Deformation and exhumation of the Mount Igikpak region, central Brooks Range, Alaska

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ABSTRACT

The Mount Igikpak–Arrigetch region exposes a crustal section more than 15 km thick that includes the highest-grade metamorphic rocks in the central Brooks Range. We carried out detailed field mapping and thermochronologic analyses along a north-south transect from the deepest part of the crustal section, near Mount Igikpak, to its northern flank.

Metamorphic grade along the transect ranges from upper greenschist grade around the Mount Igikpak orthogneiss to very low grade in the Endicott Mountains allochthon to the north. Evidence of two foliations is found throughout the metasedimentary section. The $S_2$ fabric controls the metamorphic layering and the distribution of map units. $^{40}$Ar/$^{39}$Ar analyses from white mica in the lowest-grade rocks indicate that peak metamorphism associated with the development of the $S_1$ foliation occurred before 112 Ma, probably owing to crustal thickening caused by collision of an island arc against the Arctic Alaska margin.

The dominant foliation ($S_3$) formed during rapid cooling from hornblende closure temperature ($500 \pm 50 ^\circ C$) ca. 98 Ma to biotite closure temperature ($-300 \pm 50 ^\circ C$) by ca. 90 Ma. Mineral and elongation lineations associated with $S_2$ plunge to the north, and associated kinematic indicators show top-to-the-north shear, reflecting north-south extension of the crust in mid-Cretaceous time. The episode of extensional deformation took place on a crustal scale and resulted in tectonic exhumation of the core of the orogen. The driving mechanism may have been gravitational collapse of crust that was overthickened during the preceding collision.

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Data Repository item: 2002075 contains additional material related to this article.

INTRODUCTION

The Brooks Range orogen extends for ~1000 km across northern Alaska from east of the Canadian border to the Chukchi Sea, where it disappears abruptly (Fig. 1). The metamorphic rocks of the Brooks Range have been correlated with those of the Seward Peninsula and the Chukotka Peninsula of Russia (Natal’in et al., 1999). In Early Cretaceous time a continuous compressional orogenic belt probably extended across the Bering Strait. Extensional tectonic processes that affected the Bering-Chukchi region during the mid-Cretaceous (Dumitrut et al., 1995) and early Tertiary (Tolson, 1987) contributed to the collapse of the orogen and the formation of the wide expanse of submerged shelves and basins that are imaged by the deep crustal seismic line presented in this volume (Klemperer et al., this volume, Chapter 1). Thus the tectonic history of rocks exposed today in the Brooks Range has a direct bearing on the evolution of the Bering-Chukchi crust.

We chose to study the Mount Igikpak area, located 350 km east of the Chukchi Sea, because the highest-grade metamorphic rocks of the central Brooks Range are exposed in a large-scale antiformal structure (Fig. 1). The north flank of this antiform exposes a crustal section more than 15 km thick that presumably was once above the Mount Igikpak orthogneiss (Figs. 2 and 3). Metamorphic grade ranges from lower amphibolite facies in the core of the Mount Igikpak gneiss to subchlorite grade in the sedimentary rocks of the Endicott Mountains allochthon north of the Atalna River (Nelson and Grybeck, 1981). This structural relief makes it possible to study deep-seated features that are not exposed elsewhere in the region but are critical to understanding the development of its metamorphic infrastructure.

In the Brooks Range of northern Alaska (Fig. 1) large-magnitude shortening within the thrust belt has been attributed to the Late Jurassic to Early Cretaceous collision of an island arc against the Arctic Alaska margin along the present-day southern boundary of the range (Roeder and Mull, 1978). It has been proposed that metamorphic fabrics exposed in the internal part of the range also formed during this collisional event (Gottschalk, 1990; Moore et al., 1994; Till and Snee, 1995). In this chapter we present structural and geochronologic evidence suggesting that the gently north dipping dominant foliation (S_n) in the Mount Igikpak area of the central Brooks Range is related to

Figure 1. A: Location map of Brooks Range modified from Mull et al. (1987) and Moore et al. (1994). DW is Doonerak window. Seismic line is Bering-Chukchi deep crustal seismic profile (Klemperer et al., this volume, Chapter 1). B: Regional cross section through central Brooks Range after Mull et al. (1987) and Moore et al. (1994). Approximate depth to Moho was determined by correlating Bouguer gravity along this cross section with that of TACT transect, where Moho depths are known from seismic refraction and reflection (Fuis et al., 1997).
mid-Cretaceous uplift and exhumation. The S₂ foliation overprints a higher-grade fabric (S₁) presumably related to the Late Jurassic to Early Cretaceous collisional event.

Stratigraphic and seismic reflection data from the northern foothills of the Brooks Range (Cole et al., 1997) indicate that minor north-vergent thrusting in the foreland was active in the mid-Cretaceous at the same time that the metamorphic hinterland was undergoing down-to-the-north ductile deformation and exhumation. This suggests that the two modes of deformation were kinematically linked. Mid-Cretaceous exhumation of the core of the Brooks Range may have been driven by the gravitational instability of an orogenic pile thickened during the preceding collision. Contemporaneous and parallel thrusting and extension have been documented in other collisional orogens such as the Himalayas (Burchfiel et al., 1992), and have been attributed to gravity-driven topographic collapse.
The Mount Igikpak antiform is at first sight similar to the Doonerak antiform, a much better studied structure located near the Dalton Highway to the east (Fig. 1A). The Doonerak structure has been interpreted to be underlain by a duplex of Tertiary age (Oldow et al., 1987; Moore et al., 1997; O’Sullivan et al., 1998). In this chapter we argue for a very different origin for the Mount Igikpak antiform. Our data show that doming of the area started concurrently with north-vergent ductile deformation associated with S₂. Therefore the S₂ fabric in the north flank of the dome is a primary feature and corresponds to a broad down-to-the-north extensional shear zone.

In further contrast to the Doonerak area, the youngest structures mapped on Mount Igikpak are north- to northwest-trending brittle normal faults of probable early Tertiary age. The orientation of these normal faults indicates that this part of the central Brooks Range underwent a period of orogen-parallel extension perhaps even as north-south shortening was active to the east.

**GEOLOGIC SETTING**

The Brooks Range orogen is believed to have originated during collision of an island arc with a portion of the North American continental margin (the Arctic Alaska terrane) during the Late Jurassic and Early Cretaceous (see review in Moore et al., 1994). The Arctic Alaska terrane consists of continental margin rocks of differing lithology, metamorphic grade, and structural style, which, in the central Brooks Range, include the Endicott Mountains allochthon, the Central Belt, and the Schist Belt (Figs. 1 and 2) (Moore et al., 1994). The Angayucham terrane, which occupies the highest structural level, represents oceanic assemblages and parts of the island arc that collided with Arctic Alaska.

The Endicott Mountains allochthon is composed of Upper Devonian through Lower Cretaceous sedimentary rocks imbricated by north-directed Brookian thrust faults. Stratigraphic re-
Figure 3. A: Structural cross section through study area. U-Pb, 40Ar/39Ar, and fission-track ages are projected onto plane of cross section. B: Legend to geological map (Fig. 2) and cross section (Fig. 3A).

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lations in the higher thrust sheets indicate that they have undergone several hundred kilometers of northward displacement (Mayfield et al., 1988; Moore et al., 1994). The strata at the base of the Endicott Mountains allochthon are low greenschist grade, but at higher structural levels sedimentary rocks are not metamorphosed. Data presented in this study suggest that the Endicott Mountains allochthon and the Central Belt are separated by a broad zone of distributed ductile strain.

The Central Belt is composed mostly of greenschist facies metasedimentary and metavolcanic rocks of probable Late Proterozoic to Paleozoic age and includes marble, calcareous schist, chlorite schist, and minor pelitic schist. Granitic gneisses, surrounded by greenschist- to amphibolite-grade metasedimentary rocks, are exposed along the crest of the Brooks Range. The orthogneiss bodies are variably deformed and have yielded Proterozoic and Late Devonian ages. The Central Belt is distinguished from the Schist Belt, which is to the south, by the lack of blueschist facies minerals and the presence of mainly lower Paleozoic marbles of the Skagit Limestone unit (Moore et al., 1994, 1997).

