crystallization of magma, as the residual liquid would have been saturated in biotite until completely solidified.

In view of the uncommon occurrences of inferred melt-filled fractures, we suggest that the amount of melt present during the deformation was not large enough to prevent predominant grain–grain contacts, accounting for the abundant evidence of fracturing and dislocation creep. However, we speculate that it may have been large enough to assist deformation by lubricating potential slip surfaces along some grain boundaries, although no clear microstructural evidence of this process has been observed. Penetration would be most effective if the residual melt was very hydrous.

Although he noted evidence for synmagmatic fracturing, Rosenberg (2001, p. 73) stated: “Except for pseudotachylites, cataclastic flow in partially molten crustal rocks has not been observed in nature.” The foregoing evidence of brittle deformation in the presence of melt, suggests that the marginal deformation of the San José pluton may be a natural example of predominantly brittle deformation in rock with a small proportion of melt. Emplacement-related deformation of earlier, partly crystallized granitoids may provide an ideal setting for brittle deformation, owing to the presence of melt and the high strain rates that may accompany emplacement (Johnson et al., 2001; Gerbi et al., 2004).

5.3. Evidence for contact melting

Clear microstructural evidence of melting at grain contacts would provide additional support for an interpretation of melt-present deformation. Park and Means (1996) used the term ‘contact melting’ to describe a solution process in deforming melt-present systems, similar to pressure solution or strain solution in melt-absent systems. They referred to it as ‘contact melting’, so as not to imply that the state of stress or high defect densities were playing a direct role in the dissolution. They observed that contact melting generally occurs at boundaries oriented at high angles to the direction of shortening, suggesting that the melting could be related to high normal stress at these boundaries. As with pressure solution and strain solution, contact melting accommodates strain, and so may be an important deformation mechanism in crystal mushes at the submagmatic stage of deformation.

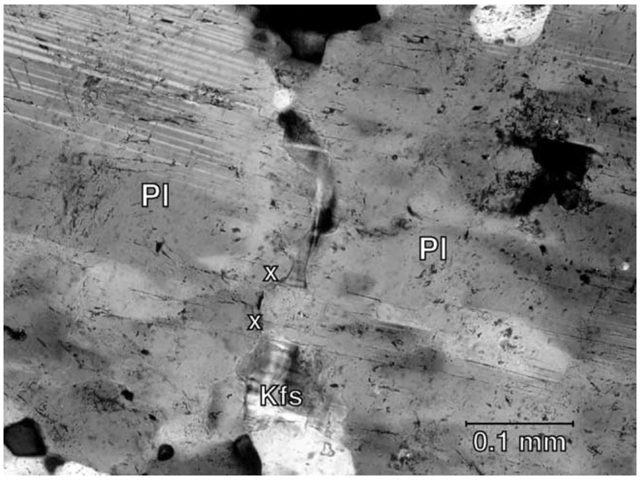

Fig. 29. Curved crack partly filled with single-crystal microcline making lenticular terminations (x) along the crack. Crossed polars.

Nicolas and Ildefonse (1996) and Rosenberg (2001) inferred contact melting from photographic evidence of truncation of mineral zoning in plagioclase. Nicolas and Ildefonse (1996) noted that truncated zoning can be explained by (1) growth of two crystals already in contact or (2) stress-driven grain dissolution or melting, which they called ‘indentation’, of one crystal by the other. They ruled out the first interpretation because they expected zoning to at least partly surround the contact surface. However, if one crystal nucleates heterogeneously on the other, complete zoning would not be expected. For example, sections through concentrically zoned, simply twinned, single crystals can resemble truncated zoning (Dowty, 1980, fig. 1), as can sections through concentrically zoned crystals, one of which nucleates epitaxially on the other (Dowty, 1980, fig. 2). The same result can even be obtained from sections of adjacent crystals that nucleate obliquely on other crystals, which is common in plagioclase aggregates in basalts (e.g. Kirkpatrick, 1977).

In both examples of truncated zoning illustrated by Rosenberg (2001, figs. 5a and 9), the crystal inferred to have been indented has its core right against the crystal inferred to have been the indenter, and does not appear to show any continuous zones around the core. This could be interpreted as indicating heterogeneous nucleation of the ‘indented’ crystal on the other one, rather than contact melting.

Separated cores with at least some later zones completely surrounding them suggest separately nucleated crystals in which truncated zoning is consistent with removal of material, either by solution or contact melting. The truncating, irregular contact shown in Fig. 30 may be an example of contact melting, especially in view of the other evidence for melt-present deformation in the San José rocks. However, irregular contacts between plagioclase grains (e.g. Rosenberg, 2001, fig. 9) are common in granitic rocks, and so cannot be taken to indicate contact melting in the absence of other evidence, although both this process and grain-boundary migration are possible interpretations in deformed granites.

