Journal of Structural Geology, Vol. 10, No. 6, pp. 555 to 565, 1988
Printed in Great Britain

0191–8141/88 $03.00 + 0.00

Synmetamorphic structural evolution of the Seward Peninsula blueschist terrane, Alaska

BRIAN E. PATRICK*

Department of Geological Sciences, AJ–20, University of Washington, Seattle, WA 98195, U.S.A.

(*Received* 10 *March* 1987; *accepted in revised form* 19 *March* 1988)

Abstract—Blueschist-facies rocks of the central Seward Peninsula cropout over 8000 km^2. Protoliths were Lower Paleozoic–Precambrian(?) shallow-water miogeoclinal sediments that were metamorphosed during the Middle Jurassic. Thermobarometric estimates yield 'peak' metamorphic conditions of 10–12 kbar at 460 ± 30°C. Crystallization of blueschist-facies minerals was synkinematic with development of a transposition foliation. This foliation is parallel to lithologic contacts and is axial planar to recumbent mesoscopic isoclinal folds. These folds are refolded by larger scale recumbent tight to isoclinal folds. Both fold sets have hinges parallel to a well-developed N–S stretching lineation. Sheath folds are also present. The long axes of the sheath folds also parallel the stretching lineation. This deformation was non-coaxial as indicated by microstructures and quartz *c*-axis fabrics. Folds nucleated, then rotated into parallelism with the stretching direction. Kinematic indicators show unequivocal top-to-the-north shear sense, compatible with blueschist formation during mid-Jurassic collision between the Brooks Range continental margin and a N-facing island arc (Yukon-Koyukuk). Convergence of these two plates is believed to have been nearly N–S (in present co-ordinates).

INTRODUCTION

HIGH-pressure, low-temperature metamorphic rocks are often found in the hinterland of mountain belts and commonly record the earliest stages of continent–continent or arc–continent collision (Cannat 1985, Baird & Dewey 1986, Choukroune *et al.* 1986, Gillet *et al.* 1986, Warburton 1986). Ductile stretching fabrics and microstructures in high pressure greenschist–blueschist–eclogite-facies rocks typically indicate prolonged non-coaxial deformation on a crustal scale (Quinquis *et al.* 1978, and references cited above). These fabrics can be utilized to determine regional sense of shear and in some cases, relative plate motions (Shackleton & Ries 1984, Baird & Dewey 1986, Choukroune *et al.* 1986).

The Nome Group blueschist terrane on the Seward Peninsula is internal to the Brooks Range fold and thrust belt (Fig. 1) and shares many characteristics with the blueschists of the Western Alps (Pollock 1982, Thurston 1985, Patrick 1986). The relationship between the Seward Peninsula blueschists and the late Mesozoic Brookian orogeny, however, remains poorly understood (Mayfield *et al.* 1983). Many workers have postulated links between the schist belt of the southern Brooks Range and the Seward Peninsula blueschists on the basis of metamorphic grade, lithological similarity and protolith ages (Forbes *et al.* 1981, 1984, Mayfield *et al.* 1983, Thurston 1985). The unusual geographic position of the Seward Peninsula blueschists with respect to the Brooks Range led Patton & Tailleur (1977) to suggest that an oroclinal bend ('the Chukchi syntaxis') had formed due to differential movement of North America and Asia in the Tertiary.

This paper will focus on the synmetamorphic structural history of the central Seward Peninsula, with particular emphasis on fabric development, regional patterns of stretching lineations, quartz *c*-axis fabrics and other microstructural indicators of shear sense. These data will be examined within the model framework of a crustal scale shear zone. In addition, the relationship between the Seward Peninsula blueschists and the tectonic evolution of northern Alaska will be explored.

GEOLOGIC SETTING

The Nome Group is an areally extensive sequence of earliest Cambrian (Precambrian?) to Devonian marbles, pelitic schists, chloritic schists and graphitic schists that were intruded by what are now metabasites and granitic orthogneisses sometime in the mid-Paleozoic. Regional mapping by Till *et al.* (1986) outlined a coherent metamorphic stratigraphy that has a present minimum thickness of 4.5 km and can be traced throughout much of the eastern and central Seward Peninsula. The top and base of the Nome Group are not exposed, so the original thickness of the sequence is unknown.