In the Mount Igikpak area (Figs. 2, 3, and 4), rocks assigned to the Central Belt are polydeformed and the dominant foliation has completely transposed original sedimentary layering. Sedimentary structures are detectable only at the highest structural levels of the Central Belt, near the transition into the overlying Endicott Mountains allochthon. South of the Mount Igikpak orthogneiss the nature and location of the boundary between the Central Belt and the Schist Belt are not clear. Along the Dalton Highway, 200 km to the east of our study area, the boundary is a north-dipping ductile shear zone (Moore et al., 1997). Mull (1977) and Till et al. (1988) proposed that the Schist Belt–Central Belt boundary is a backthrust, and Oldow et al. (1998) argued that it corresponds to the basal detachment of the overlying Skagit allochthon. Although we do not have data to resolve this controversy, we choose the first hypothesis because it leads to a simpler kinematic history (Fig. 1A).

Geobarometric studies have shown that peak pressures in the Schist Belt exceeded those in the Central Belt (Patrick, 1995). The Schist Belt is composed of polydeformed greenschist to epidote-amphibolite facies metasedimentary and metavolcanic rocks with relict blueschist facies minerals. Protoliths appear to be mostly Proterozoic to mid-Paleozoic continental margin- to arc-type sedimentary and volcanic rocks (Till et al., 1988). High-pressure–low-temperature metamorphism within the Schist Belt took place during collision that led to underthrusting of the Arctic Alaska continental margin beneath oceanic and arc rocks (Till, 1992a). The timing of this event has been difficult to determine due to effects of younger thermal

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**Figure 4.** Synthetic block diagram showing main textural and structural features of north flank of Mount Igikpak. Inspired by a figure of Malavieille (1993).
overprints, but is thought to have taken place in Middle to Late Jurassic time, between 171 Ma (Christiansen and Snee, 1994) and 142 Ma (Gottschalk and Snee, 1998). Remnants of the overriding oceanic allochthon are represented by the Angayucham terrane, a belt of mafic and ultramafic rocks, chert, and pelagic sedimentary rocks ranging from Devonian to Jurassic age, extending along the southern flank of the Brooks Range. A narrow belt of low-grade metamorphic rocks intervenes between the Angayucham terrane and the Schist Belt and is bounded by down-to-the-south normal faults (Box, 1987; Gottschalk and Oldow, 1988; Little et al., 1994). The presence of this system of normal faults has been cited as evidence that the Brooks Range collisional orogen was modified by extensional deformation after emplacement of the allochthons (Miller and Hudson, 1991; Little et al., 1994).

The structure of the southern boundary of the Brooks Range is further complicated by the Kobuk fault zone (Fig. 1), a complex zone of brittle deformation that operated as a right-lateral strike-slip fault zone during Tertiary time (see review in Avé-Lallemant et al., 1998). This fault is thought to be part of the system of major strike-slip faults that cut through interior Alaska including the Tintina, Denali, and Kaltag faults. The Kobuk fault dies out to the west in the Chukchi Sea and has been linked with Eocene transiensional deformation in the offshore Hope Basin (Tolson, 1987).

**GEOLGY OF THE MOUNT IGIKPAK REGION**

The Mount Igikpak area is mainly within the Central Belt (Figs. 1A and 2). This area was previously mapped in reconnaissance as part of the Survey Pass quadrangle (Nelson and Grybeck, 1980). The Mount Igikpak orthogneiss, exposed at the headwaters of the Noatak River, is overlain to the north by a thick section of probable Paleozoic greenschist-grade metasedimentary rocks; the rocks have a strong foliation that dips consistently to the north at ~25° (Figs. 2 and 3).

Evidence for at least two foliations is found in the surrounding metasedimentary section. S1 is only locally preserved. The younger foliation (S2) is the dominant fabric that controls the geometry and orientation of compositional layering in the metasedimentary rocks. S2 is a high-strain transposition fabric and is axial planar to recumbent isoclinal folds of all scales. The degree of penetrative strain within the metasedimentary sequence decreases upsection and to the north. Likewise, metamorphic grade decreases from upper greenschist grade near the gneiss to very low grade at the north end of the transect (Fig. 3A).

Repetition of distinctive marble panels of Skajit Limestone (Fig. 3A) provides the only direct evidence of thrust imbrication in the Mount Igikpak area (e.g., Nelson and Grybeck, 1980). There is no remaining evidence for brittle fault motion along these contacts, because the entire section has been overprinted by intense ductile strain.

**Mount Igikpak orthogneiss**

The Igikpak orthogneiss is the largest body of metagranite exposed in the central Brooks Range. It crops out as two irregular lobes, but our mapping was confined to the western and larger one (Figs. 1 and 2). There are two texturally distinct phases within the orthogneiss: (1) megacrystic K-feldspar augen gneiss, which constitutes most of the internal portion of the metaplaton (Fig. 5A) and (2) fine-grained unfoliated granite, which is found along its northern rim. K-feldspar porphyry and aplite also occur as border phases.

Throughout the augen gneiss there is a well-developed north-trending mineral and stretching lineation that parallels the lineation in the overlying metasedimentary rocks. S-C fabrics and the asymmetry of porphyroclasts consistently indicate top-to-the-north shear (Fig. 5A). Deformation at ~450–500 °C is evidenced by the dynamic recrystallization of the K-feldspars, which display core-mantle textures, and by the recrystallized rims of myrmekite (Passchier and Trouw, 1996). On the basis of these petrographic and structural data, and the field relationships and cross-sectional geometry (Figs. 2 and 3), we interpret the coarse augen gneiss to represent a deeper structural level than the finer-grained phase of the orthogneiss.

The finer-grained, marginal northern portion of the orthogneiss does not have a pervasive macroscopic metamorphic fabric. Instead, deformation is concentrated along spaced mylonitic bands. In thin section, quartz shows dynamic recrystallization, manifested by mantles of small new grains around larger internally strained quartz grains, and by convoluted grain boundaries.

We attribute the lack of a penetrative metamorphic fabric in the northern portion of the orthogneiss to the rheological contrast between the metagranite and the more readily deformed metasedimentary rocks that overlie it. During deformation, strain was concentrated within the schist and carbonate section, but only along discrete shear zones in the upper portion of the orthogneiss.

There are many contact relationships between the Mount Igikpak orthogneiss and surrounding schists. The upper contact of the gneiss is generally coplanar with the foliation in metasedimentary rocks. In many places a low-angle mylonitic shear zone is developed along the orthogneiss-metasediment boundary (Figs. 2 and 5B). Within this zone, the orthogneiss exhibits dramatic grain-size reduction and quartz ribbons are developed. S-C fabrics and porphyroclasts with asymmetric strain shadows in the shear zone consistently indicate top-to-the-north sense of shear (Fig. 5B). Brittle normal faults, oriented north-south, and with down-to-the-west displacement, also offset the margin of the gneiss (Figs. 2 and 3).

There is evidence for an original intrusive relationship along the northern and western flanks of the gneiss. Skarns are developed in calcareous lithologies (Newberry et al., 1986) and numerous aplite dikes are found adjacent to the orthogneiss.
The aplite dikes, which are concordant with the dominant foliation, rapidly diminish in abundance upsection. Some are conspicuously boudinaged, indicating at least 100% extension in a north-south direction.

Granitic gneisses in the central Brooks Range were initially believed to be Cretaceous in age and syntectonic with Brookian metamorphism (Brosché and Pessl, 1977). Nelson and Grybeck (1978) were the first to suggest a pre-Cretaceous age for the Arridgeth and Igikpak orthogneisses on the basis of textural and field relationships. Later Rb/Sr (Silberman et al., 1979) and U-Pb isotopic work demonstrated that many of the Brooks Range orthogneisses are Devonian, although the U-Pb zircon data are often complex (Dillen et al., 1980; Aleinikoff et al., 1993). Pat-}

ton et al. (1987) carried out U-Pb analyses on samples from the foliated core and the unfoliated northern portion of the Mount Igikpak orthogneiss. Both analyses yielded discordant Devonian ages (Table 1; Fig. 6A).