Evidently care should be taken when evaluating these microstructures. Nevertheless, if the two impinging crystals, both with truncated zoning, have well separated cores, and if each core is completely surrounded by some zones (Fig. 30),

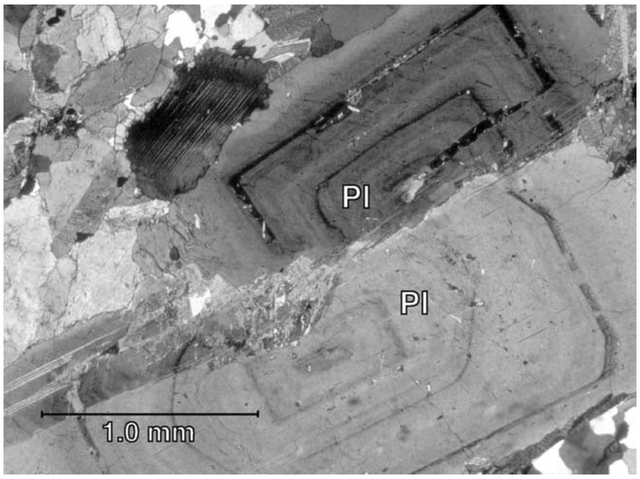

Fig. 30. Two zoned plagioclase grains with an irregular contact that may have resulted from contact melting. The cores of both grains are separated, and are surrounded by some oscillatory zones, suggesting initial separation of the two crystals, rather than nucleation and growth of one on the other. Crossed polars.

separate nucleation and growth of the crystals followed by impingement can be inferred. Two alternative explanations for the truncated zoning are: (1) fortuitous impingement and consequent cessation of growth at the contact, and (2) contact melting. A third possible alternative interpretation is that large fragments of plagioclase, with zoning truncated by fracture surfaces, impinge and grow together in a kind of sintering process. However, in the San José rocks, inferred plagioclase fragments are typically much smaller than the original crystals, and occur along original grain boundaries and in microshear zones. Large fragments, similar in size to the primary grainsize of the rock, as described for melt-filled interboudin zones, are uncommon.

5.4. Deformation history from a spatial deformation gradient?

An important question is whether the deformation and microstructural history of a high-strain rock near the margin of the San José pluton can be inferred by evaluating the progressive microstructural evolution along transects from the margin inwards. Means (1995) discussed two end-member shear-zone types, one of which (Type 1) widens with time as deformation ceases in the most highly strained rocks, owing to work hardening, and the other (Type 2) maintains constant width, but the most actively deforming zone narrows with time. Means (1995) noted that Type 2 shear zones are much more effective recorders of rock history, because a large part of the history is preserved in a traverse from low to high strain.

One way to evaluate the San José pluton is to consider the marginal northern unit as a deforming shell undergoing some component of outward stretching. In this instance, the width of rock affected by solid-state deformation would increase with time as the solid-state deformation front migrates inward from the pluton margins (similar to a Type 1 shear zone). Additional widening of this zone of solid-state deformation may (or may not) occur owing to (a) crystallization as the rocks cool, and (b) expulsion of melt owing to partitioning of deformation, with zones of highly non-coaxial deformation forming linked pathways for melt migration (Vigneresse and Tikoff, 1999; Rosenberg and Handy, 2000; Brun and Vigneresse, 2002).

Such a deforming shell with a component of outward stretching forms an unusual shear zone in that deformation cannot easily stop in one part of the stretching zone while continuing in another, unless the part that has stopped deforming is isolated by boudinage. We have found no compelling evidence for large-scale boudinage in the marginal northern unit. Thus, even though the zone of solid-state deformation widened with time, the progressive evolution of microstructure seen from low-strain to high-strain rocks across the zone may reflect the microstructural evolution experienced by the most highly strained rocks. This would allow us to infer progressive histories of individual rocks from a microstructural progression.

Although the stretching shell hypothesis described above is consistent with the data that we have collected in and around the San José pluton, as well as the boundary conditions that we have inferred for the deformation of the marginal northern unit (Johnson et al., 2003), extracting deformation histories from finite microstructures is a notoriously difficult problem, and further work on the pluton may well lead to the favoring of a different hypothesis.

6. Conclusions

The most general inferences on deformation history that we can make from our microstructural observations are: (1) the deformation was superimposed on an existing magmatic foliation; (2) some submagmatic deformation occurred; (3) abundant brittle deformation occurred, at least some of which may have accompanied submagmatic flow, as indicated by melt-filled fracture-controlled openings in plagioclase; (4) much of the solid-state deformation occurred at relatively high-temperatures, as indicated by local *c*-slip in quartz, grain-boundary movement (recrystallization) in plagioclase and probably hornblende, and formation and recrystallization of myrmekite; (5) crystal–plastic deformation is suggested by apparently recrystallized aggregates of plagioclase, quartz and biotite, and by apparent recovery structures in quartz, although we are unable to determine whether or not brittle deformation was involved in the formation of these aggregates; and (6) brittle deformation continued at subsolidus conditions, as indicated by fracture-controlled openings in plagioclase filled by deuteric minerals.

Crystal–plastic strain may have followed initially rapid, brittle, melt-present deformation, which we infer to be a result of intrusion of later pulses of magma. Alternatively crystal–plastic and brittle deformation processes may have operated simultaneously (McLaren and Pryer, 2001). We infer that the deformation began in the presence of a small amount of melt, and probably continued in the absence of melt. Early melt–present deformation is supported by the occurrence of coarse-grained aggregates and single grains of late magmatic quartz, sodic plagioclase and K-feldspar in inferred fracture-controlled openings in plagioclase, which implies instantaneous fracturing and filling.

The formation and development of the foliation appears to have been controlled mainly by slip on cleavage fragments of biotite, accompanied by the formation of fine-grained folia of quartz or plagioclase, at least some of which may have originated by fragmentation. Hornblende fragments also contributed to the folia in the more strongly deformed rocks. However, much, though not all of the evidence of brittle deformation may have been obscured by grain growth during cooling of the pluton.

Acknowledgements

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