The depositional environment for the largely metasedimentary Nome Group is believed to have been shallow water shelf and slope facies (Till *et al.* 1986), possibly representative of a passive continental margin. Similar lithologies are found throughout the western Brooks Range schist belt (Armstrong *et al.* 1986). Devonian granitic rocks in the southern Brooks Range (Dillon *et al.* 1980) may extend westward to the Kiwalik orthogneiss in northeast Seward Peninsula (Till 1983).

Along with the low-grade metamorphic rocks on the Seward Peninsula, amphibolite- to granulite-facies rocks

* Present address: Department of Geology and Geophysics, University of Calgary, Calgary, Alberta, Canada T2N 1N4.

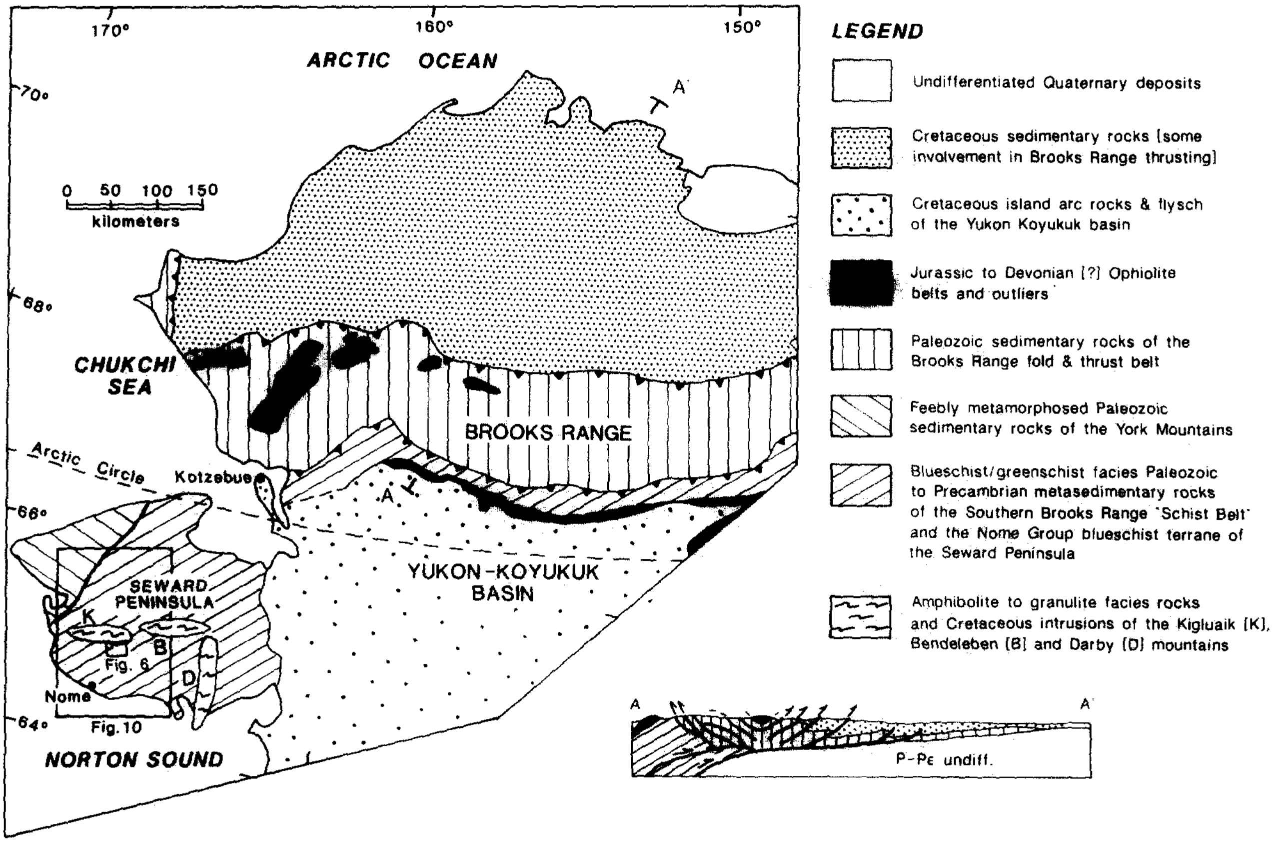

Fig. 1. Generalized geological map and cross-section of northwestern Alaska showing north-vergent Brooks Range fold and thrust belt and associated metamorphic hinterland. Cross-section after Till (unpublished). Thrust faults shown with barbs on upper plate.