In an attempt to better define the age of crystallization of the Mount Igikpak orthogneiss, we analyzed zircons from a sample of the unfoliated northern portion of the gneiss (sample 93–JT-88; location in Fig. 2). Three conventional zircon fractions yielded discordant analyses with U/Pb and Pb/Pb ages ranging from 320 to 627 Ma (Table 1; Fig. 6A). To evaluate the zircon systematics of this sample, a fraction of coarse zircon was subject to a step-wise dissolution experiment following the technique of Mattinson (1994) and McClelland and Mattinson (1996). Sequential dissolution steps resulted in progressively increasing U/Pb and 207Pb/206Pb ages (Table 1), such that the analyses define a line that generally parallels the concordia curve (Fig. 6A). The first three steps, interpreted to reflect the effect of Pb-loss on primary magmatic components, define a chord with a lower intercept of 120 Ma. This is consistent with the age of peak metamorphism in the area as independently defined by the 40Ar/39Ar data presented in the following. Step L4 and the residue define a chord from ca. 385 to 1375 Ma. We interpret this chord to reflect a mixture between primary Devonian magmatic zircon and the average age of Proterozoic inherited components. This interpretation is consistent with earlier observations, based on bulk composition and εNd signature, that the Mount Igikpak orthogneiss represents a highly evolved magma with a large component of contamination from Proterozoic continental crust (Nelson et al., 1993; Newberry et al., 1986).

We assign an approximate crystallization age of 375–395 Ma to this phase of the Igikpak orthogneiss, which is consistent with the ages determined for similar metatutonic bodies near the Dalton Highway (Aleinikoff et al., 1993).

The tectonic setting of Devonian magmatism in the Brooks Range is not well understood. Both arc (Dillen et al., 1980) and extensional (Hitzman et al., 1986) environments have been proposed. A possible scenario is a continental arc undergoing extension in the backarc region to account for the inherited crustal component seen in the U-Pb analyses.

Patrick (1995) carried out a thermobarometric study of the Brooks Range orthogneisses and estimated metamorphic conditions for two samples of the Mount Igikpak gneiss to be in excess of 5.5 kbar and 373–416 °C. The moderately high pressure resulted from crustal thickening during Late Jurassic–Early Cretaceous collisional deformation in the Brooks Range.

**Metasedimentary rocks**

The metamorphic section overlying the Mount Igikpak orthogneiss consists of alternating layers of calcareous schist, marble, chlorite schist, carbonaceous slate, and graphitic phyllites (Fig. 3). In addition, minor pelitic schist, micaceous quartzite, metaconglomerate, and small metagabbro intrusives
within the marbles are present. The protoliths are probably Late Proterozoic to middle Paleozoic continental margin sedimentary and volcaniclastic rocks (Nelson and Grybeck, 1980).

Graphic marble overlying the Mount Igikpak orthogneiss contain Middle Ordovician conodonts (Toro, 1998). Massive white marbles assigned to the Skajit Limestone (Silurian to Devonian) are exposed along the divide between the Noatak and Alatna Rivers (Figs. 2 and 3). Graphic phyllites that overlie the Skajit Limestone have been correlated with the Upper Devonian Hunt Fork Shale of the Endicott Mountains allochthon (Nelson and Grybeck, 1980).

Metamorphic fabrics and structures

The study area is divided into two structural domains: (1) a southern high-strain domain adjacent to the Mount Igikpak orthogneiss and continuing northward as far as the upper marble panel of Skajit Limestone and (2) a northern lower-strain domain comprising the phyllitic and slaty rocks that extend into the Endicott Mountains allochthon (Figs. 2, 3A, and 4).

**High-strain domain.** The high-strain domain is characterized by a well-developed schistosity that completely transposes all sedimentary bedding. In addition, north-trending mineral and stretching lineations are present in all lithologies. Axes of isoclinal folds are parallel to these lineations and to the direction of tectonic transport. Two metamorphic fabrics can be recognized. An S₁ foliation is locally preserved, while the dominant fabric, S₂, controls mesoscopic and map-scale layering.

*Early foliation (S₁).* This older fabric is preserved within the high-strain domain in the hinges of folds, in inclusion trails within garnets (Fig. 7A), and by bands of garnet porphyroblasts that have not been entirely transposed into parallelism with S₂. Garnet-biotite schists are present in pelitic lithologies along the northern and eastern margin of the Mount Igikpak gneiss. Microscopic evidence indicates that garnet is associated with this older fabric that predates the dominant foliation. The garnets exhibit spiral inclusion trails and snowball textures that are clearly syntectonic. Coarse porphyroblastic S₁ biotite is locally preserved at high angle to the dominant fabric. Early garnet and biotite are partially replaced by chlorite, muscovite, and new fine-grained biotite S₂ biotite (Fig. 7A). Relict S₁ mica survives in fold hinges and within small domains that were not fully transposed.

Previous workers noted that the garnet-in and biotite-in iso-
grads generally follow the margin of the orthogneiss (Nelson and Grybeck, 1979; Newberry et al., 1986) (Fig. 2). This suggests the possibility that the higher-grade assemblages are related to the intrusion of the orthogneiss. However, in a thermobarometric study on garnet-biotite schists from the north side of the Arrigetch gneiss, where similar relationships exist, Vogl and Patrick (1995) obtained temperatures in the 500-600 °C range and pressures of 6-8 kbar. Because of the dynamic textures ob-
served in garnet, the moderately high pressures and temperatures suggested by thermobarometry, and the evidence of an S₁ fabric throughout the structural section, we conclude that the early metamorphic and deformational event associated with S₁ took place during the initial phase of Brookian metamorphism, rather than with the intrusion of the gneiss in Devonian time. The presence of the higher-grade assemblages in the vicinity of the orthogneiss is a function of the structural level exposed.

**Dominant foliation (S₂).** The gently north dipping foliation (Figs. 3 and 8A) is the most obvious and laterally continuous structural feature of the metamorphic rocks on the north flank of Mount Igikpak. This fabric is axial planar to recumbent isoclinal folds that range from a few centimeters to hundreds of meters in wavelength (Figs. 3 and 7B). The fold axes range from north trending, with east-vergent asymmetry in the high-strain domain, to more east-west trending in the low-strain areas at higher structural levels (Figs. 4 and 8C). All the metamorphic assemblages associated with S₂ in the metasedimentary rocks correspond to medium and low greenschist facies.

The S₂ foliation is defined by well-aligned late biotite, muscovite, and chlorite and by elongated grain-shape fabric in matrix minerals. Chlorite is common, generally growing after early biotite and garnet in the deeper part of the section (Fig. 7A). The metamorphic event associated with S₂ produced retrograde as-

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**Figure 6.** U-Pb zircon data from Mount Igikpak orthogneiss. A: Tera-Wasserberg plot showing both conventional and partial dissolution analyses of sample 93-JT-88. Conventional fractions b and c were abraded. Two samples from eastern and central part of orthogneiss, previously published by Patton et al. (1987), are shown for comparison. B: U release spectra for partial dissolution experiment.
Figure 7. Metasedimentary rocks overlying Mount Igikpak orthogneiss. A: Photomicrograph showing garnet porphyroblast with spiral inclusion trails at high angle to \( S_2 \) foliation and with alteration rim of chlorite. Garnet + biotite + white mica + quartz is \( S_1 \) assemblage, white mica + chlorite + quartz is \( S_2 \) assemblage. Field of view is 7.2 mm wide. Sample was collected near 93-JF-124 locality shown in Figure 2. B: Isoclinal folds in banded marble and quartzite (map unit Prqm).

Assemblages that recrystallized the earlier, upper greenschist-grade minerals.