are exposed in the Kigluaik, Bendeleben and Darby Mountains (Fig. 1). These rocks possess similar structures and lithologies to those found in the Nome Group blueschists (Patrick 1986, Lieberman 1986).

In the northwestern Seward Peninsula, the Nome Group blueschist terrane is juxtaposed against feebly metamorphosed rocks of the York Mountains (Fig. 1), along a poorly exposed fault of undetermined displacement and nature.

METAMORPHISM

The blueschists of northern Alaska were cited by Sainsbury *et al.* (1970) and subsequent workers as examples of Precambrian high-pressure metamorphism. However, a recent Rb–Sr and K–Ar study by Armstrong *et al.* (1986) on whole rock-mineral (phengite, amphibole and paragonite) pairs points to a middle to late Jurassic blueschist-forming event.

Blueschist-facies parageneses are widespread on the Seward Peninsula, covering an area greater than 7000 km^2 (Fig. 2). Diagnostic assemblages include: quartz +

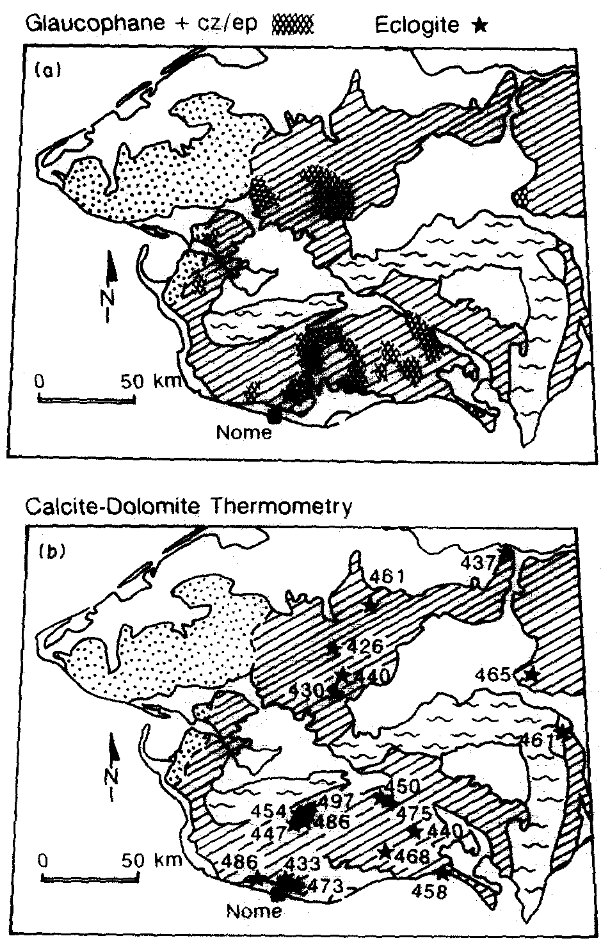

Fig. 2. (a) Distribution of glaucophane + clinozoisite in mafic and pelitic rocks and eclogite localities of the Seward Peninsula blueschist terrane. Patterns are the same as for Fig. 1. (b) Metamorphic temperatures obtained from calcite–dolomite mineral pairs utilizing the equation of Rice (1977). Application of a 10 kb pressure correction based on the work of Goldsmith & Newton (1969) systematically lowers all temperatures by ~40°C (Patrick *et al.* 1985).