Mineral and elongation lineations are strongly developed on \( S_2 \) planes in the high-strain domain. They consistently plunge to the north, parallel to isoclinal fold axes (Fig. 8C). Both microscopic and macroscopic kinematic indicators, such as asymmetric strain shadows of porphyroclasts in granitic mylonites, oblique grain-shape fabrics, shear bands, and asymmetric boudins, consistently indicate a top-to-the-north sense of shear. There are few good strain markers, but judging by the degree of quartz grain-size reduction, the complete transposition of all layering, and the intensity of development of the foliation and lineation, it is clear that deformation involved high strains.

Nature of the contacts. Most major lithologic contacts were interpreted as thrust faults during reconnaissance mapping of the metamorphic rocks of the central Brooks Range (Nelson and Grybeck, 1980). However, unambiguous older-over-younger relationships are difficult to find in the field because of poor age control and because a strong \( S_1 \) fabric overprints all contacts. The best evidence for structural imbrication is the repetition of the marbles of the Skajit Limestone.

The base of both panels of the Skajit Limestone is transitional, but the upper contacts are sharp. Mylonitic textures are developed near the top of the upper Skajit panel with strong oblique grain-shape fabric, reduced grain size, and dolomite porphyroclasts with asymmetric tails indicative of top-to-the-north shear. In addition, this marble is involved in large-scale recumbent isoclinal folds so that its upper contact is not a planar fault zone, although it may be a folded fault (Figs. 2 and 3). We conclude that ductile deformation and metamorphism continued after the initial juxtaposition of the units, probably by thrust faults, erasing much of the textural evidence of the original nature of the contacts, and modifying the initial geometry of the structures. Similar relationships are found in the Central Belt along the TACT transect (Moore et al., 1997).

Low-strain domain. North of the upper panel of Skajit Limestone, textures become phyllitic or even slaty. Metamorphic minerals are fine grained, sedimentary layering is only partially transposed, and penetrative strain is noticeably lower than deeper in the section (Fig. 4). The phyllites and slates of the low-strain domain have a spaced crenulation cleavage that is parallel to the dominant fabric in the high-strain domain, while \( S_1 \) constitutes the main penetrative fabric in hand specimen and thin section. We tentatively correlate \( S_1 \) in the low-strain domain with \( S_1 \) in the schists of the high-strain domain.

In the low-strain structural domain, folds are typically open or kink shaped with east-west–trending fold axes (Fig. 8C). The orientation of fold axes in the low-strain domain is in sharp contrast to the high-strain domain where fold axes are consistently north trending and parallel to the stretching lineations. The transition between structural domains appears to be abrupt: fold axes with intermediate orientations are not easily found (Fig. 8C). This suggests that different modes of deformation were at play at different structural levels of the orogenic system. Similar relationships have been documented in other mountain belts. For example, in the western Alps, fold axes and stretching lineations are east-west trending and are parallel to the dominant direction of tectonic transport in the high-grade internal crystalline massifs, while in the external zones of the Briançonnais and the Jura, the fold axes are north-south trending and perpendicular to the direction of tectonic transport (Malavieille et al., 1984).

Late-stage brittle structures

A system of normal faults trending northwest-southeast to north-south cut the Mount Igikpak massif (Figs. 2 and 3) (Nelson and Grybeck, 1980). A very well exposed example of these faults is the Tupik fault (Fig. 3). This fault places marble and quartzite in the hanging wall against hydrothermally altered and
fractured Mount Igikpak orthogneiss in the footwall. The fault surface dips 40°SW, is marked by a zone of silicified breccia several meters thick, and has 1-m-wide millons that plunge downdip. The metamorphic fabric in the quartzites and marbles of the hanging wall is folded by normal-sense drag. The offset on the Tupilak fault is ~1.8 km, based on restoration of the cross section (Fig. 3).

The most spectacular normal fault is exposed in the headwaters of the Reed River (Fig. 2). It places high-level, hydrothermally altered Mount Igikpak orthogneiss on deep-level augen gneiss. The fault surface, which dips 40°W, is marked by a 5-m-thick zone of chloritic breccia underlain by a 30 m thickness of fractured augen gneiss. The fault has a minimum displacement of 1.2 km because the augen gneiss in the footwall extends to the summit of Mount Igikpak (2594 m), high above the metasedimentary roof pendants on the hanging-wall orthogneiss. The maximum displacement is undefined. The brittle nature of these faults, together with thermochronologic data discussed in the following, suggests that they were active during the Tertiary. Their geometry shows that the Brooks Range underwent extension parallel to the length of the orogen at that time.

K-Ar AND $^{40}$Ar/$^{39}$Ar GEOCHRONOLOGY

**Previous K-Ar and $^{40}$Ar/$^{39}$Ar studies**

Turner et al. (1979) reported 80 K/Ar ages from the southern Brooks Range Schist Belt. The ages range from 86 to 756 Ma, but most are in the 100–130 Ma range. This latter cluster of ages was interpreted as resulting from cooling after a green schist facies metamorphic event that ended in the mid-Cretaceous and overprinted earlier blueschist facies assemblages.

A few of the ages reported by Turner et al. (1979) were obtained from schists collected near our field area. White mica from 12 km south of the Mount Igikpak orthogneiss was 114 ± 3 Ma, and three biotite samples from the same structural level were 102–100 Ma. One muscovite from 5 km south of the orthogneiss yielded 106 ± 3 Ma. Four mica samples from the Arrigetch orthogneiss, which occupies a structural position similar to that of the Mount Igikpak gneiss, ranged between 92 ± 5 Ma (muscovite) and 86 ± 4 Ma (biotite). In addition, biotite and hornblende from a skarn at the eastern margin of the Arrigetch orthogneiss yielded 95 ± 3 Ma and 109 ± 3 Ma K/Ar ages (Silverman et al., 1979). These ages are consistent with the pattern Turner et al. (1979) observed along the southern Brooks Range: biotites are slightly younger than white micas, reflecting their lower closure temperature; and ages become younger toward the core of the Brooks Range. The biotite ages from the Arrigetch Peaks, which are some of the youngest K/Ar ages in the Brooks Range, suggest that the orthogneiss was exhumed later than the Schist Belt.

Patrick et al. (1994) mapped a metamorphic field gradient south of the Arrigetch orthogneiss that ranges from amphibolite facies in the north to low greenschist, with relict blueschist facies minerals, to the south. A hornblende sample from the high-grade rocks yielded a slightly disturbed $^{40}$Ar/$^{39}$Ar spectrum and an isochron age of 110 ± 1 Ma, which was interpreted as the time of peak metamorphism. Coexisting muscovite yielded a plateau age of 96 Ma. White mica samples from the lower-grade rocks to the south produced hump-shaped spectra with steps ranging from 102 to 133 Ma, indicative of incorporation of excess argon. In general, these recent $^{40}$Ar/$^{39}$Ar data are in agreement with previous K-Ar work.

**New $^{40}$Ar/$^{39}$Ar results**

We present here 12 new $^{40}$Ar/$^{39}$Ar ages obtained by step-heating experiments (Table 2; Figs. 9 and 10) of samples col-
<table>
<thead>
<tr>
<th>Sample</th>
<th>Lithology</th>
<th>Mineral</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Total Fusion Age (Ma)</th>
<th>Isochron Age (Ma)</th>
<th>MSWD</th>
<th>$^{40}$Ar/$^{39}$Ar</th>
<th>WM Plateau Age (Ma)</th>
<th>Steps Used</th>
<th>$^{39}$Ar Used</th>
</tr>
</thead>
<tbody>
<tr>
<td>93-JT-85</td>
<td>Qtz Mu Schist</td>
<td>White mica</td>
<td>67°31.5'</td>
<td>154°57.5'</td>
<td>72.9</td>
<td>74.5 ± 0.4</td>
<td>3.38</td>
<td>286 ± 12</td>
<td>74.3 ± 0.2</td>
<td>1/7/12</td>
<td>77</td>
</tr>
<tr>
<td>93-JT-88</td>
<td>Granite</td>
<td>K-feldspar</td>
<td>67°30.1'</td>
<td>154°59.4'</td>
<td>48.4</td>
<td>47</td>
<td>9.08</td>
<td>1111 ± 45</td>
<td>47.0 ± 0.2</td>
<td>2/20/21</td>
<td>96</td>
</tr>
<tr>
<td>93-JT-88</td>
<td>Granite</td>
<td>Biotite</td>
<td>67°30.1'</td>
<td>154°59.4'</td>
<td>77</td>
<td>78</td>
<td>26.95</td>
<td>179 ± 23</td>
<td>77.5 ± 0.1</td>
<td>2/11/12</td>
<td>91</td>
</tr>
<tr>
<td>93-JT-96</td>
<td>Mu Marble</td>
<td>White mica</td>
<td>67°34.7'</td>
<td>154°52.8'</td>
<td>96.2</td>
<td>97.2 ± 0.3</td>
<td>2.14</td>
<td>266 ± 51</td>
<td>97.2 ± 0.2</td>
<td>2/10/12</td>
<td>94</td>
</tr>
<tr>
<td>93-JT-113</td>
<td>Graph. Phyllite</td>
<td>White mica</td>
<td>67°30.8'</td>
<td>154°38.7'</td>
<td>113</td>
<td>111.4 ± 1.2</td>
<td>7.93</td>
<td>311 ± 42</td>
<td>112 ± 1</td>
<td>3/8–11–12–12</td>
<td>84</td>
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<tr>
<td>93-JT-124</td>
<td>Bio Gt Schist</td>
<td>Biotite</td>
<td>67°29.8'</td>
<td>154°58.5'</td>
<td>102</td>
<td>102.5 ± 2</td>
<td>2.27</td>
<td>248 ± 7</td>
<td>102 ± 0.2</td>
<td>2/12/12</td>
<td>97</td>
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<tr>
<td>93-JT-142</td>
<td>Bio Gt Schist</td>
<td>Biotite</td>
<td>67°32.4'</td>
<td>154°58.4'</td>
<td>94.9</td>
<td>94.9</td>
<td>15.26</td>
<td>293 ± 15</td>
<td>94.2 ± 0.2</td>
<td>2/8/13</td>
<td>59</td>
</tr>
<tr>
<td>93-JT-143</td>
<td>Bio Gt Schist</td>
<td>Biotite</td>
<td>67°33.1'</td>
<td>154°59.4'</td>
<td>89.2</td>
<td>89.5 ± 0.2</td>
<td>4.11</td>
<td>240 ± 21</td>
<td>89.3 ± 0.1</td>
<td>2/10/11</td>
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<tr>
<td>94-JT-19</td>
<td>Skarn</td>
<td>Hornblende</td>
<td>67°30.1'</td>
<td>155°08.4'</td>
<td>94</td>
<td>95.5</td>
<td>156.54</td>
<td>317 ± 28</td>
<td>96 ± 1</td>
<td>3/5/09</td>
<td>72</td>
</tr>
<tr>
<td>94-JT-79</td>
<td>Augen Gneiss</td>
<td>White mica</td>
<td>67°25.2'</td>
<td>155°02.2'</td>
<td>84.6</td>
<td>84</td>
<td>87.09</td>
<td>311 ± 35</td>
<td>84.2 ± 0.8</td>
<td>2/12/15</td>
<td>95</td>
</tr>
<tr>
<td>94-JT-79</td>
<td>Augen Gneiss</td>
<td>Biotite</td>
<td>67°25.2'</td>
<td>155°02.2'</td>
<td>85.1</td>
<td>85</td>
<td>91.36</td>
<td>305 ± 47</td>
<td>85.6 ± 0.8</td>
<td>2/16/21</td>
<td>90</td>
</tr>
<tr>
<td>94-JT-79</td>
<td>Augen Gneiss</td>
<td>Kspar</td>
<td>67°25.2'</td>
<td>155°02.2'</td>
<td>53.9</td>
<td>47</td>
<td>147.2</td>
<td>806 ± 6</td>
<td>47.0 ± 0.1</td>
<td>2/9/12</td>
<td>89</td>
</tr>
</tbody>
</table>

Note: $J$ is the irradiation parameter; MSWD is the mean square weighted deviation. WM Plateau Age is the weighted mean plateau age of the release spectrum. Steps Used and $^{39}$Ar Used refer to the plateau age. Analyses with prefix 93 were preformed at P.B. Gans's laboratory at UCSB. Analyses with prefix 94 were preformed at M. McWilliam's laboratory at Stanford University. Uncertainties reported are 1 sigma. See Toto (1998) for analytical details. Full data tables can be found in GSA Data Repository (see footnote 1).

Selected along a transect from the core of the Mount Igikpak gneiss to the base of the Endicott Mountains allochthon, 35 km to the north (Figs. 2 and 3). This represents a nearly complete sampling of a crustal transect through the Central Belt of the Brooks Range. We analyzed minerals with different closure temperatures (biotite, white mica, K-feldspar, and hornblende) and from different structural levels in order to characterize the cooling history of each sample and of the area as a whole.

**Mica ages.** These samples are discussed in order of their geographic position from north to south along the transect, i.e., toward increasing structural depth and metamorphic grade. The approximate Ar closure temperature of biotite is 300 ± 50 °C and of white mica is 350 ± 50 °C for moderate cooling rates (McDougall and Harrison, 1988).

Sample 93-JT-113 was collected from a deformed quartz vein concordant with $S_1$ in graphitic slates within the low-strain domain at the northern end of the transect (Figs. 2 and 3). The spectrum has a near-plateau age of 112 ± 1 Ma, calculated using 78% of the released $^{39}$Ar (Fig. 9A). The $^{40}$Ar/$^{39}$Ar intercept for this sample is within error of the atmospheric ratio, indicating that the age is reliable. Because the metamorphic grade in this part of the transect is very low, it is not likely that these rocks were ever heated above the closure temperature of white mica. We interpret the 112 Ma age as dating the growth of $S_1$ mica at this structural level, which probably corresponds to local peak metamorphic conditions.

Sample 93-JT-96 is white mica from a coarse-grained, well-foliated, white marble of the Skajit Limestone from the upper part of the high-strain structural domain (Figs. 2 and 3). The micas are aligned with the dominant ($S_2$) fabric. The sample yielded a reliable near-plateau age of 97.2 ± 2 Ma (Fig. 9B). The $^{40}$Ar/$^{39}$Ar age from this sample may either represent the age of recrystallization of the mica, in which case it would be dating the $S_2$ deformational event, or alternatively, cooling after the metamorphic event.

Five mica samples from rocks that directly overlie the Mount Igikpak orthogneiss, and from the margin of the orthogneiss, yielded ages ranging from 102 ± 0.2 to 74.3 ± 0.2 Ma (Table 2; Figs. 2 and 3). The oldest age came from sample 93-JT-124 (Fig. 9C), a garnet-biotite–white mica schist that contains both garnet and coarse biotite porphyroblasts associated with the $S_1$ fabric. Samples 93-JT-142 and 93-JT-143, also garnet-biotite schists, yielded 94.2 ± 2 and 89.3 ± 0.1 Ma, respectively (Fig. 9, D and E). The remaining two samples from this part of the section were of metamorphic white mica from schists (93-JT-85), which had a $^{40}$Ar/$^{39}$Ar age of 74.3 ± 0.2 Ma, and relict magnatic biotite from the unfoliated portion of the Mount Igikpak gneiss (93-JT-88), which yielded 77.5 ± 0.1 Ma (Fig. 9, F and G). These two ages are significantly younger than the samples from nearby, and are the youngest $^{40}$Ar/$^{39}$Ar mica ages reported so far from the central Brooks Range.

The wide range of ages (ca. 102–74 Ma) in micas from a small area and similar structural levels is difficult to explain.