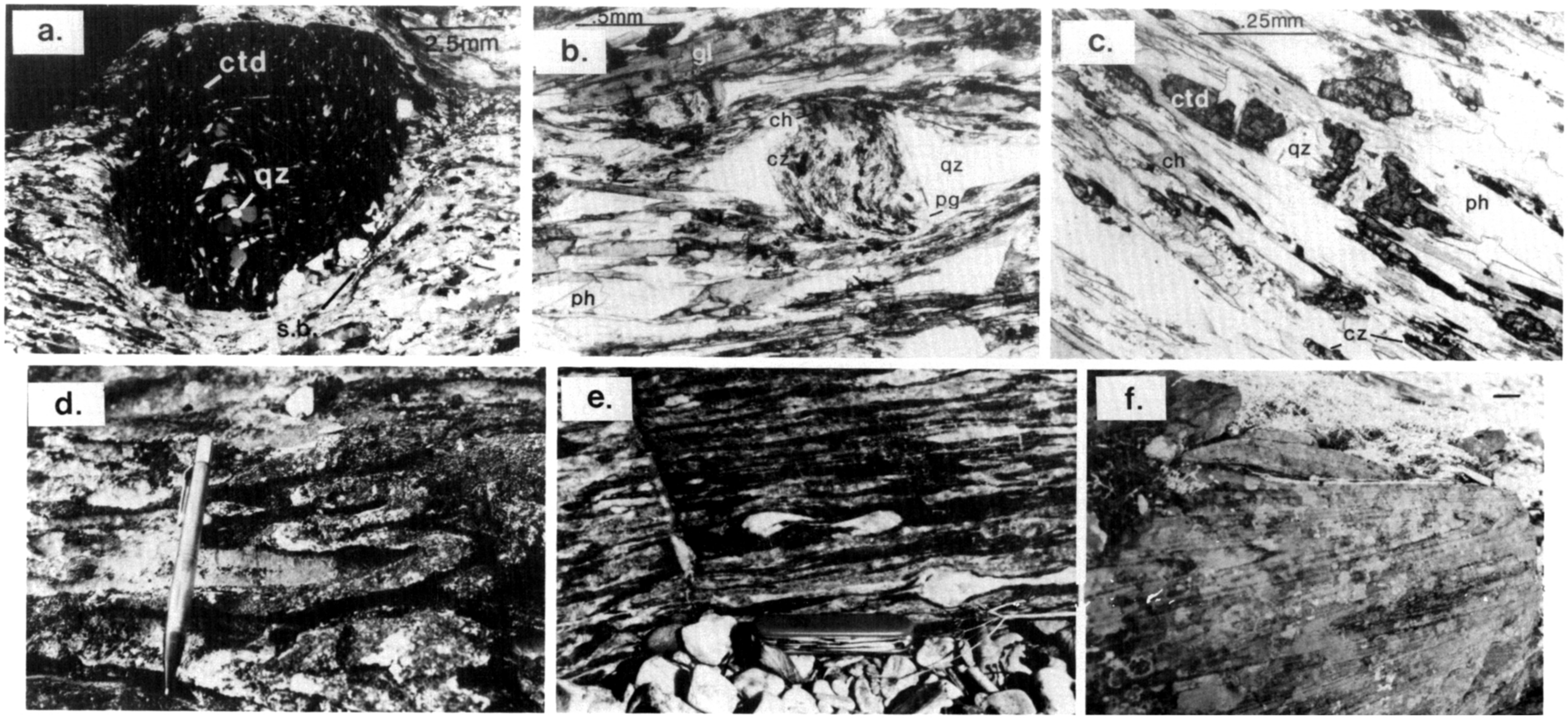

Fig. 3. Direct and indirect evidence of synkinematic blueschist metamorphism and mesoscopic structures in the Seward Peninsula blueschist terrane. (a) Snowball garnet in sample SL81-286-2. Inclusion trails are composed of quartz (qz) and chloritoid (ctd). Note post-metamorphic shear band (s.b.) indicating dextral shear while garnet rotation indicates sinistral shear. (b) Helicitic lawsonite pseudomorph in matrix of pelitic schist (sample AB80-15-5), composed of clinozoisite (cz), chlorite (ch) and paragonite (pg). Quartz (qz) in strain shadow, glaucophane (gl) and phengite (ph) are parallel to the foliation. (c) Microboudinage of chloritoid grain in glaucophane-bearing pelitic schist (sample AB80-58-5). Note the ubiquitous parallelism of all prismatic minerals with the foliation. cz—clinozoisite, ctd—chloritoid, ph—phengite, ch—chlorite. (d) Rootless intrafolial fold defined by quartz layer in pelitic schist. Axial plane is parallel to transposition foliation. (e) Rootless isoclinal folds in banded marble. Note the opposite sense of vergence of these folds (photo by A. B. Till). (f) Recumbent isoclinal folds in micaceous marble, interlayered with transposed pelitic schists.

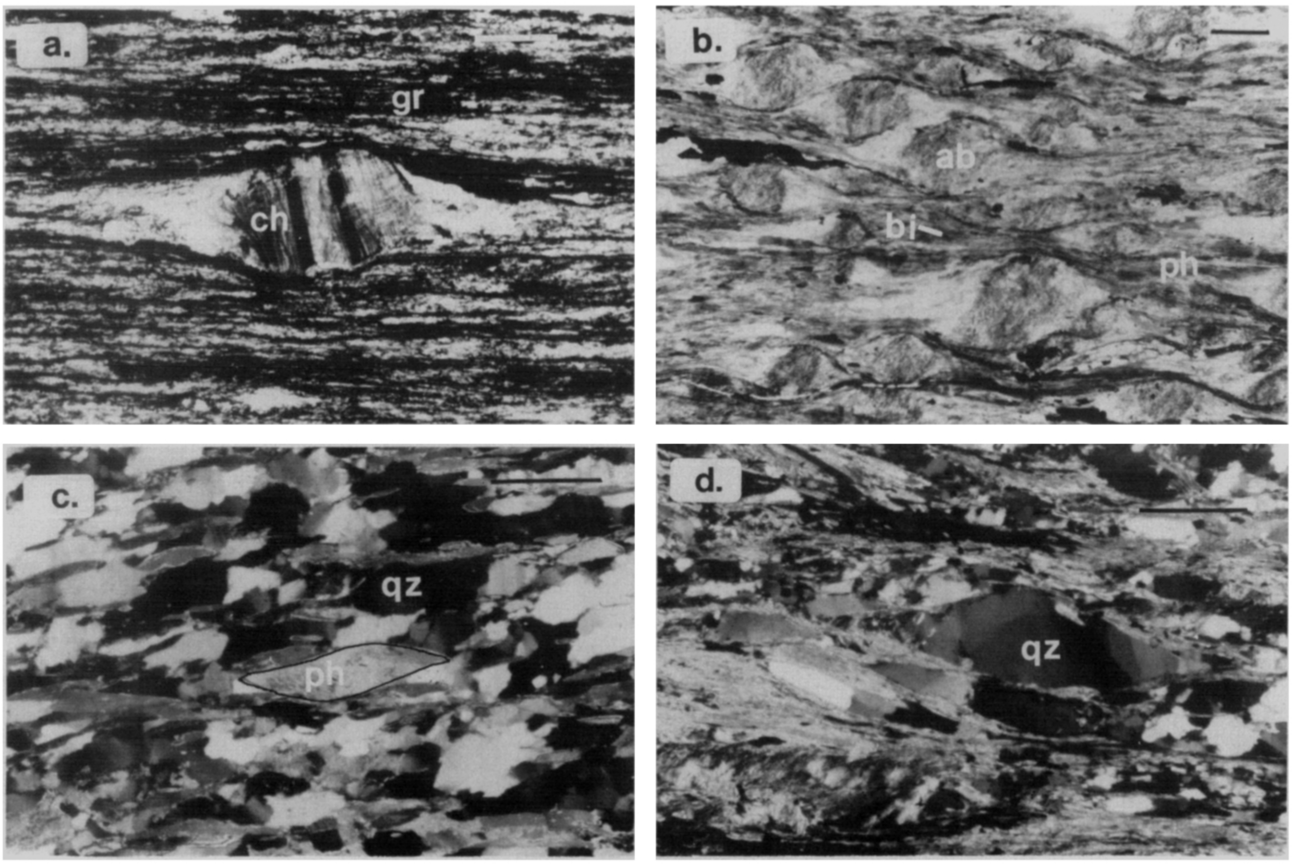

Fig. 4. Microscopic sense of shear indicators in the Nome Group blueschists. (a) Slightly asymmetric strain shadow around boudinaged chlorite (ch) grain in quartz–graphite (gr) schist indicating dextral shear. (b) Asymmetric strain shadows around albite (ab) in psammitic schist with phengite (ph) and static-mimetic biotite (bi) indicating dextral shear (sample AB85-15). (c) Phengite 'fish' in granitic orthogneiss showing dextral shear (sample AB84-194). (d) Sigmoidal flattened quartz aggregates showing sinistral shear. All sense of shear indicators shown here indicate top-to-the-north shear sense. Scale bar in all photomicrographs is 1 mm.

phengite + glaucophane ± chloritoid ± garnet ± clinozoisite ± paragonite in pelitic lithologies and glaucophane + garnet +epidote + chlorite ± Na-actinolite ± phengite ± quartz in metabasites. Interlayered eclogites have been found in three localities and are thought to be isofacial with the blueschists (Forbes *et al.* 1984, Thurston 1985). Lawsonite pseudomorphs are common in both pelitic and mafic rocks, signifying a progradation from lawsonite- to epidote-zone blueschists (Forbes *et al.* 1984, Thurston 1985).

Nearly all blueschist-facies parageneses show partial post-kinematic overgrowth of albite–epidote–amphibolite and greenschist-facies minerals: barroisitic rims on glaucophanitic amphibole, pseudomorphic replacement of glaucophane by albite + chlorite ± oxychlorite or actinolite, breakdown of fairly magnesian chloritoid to chlorite + paragontite and incipient replacement of high-Si phengite by biotite are but some examples. This overprint is believed to be a consequence of decompression along a clockwise *P–T* path (Forbes *et al.* 1984, Thurston 1985, Evans & Patrick 1987).