Figure 9. $^{40}$Ar/$^{39}$Ar and K/Ca spectra. Steps used in calculated weighted mean plateau age are shown in solid black. A, B: White mica samples from shallow structural level. C–E: Biotite samples from schists directly overlying Mount Igikpak gneiss. F, G: Anomalously young biotite and white mica from shallow-level Igikpak gneiss and overlying schist. See GSA Data Repository1 or accompanying CD-ROM for $^{40}$Ar/$^{39}$Ar data tables for samples.

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1GSA Data Repository item 2002075. $^{40}$Ar/$^{39}$Ar Data Tables for Samples, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA, editing@geosociety.org, or at www.geosociety.org/pubs/f102002.htm, or on the CD-ROM accompanying this volume.
There is no field evidence for late magmatic activity that could have thermally reset some of the samples, nor are there any major faults or breaks in metamorphic grade separating the samples. One possibility is that the ages reflect distinct episodes of metamorphic growth, the older ages reflecting a larger component of $S_1$ micas and the younger ages being dominated by $S_2$ micas. However, this explanation cannot be applied to all the samples. One of the youngest ages (77 Ma, sample 93-JT-88) came from what is clearly relict magmatic biotite of the Devonian orthogneiss.

We also analyzed coexisting metamorphic white mica and biotite from the coarse and well-foliated interior of the orthogneiss (sample 94-JT-79), which yielded near-plateau ages of $84.2 \pm 0.8$ and $85.0 \pm 0.8$ Ma (Fig. 10, A and B). We interpret
these ages as dating postmetamorphic cooling. These two analyses behaved as expected, the biotite age being slightly younger than the white mica age and both being younger than most of the mica samples from shallower structural levels.

**Hornblende age.** We obtained a $^{40}\text{Ar}/^{39}\text{Ar}$ age from hornblende from a garnet-bearing skarn collected from the northwestern margin of the Mount Igikpak orthogneiss (sample 94-JT-19, Figs. 2 and 3). A plateau age calculated using the flat portions of the spectrum (Fig. 10C), representing 75% of the released $^{40}\text{Ar}$, is $96 \pm 1$ Ma. That metamorphic hornblende, likely of Devonian age, was completely reset in the mid-Cretaceous, suggests that the peak temperature during the metamorphic event associated with $S_2$ exceeded the $500 \pm 50$ °C closure temperature of hornblende.

**K-feldspar ages.** Two K-feldspar samples (93-JT-88 and 94-JT-79) from the Mount Igikpak gneiss help define low-temperature portions of the cooling history. Both samples yielded $^{40}\text{Ar}/^{39}\text{Ar}$ age gradients (Fig. 10D and E), as is commonly observed in plutonic and metamorphic K-feldspars (Harrison, 1990; Lovera et al., 1989). In both samples the lowest-temperature steps are probably disturbed by excess $^{39}\text{Ar}$ trapped in fluid inclusions. In sample 93-JT-88 (Fig. 10D), the spectrum decreases from a high temperature at $97$ Ma to a pseudoplateau of $50$ Ma. This age pattern probably reflects cooling from above the closure temperature of biotite ($300 \pm 50$ °C), which is $74$ Ma for this sample, down to $200$ °C ca. $50$ Ma. Sample 94-JT-79 (Fig. 10E) yielded an age gradient that climbs from $46$ to $64$ Ma, both samples suggest that cooling through $200 \pm 50$ °C took place during early Tertiary time. This is consistent withapatite fission-track data (see following).

**APATITE FISSION-TRACK DATA**

**Previous apatite fission-track studies**

A total of 22 apatite fission-track ages from the Arrigetch and Mount Igikpak orthogneisses were previously published (Blythe et al., 1997; Murphy et al., 1994; O'Sullivan et al., 1993). The samples from the Arrigetch gneiss indicate cooling through the apatite annealing zone (65–110 °C) between 22 and 25 Ma (Murphy et al., 1994). Three samples from the eastern lobe of the Mount Igikpak orthogneiss, which is geographically closest to the Arrigetch orthogneiss (Fig. 1), averaged $25 \pm 2$ Ma. In contrast, ages from the western lobe of the Mount Igikpak gneiss, the site of our study, were distinctly older, ranging between $42 \pm 6$ and $33 \pm 6$ Ma (samples 77ANS87, 77ADG103, 77ADG128, Fig. 2). The age difference between the western and eastern lobes of the Mount Igikpak orthogneiss was attributed to differential uplift across a north-trending vertical fault separating the two areas that was active sometime between 42 and 25 Ma (Murphy et al., 1994; Blythe et al., 1997).

**New apatite fission-track results**

Only two samples yielded enough apatite to obtain reliable ages (Table 3; Fig. 11). Sample 93-JT-87, from highly strained quartzite near the margin of the Igikpak orthogneiss (Fig. 2), yielded an age of $29.6 \pm 4.0$ Ma with a mean track length of $13.2 \pm 0.3$ μm. This age overlaps within the uncertainty with the previously reported ages from the Igikpak gneiss (O’Sullivan et al., 1993). All the samples from the Igikpak area are characterized by reduced track lengths and often negatively skewed length distributions. This suggests that the samples resided for a considerable time within the partial annealing zone of apatite, and therefore the ages that are reported may not reflect a time of rapid cooling.

Sample 93-JT-106, from low greenschist-grade quartz-pebble metaconglomerate located 20 km north and 9 km up structural section from the orthogneiss (Figs. 2 and 3), yielded an age of $60.0 \pm 8.1$ Ma. The difference between this age and the ca. 42–30 Ma ages from the core of the gneiss indicates that uplift and denudation of the core of the Brooks Range with respect to the flanks was ongoing during the early Tertiary.

**Thermal and tectonic evolution**

We have integrated the available stratigraphic, structural, thermochronologic, and petrological data from the Mount Igikpak area with regional tectonic considerations in order to construct a thermal history (Fig. 12A) and a pressure-temperature

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**TABLE 3. FISSION-TRACK DATA**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Elevation (m)</th>
<th>No. Xls.</th>
<th>Spontaneous</th>
<th>Induced</th>
<th>$P(\chi^2)$ (%)</th>
<th>Dosimeter</th>
<th>Age ± 1σ Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td>93-JT-87</td>
<td>67°31.1'</td>
<td>154°57.5'</td>
<td>1280</td>
<td>15</td>
<td>0.206</td>
<td>74</td>
<td>2.339</td>
<td>839</td>
<td>12</td>
</tr>
<tr>
<td>93-JT-106</td>
<td>67°38.5'</td>
<td>154°43.3'</td>
<td>780</td>
<td>6</td>
<td>0.416</td>
<td>65</td>
<td>2.332</td>
<td>365</td>
<td>12</td>
</tr>
</tbody>
</table>

*Note: Abbreviations: No Xls, number of individual grains dated; Rho-S, spontaneous track density ($\times 10^6$ tracks per square centimeter); NS, number of spontaneous tracks counted; Rho-I, induced track density in external detector (muscovite) ($\times 10^6$ tracks per square centimeter); NI, number of induced tracks counted; $P(\chi^2)$, $\chi^2$ probability; Rho-D, induced track density in external detector adjacent to CNS dosimetry glass ($\times 10^6$ tracks per square centimeter); ND, number of tracks counted in determining Rho-D. Age is the sample fission-track pooled age, calculated using zeta of 385.9 for CNS. See Toro (1998) for analytical details.*
(P-T) path (Fig. 12B) for rocks along our structural transect. The thermal history plot tracks the progress of two points in our structural transect (Figs. 3, 12, and 13): (1) a point at the base of the Endicott Mountains allochthon (path A, shown in white in Fig. 12a) and (2) a point in the metasedimentary rocks that directly overlie the Mount Igikpak gneiss (path B, shown in gray in Fig. 12A). These two points represent the top and bottom of a crustal section through the Brooks Range Central Belt.