In addition to the greenschist overprint, a superimposed thermal overprint has been observed on the south flank of the Kigluaik Mountains. This is manifested by static-mimetic growth of biotite in pelitic schists and replacement of chloritoid by margarite immediately south of the Kigluaik Mountains. This 'Lepontine-style' overprint on the blueschists has also been observed in the northern Darby Mountains (Till *et al.*, 1986).

Metamorphic conditions have been determined utilizing calcite–dolomite, garnet–biotite, garnet–clinopyroxene and two-feldspar thermometry with jadeite–albite–quartz and phengite–biotite–K-feldspar barometry (Pollock 1982, Patrick *et al.* 1985, Thurston 1985, Evans & Patrick 1987, Patrick 1987). Blueschist-facies conditions are estimated to have been 10–12 kbar at 460 ± 30°C followed by decompression to <5 kb (Evans & Patrick 1987).

Regional calcite–dolomite thermometry by Patrick *et al.* (1985) shows a remarkable uniformity in metamorphic temperatures over much of the central Seward Peninsula (Fig. 2). This is in contrast to other coherent blueschist terranes such as the Western Alps and New Caledonia where mineral isograds and changes in metamorphic grade can be readily recognized (cf. Frey *et al.* 1974, Yokoyama *et al.* 1986).

Evidence that blueschist metamorphism was synkinematic is provided by rare snowball garnets (Fig. 3). Garnets commonly contain inclusions of chloritoid, glaucophane and lawsonite pseudomorphs. Rare lawsonite pseudomorphs in the matrix, containing internal *S*-surfaces, broadly constrain the deformation to be synkinematic with the early lawsonite-zone blueschist metamorphism (Fig. 3). The relative scarcity of lawsonite pseudomorphs in the matrix vs those found within garnet, however, indicates that deformation in most areas continued after the stability limit of lawsonite was exceeded.