We use a geothermal gradient of 25 °C/km to estimate paleotemperature in pre-Brookian time. We assume that a lower geothermal gradient (15 °C/km) existed during the initial phase of Brookian orogeny, as suggested by the development of blueschist facies assemblages in the Schist Belt (Till and Snee, 1995), and that the gradient reequilibrated to 25 °C/km after crustal thickening was accomplished.

The following discussion is keyed to numbered stages along the time-temperature and P-T paths (Fig. 12). In addition, the Mesozoic and Tertiary evolution is schematically illustrated in a series of cross sections (Fig. 13) that incorporate the available geological and geochronological data.

**Stage 1. Preinvasion.** Our burial and thermal histories begin in pre-Devonian time, prior to the intrusion of the Mount Igikpak pluton and prior to deposition of the Endicott Group. A moderate depth of emplacement of ~5 km is consistent with textures observed in the shallow portion of the Mount Igikpak orthogneiss and with the mineralogy of the skarns in the contact aureole (Newberry et al., 1986).

**Stage 2. Intrusion of granite.** The Mount Igikpak granite was emplaced ca. 380 Ma, as determined by our U/Pb data. The
temperature shown at stage 2 in Figure 12A (450 °C) was calculated using a one-dimensional conductive thermal model (e.g., Peacock, 1989). This is the temperature in the country rock 200 m away from an intrusive body comparable in size to the Mount Igikpak orthogneiss (20 km thick), with a magmatic temperature of 825 °C. This calculation is consistent with paleotemperature estimates based on phase relations of the skarn mineralogy (Newberry et al., 1986). Under these conditions, the thermal pulse associated with the pluton would decay within 20 m.y. (Fig. 12).

Stage 3. Postintrusion. This stage represents postintrusion conditions in Late Devonian time (Fig. 12). Although we know little about the tectonic setting of Brooks Range Devonian magmatism, we depict some thermally driven uplift and erosion at this time, in response to the intrusion of the granitic belt.

Stage 4. Deposition of Endicott Group. This stage spans Late Devonian to Early Mississippian time, during which a thick elastic wedge was deposited. The total stratigraphic thickness of the Endicott Group in the Endicott Mountains allochthon is ~4.5 km (Moore et al., 1994). We assume that the full thickness of the Endicott Group was deposited above the location of the Mount Igikpak orthogneiss. This assumption helps to account for the crustal thickness attained during Brookian deformation, leading to a more plausible thermal history during metamorphism.

Stage 5. Deposition of Ellesmerian Group. This stage spans Mississippian to Late Jurassic time. During this time the Ellesmerian Group was deposited along the passive margin of Arctic Alaska. This stage corresponds to a period of slowly increasing temperature and burial.

Stage 6. Onset of Brookian orogeny. Brookian deformation is generally attributed to the collision of an island arc with the Arctic Alaska microplate (Fig. 13A), which led to the development of high-P, low-T metamorphic assemblages in the southern Schist Belt, the emplacement of ophiolitic allochthons on the continental margin, and large-scale crustal shortening within the orogen (Mayfield et al., 1988; Mull, 1982; Roeder and Mull, 1978; Till, 1992a).

The age of initiation of deformation is not directly defined by our data. The earliest indications of Brookian tectonism are the Middle Jurassic cooling ages from the metamorphic sole of the western Brooks Range ophiolites (Wirth and Bird, 1992), but it is likely that initial obduction of the ophiolite took place in an oceanic setting some distance from the continental margin, because shelf sedimentation continued uninterrupted until the Late
Jurassic along the Arctic Alaska passive margin (Moore et al., 1994; Patton and Box, 1989). Reliable $^{40}\text{Ar}/^{39}\text{Ar}$ white mica ages from the southern Schist Belt are as old as 130 Ma (Christiansen and Snee, 1994; Gottschalk and Snee, 1998), indicating that crustal thickening and deformation associated with the collision began prior to this time. However, one relict 171 Ma phengite age reported by Christiansen et al. (1994), from an area of the Schist Belt that yielded mostly Early Cretaceous ages, suggests that an earlier metamorphic event could have affected the southern Brooks Range.

**Stage 7. Crustal thickening.** This stage represents Early Cretaceous time. Most of the shortening and crustal thickening in the Brooks Range is inferred to have taken place at this time (Fig. 13B), based in part on the age of southerly derived synorogenic clastic rocks found in the Endicott Mountains allochthon (Moore et al., 1994). This is the most likely age of high-pressure metamorphism of the continental-margin rocks of the Schist and Central Belts (Till and Snee, 1995), and also corresponds to an episode of very rapid subsidence in the Colville foreland basin, which has been attributed to flexure of the lithosphere due to the emplacement of the Endicott Mountains and higher allochthons (Nunn et al., 1987; Cole et al., 1997). The pressures attained within the Brooks Range orthogynes of the Central Belt during metamorphism have been estimated as 5–8 kbar (Patrick, 1995). We have chosen the lower end of this pressure range for the rocks that directly overlie the Mount Igikpak orthogneiss (gray curve in Fig. 12C). This implies a doubling of the thickness of the overburden from ~10 km to 23 km, assuming an average density of 2700 kg/m$^3$. These values of thickening are consistent with estimates of shortening from the foreland fold and thrust belt to the north (Cole et al., 1997).

**Stage 8. Peak metamorphism.** This stage corresponds to peak temperatures, recorded by the garnet-biotite schists associated with relict $S_3$ foliation in the pelitic metasedimentary rocks that overlie the Mount Igikpak gneiss. The last stages of this metamorphic and deforming event were probably recorded by the $112 \pm 1$ Ma $^{40}\text{Ar}/^{39}\text{Ar}$ age of $S_4$ white mica in the low-grade slates of the base of the Endicott Mountains allochthon, although it is risky to correlate foliations from such different structural levels and metamorphic grades. Thermobarometry from lithologies and structural levels similar to those found adjacent to the Mount Igikpak orthogneiss yielded temperatures in the 500–600 °C range and pressures of 6–8 kbar on the north side of the Arrigetch orthogneiss (Vogl and Patrick, 1995).

**Stage 9. Extensional collapse and exhumation.** This stage represents initial exhumation of the metamorphic rocks of the Brooks Range Central Belt synchronous with ductile deformation that produced the $S_4$ fabrics (Fig. 13C). The dominant foliation ($S_2$) in the metasedimentary rocks that overlie the Mount Igikpak orthogneiss is mostly defined by white mica and chlorite that replaced older biotite and garnet. We interpret this texture as evidence that the $S_2$ fabric formed along a retrograde metamorphic path, as the crustal section was being exhumed (Fig. 12B). This deformational and metamorphic event is also recorded in the $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology from these rocks. Hornblende from the Devonian skarns along the margin of the Igikpak orthogneiss yielded an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 96 ± 1 Ma, while most of the mica ages from this structural level range between 89 and 77 Ma (Figs. 9 and 10). We interpret this pattern of ages as the result of initial rapid cooling from above the hornblende Ar closure temperature (~500 °C) ca. 96 Ma to near the closure temperature of biotite (~300 °C) ca. 90 Ma, followed by a period of residence at this temperature (Fig. 12A).

Many 90–100 Ma mica $^{40}\text{Ar}/^{39}\text{Ar}$ and $K$-Ar ages have been reported from the Brooks Range Schist and Central Belts (Turner et al., 1979; Blythe et al., 1990; Christiansen and Snee, 1994; Patrick et al., 1994). These cooling ages probably date the exhumation of the metamorphic hinterland after earlier high-P metamorphism. It is widely agreed that extension played a role in the exhumation of the deep-level rocks of the Schist Belt (Gottschalk and Oldow, 1988; Miller and Hudson, 1991; Christiansen and Snee, 1994; Little et al., 1994), and our work in the Mount Igikpak region suggests that the same is true for the Central Belt. The driving mechanism for extension is still a matter of discussion (Miller and Hudson, 1993; Till et al., 1993).