Indirect evidence for pre- to synkinematic metamorphism is the parallelism of all prismatic minerals (i.e. glaucophane, chlorite, phengite, chloritoid, epidote) with a well developed schistosity (Fig. 3), asymmetric strain shadows around equant minerals (albite, garnet, chloritoid) and isoclinally folded phengite, glaucophane and chloritoid (see also Thurston 1985).

STRUCTURAL EVOLUTION

The following observations are based largely on 2 inch/mile mapping in the region of Salmon Lake (Fig. 6). Regional reconnaissance mapping by Till *et al.* (1986) and myself is consistent with the detailed work.

Fabrics

The dominant fabric in the Nome Group is a shallowly dipping foliation–schistosity (S_1). This foliation is everywhere subparallel to lithologic contacts, giving the initial impression of 'layer-cake' stratigraphy. Foliation is developed in all lithologies with the exception of deformed layers and boudins of metabasite and large orthogneiss bodies that retain relict igneous textures partially replaced by blueschist–greenschist minerals.

Individual lithological units can be traced for many kilometers along strike, without visible large-scale overturning of stratigraphy (Fig. 6). Exposure of the Nome Group on the central Seward Peninsula, however, is generally fair to poor, making large-scale structures difficult to identify.

On the outcrop scale, the foliation is axial planar to recumbent isoclinal folds. In pelitic schist units, quartz segregations define rootless intrafolial folds with an axial-planar transposition foliation (Fig. 4d & e). In other lithologies lacking suitable markers or ductility contrasts (e.g. marbles, chlorite–albite schists), transposition is not readily seen, but the foliation is nonetheless axial planar to recumbent isoclinal folds (Fig. 4f). Measured fold axes of all small-scale folds form a well-defined point maximum (Fig. 7) within an incipient great circle. This dispersal of fold axes (Fig. 7) is due to postmetamorphic open folding about an E–W axis (the Mount Distin Synform of Fig. 6). Indicators of consistent vergence in recumbent isoclinal folds are lacking on the micro- and mesoscopic scales throughout the central Seward Peninsula (cf. Fig. 4e).

Sheath folds are occasionally observed in semi-pelitic and quartzo-feldspathic schists, with the long axes of the sheath folds oriented parallel to the hinges of recumbent isoclinal folds (Fig. 5).

Recumbent isoclinal folds in pelitic schists and chlorite–albite schists are refolded *coaxially* by larger scale (to outcrop size) tight to isoclinal folds (Fig. 8). These structures fold the foliation (S_1) that is axial planar to the smaller scale isoclinal folds, but the new foliation (S_2) remains axial planar to both fold sets, except locally, where it is folded in the later fold hinges. Thus, although these large folds clearly refold the smaller scale folds, both sets are recumbent, nearly isoclinal and possess the same axial orientation. These superimposed fold sets

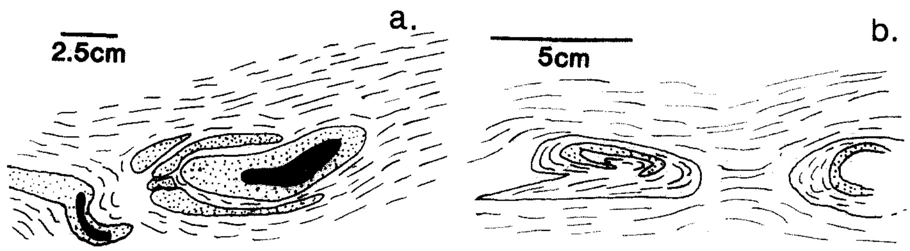

Fig. 5. Line drawings from photographs of sheath folds. (a) Impure marble; stippled pattern is calcite marble, dashes and solid pattern are micaceous marble. (b) Quartzo-feldspathic schist; stippled pattern is quartz layer, dashes are more pelitic material.

give rise to Type 3 interference patterns as defined by Ramsay (1967; see also Fig. 8).

All tight to isoclinal fold hinges throughout the central Seward Peninsula and especially in the Salmon Lake region are parallel to a well-developed stretching lineation oriented generally N–S (Fig. 7). This lineation is defined by quartz rodding in quartzites and pelitic schists, boudins formed from quartz segregations, mica streaking and microcrystalline boudinage of clinozoisite, chloritoid and glaucophane in pelitic schists. In some lithologies lacking ductility contrasts and banding (e.g. quartzites and some quartz graphite schists) the rocks appear to be L-tectonites.