Kinematic indicators in the Mount Igikpak region show that the deformation associated with the $S_4$ fabric took place under top-to-the-north shear. The structure exposed today is a north-dipping panel deformed by top-to-the-north shear, with metamorphic grade decreasing upsection (Fig. 4), as expected in an extensional shear zone (Fig. 13C). The variation of $^{40}\text{Ar}/^{39}\text{Ar}$ and fission-track ages along our transect shows that the north tilt of the section is not just a late-stage feature, but has persisted since the mid-Cretaceous.

We propose that rapid exhumation of the Central Belt took place as the overthickened orogenic wedge collapsed under gravitational stress both along the southern Brooks Range extensional system and along the northern boundary of the Central Belt. It is likely that collapse was facilitated by the loss of strength of the crust by conductive heating as the isotherms reequilibrated after arc-continent collision. Probably shallow-level normal faulting accompanied ductile deformation at depth. Low-angle normal faults have been identified at the base of the ophiolitic allochthons found at the highest structural level in the western Brooks Range (Harris, 1988).

The presence of small piggyback basins that contain folded early and middle Albian sedimentary rocks in the northern fold and thrust belt suggests that shortening was active in the foreland while the metamorphic core was collapsing (Cole et al., 1997). These contractional structures probably accommodated the ductile north-directed strain evidenced by $S_2$ in the metamorphic core. The geometry of contemporaneous thrusting and extension in the northern Brooks Range is somewhat different from that documented in the Himalayas. In the Himalayas, sense of shear on the South Tibetan detachment system was in the opposite direction from that along the Main Central thrust (Burchfiel et al., 1992). Thus the intervening rocks were extruded toward the foreland. In contrast, in the central Brooks
Range shear within the Central Belt, metamorphic rocks had the same sense as the structures in the thrust belt, producing a geometry akin to a crustal-scale gravity slide.

Stage 10. Cooling and east-west extension. This stage corresponds to the Late Cretaceous and early Tertiary portion of the thermal history as defined by the age gradients observed in the K-feldspar \(^{40}\text{Ar}^{39}\text{Ar}\) spectra from the orthogneiss and the fission-track data (Fig. 12A). Both methods are compatible with a moderate cooling rate of \(~5^\circ\text{C/m.y.}\) between 70 Ma and the present. However, a more complicated cooling path, which is not resolved by our data, is likely. Rapid cooling events ca. 60 and 25 Ma have been identified along the Dalton Highway, along the range front, and in the northeastern Brooks Range (O'Sullivan et al., 1997, 1998). The 60 Ma cooling event has been attributed to renewed shortening because rocks as young as latest Cretaceous are folded in the North Slope (Chapman and Sable, 1960). Along the Dalton highway, early Tertiary deformation is believed to have produced crustal-scale duplexes that underlie the Mount Doonerak antiform and that were reactivated at 25 Ma (Moore et al., 1997).

We see no direct evidence for Tertiary contraction along our transect. The youngest structural features are north- to northwest-trending normal faults that cut the orthogneiss. The brittle nature of these faults indicates that they were not active until late Tertiary time, when the orthogneiss was within \(-10\) km of the surface. Their timing and orientation suggest that they may be related to the Eocene-Miocene normal fault system in the Hope Basin, located directly west of the Brooks Range in the Chukchi Sea (Fig. 1) (Tolson, 1987). The origin of the Hope Basin has been attributed to transensional deformation associated with right-lateral strike-slip motion on the Kobuk fault, which extends along the southern flank of the Brooks Range. Normal faulting in the Igikpak area may also be linked to movement on the Kobuk fault. To the west of Mount Igikpak the topographic elevation of the Brooks Range decreases progressively, and there are large low-lying areas underlain by alluvium (the Noatak Lowlands, Fig. 1). We speculate that these areas of low topography may also be controlled by normal faults and correspond to the structural transition from the high central Brooks Range to the Hope Basin. The onset of normal faulting in the Hope Basin was contemporaneous with development of several other basins on the Bering Shelf (Worrall, 1991), and correlates with an important change in plate motions in the North Pacific. The rate of subduction along the paleo-Pacific margin of Alaska decreased sharply in the Eocene, and convergence changed to an oblique northwest direction (Engebretson et al., 1985). Other consequences of this plate regime were dextral movement on the major strike-slip faults that cut interior Alaska and development of continental basins in association with these faults south of the Brooks Range (Kirschner, 1994; Pfafker and Berg, 1994). The presence of north- and northwest-trending normal faults in the Mount Igikpak area suggests that the central and perhaps the western Brooks Range were also affected by this regional deformation event.

CONCLUSIONS

Structural and thermochronologic studies of a 15-km-thick crustal section exposed in the Mount Igikpak area reveals that deformation in the Central Belt of the Brooks Range is characterized mostly by penetrative ductile strain. We attribute peak metamorphic conditions and the development of \(S_1\) to crustal shortening and thickening during collision of an island arc against the Arctic Alaska margin during latest Jurassic to Early Cretaceous time. A strong north-dipping foliation \((S_2)\) is well developed throughout the area. Metamorphic grade decreases upsection in the metasedimentary rocks from upper greenschist facies near the Mount Igikpak gneiss to very low grade at the northern end of the transect. Kinematic indicators, both macroscopic and microscopic, yield top-to-the-north sense of shear. Chlorite, white mica, and biotite associated with the dominant foliation overprint earlier, higher-grade minerals related to \(S_1\). This demonstrates that ductile deformation within the Central Belt occurred during the retrograde path of the PT history, while exhumation was taking place.

Our \(^{40}\text{Ar}^{39}\text{Ar}\) thermochronologic data show that the schists near the Mount Igikpak gneiss cooled from 500 to 300 \(^\circ\text{C}\) between ca. 98 and 90 Ma, probably during development of the \(S_2\) foliation. Many \(^{40}\text{Ar}^{39}\text{Ar}\) and K-Ar ages from the Schist and Central Belts of the Brooks Range demonstrate that this cooling event was regional in nature. It also coincides with the main episode of rapid sedimentation in the Colville foreland basin (Cole et al., 1997).

It is generally assumed that the large-scale antiform of the central Brooks Range is the result of Tertiary deformation, as is the case for the Doonerak antiform near the Dalton Highway (Oldow et al., 1987; Moore et al., 1997). However, our thermochronologic analysis shows that in the Mount Igikpak region the northward tilt of the section has persisted since at least mid-Cretaceous time, while ductile deformation was taking place. Therefore, the structure on the north flank of Mount Igikpak must be interpreted as a gently north dipping extensional shear zone.

We attribute mid-Cretaceous rapid cooling to extensional denudation of the metamorphic core due to gravitational collapse of the overthickened orogenic pile. Deformation by top-to-the-north shear in the Central Belt was contemporaneous with the waiting stages of north-directed thrusting in the northern foothills. It is possible that extensional structures in the Central Belt were kinematically linked to coeval compressional structures in the foreland. This would make the mid-Cretaceous Brooks Range another example of contemporaneous and linked shortening and extension in a collisional orogen, as is the case of the Miocene to Pliocene Himalayas (Burchfiel et al., 1992).

The implications for the Bering-Chukchi region are that collisional deformation that took place in northern Alaska and Chukotka during Late Jurassic to Early Cretaceous time was immediately followed by extensional collapse of the orogen culminating ca. 90 Ma. This affected not only the Seward Peninsula.
(Dumitru et al., 1995), Chukotka (Bering Straits Geological Field Party, 1997), and the southern Brooks Range (Miller and Hudson, 1991), but also the internal part of the Brooks Range.

The youngest structures observed along our transect are large-offset north- to northwest-trending normal faults that were probably active in early Tertiary time. Normal faults of this age and orientation are also found offshore to the west of the Brooks Range, in the Hope Basin, and may be more widespread in the western Brooks Range than has been appreciated, as suggested by the large low-lying areas that are found west of Mount Igikpak. Tertiary orogeny-parallel extension may be linked to dextral slip on the Kobuk fault, related to change in the plate motions that took place in the Eocene and that affected much of interior Alaska and the Bering-Chukchi shelves.

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