The ubiquitous parallelism between fold hinges and stretching lineations raises an old structural question: was the main axis of shortening oriented parallel or perpendicular to the fold hinges? Previous workers on the Seward Peninsula (Sainsbury 1972, Pollock 1982) have used N–S-trending fold hinges as evidence of E–W shortening, although Thurston (1985) allowed for the possibility of N–S shortening.

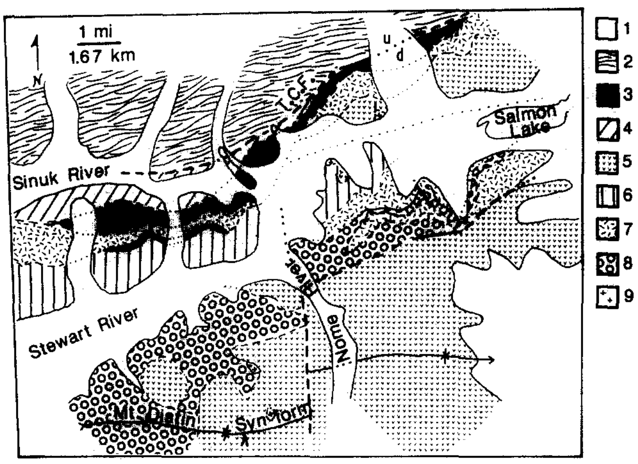

Fig. 6. General geologic map of the Salmon Lake region (see Fig. 1 for location). Mapping east of Nome River by Pollock (1982) with modifications based on fieldwork by the author and the metamorphic stratigraphy outlined by Till *et al.* (1986). 1—Undifferentiated Quaternary deposits. 2—Staurolite to sillimanite grade rocks, separated from the Nome Group (3–9) by the Tumit Creek Fault (T.C.F.). 3—Quartz–graphite and graphite schists and phyllites. 4—Biotite–plagioclase schists. 5—Chlorite–albite schists and other mafic metasediments, commonly containing mafic blueschist boudins and rare eclogite boudins. 6—Pelitic schists with phengite–quartz–albite ± chloritoid. East of Nome River, glaucophane ± garnet are locally abundant. 7—Calcareous (calcite + ankerite) psammitic schists. 8—Grey calcite marbles, locally micaceous. 9—Granitic orthogneisses.

For several reasons, I favor an alternative interpretation. The strongly developed linear fabric, the presence of rare sheath folds and the recumbent isoclinal, typically rootless, intrafolial style of folding together suggest a deformation history in which folds are tightened, attenuated and rotated into parallelism with the stretching direction of the finite-strain ellipsoid (interpreted to be equivalent to the stretching lineation).

This behavior of linear elements under progressive noncoaxial deformation (approximately simple shear) has been demonstrated theoretically (Escher & Watterson 1974, Cobbold & Quinquis 1980), experimentally (Cobbold & Quinquis 1980), and has been described in numerous field studies (Bryant & Reed 1969, Bell 1978, Quinquis *et al.* 1978, Williams 1978, Minnigh 1979, Meneilly & Storey 1986).

Since all isoclinal fold hinges are now parallel or subparallel to the stretching lineation, it is difficult to say whether they formed as active folds contemporary with deformation (as per Williams 1978) or as passive amplification of initial inhomogeneities (e.g. sedimentary structures) in the sedimentary pile (as per Cobbold & Quinquis 1980). Cobbold & Quinquis (1980) contended that passive folds tend to be more non-cylindrical (i.e. sheath-like) than active folds. If this is the case, the relative rarity of sheath folds and other non-cylindrical structures on the central Seward Peninsula indicates that folding during the early stages of deformation was active, that is, it was due to rheological differences rather than to amplification of pre-existing inhomogeneities.

If the early recumbent isoclinal folds were rotated into parallelism with the stretching direction, how were they coaxially refolded by later recumbent isoclinal folds? Three hypotheses come to mind: (1) a later defor-

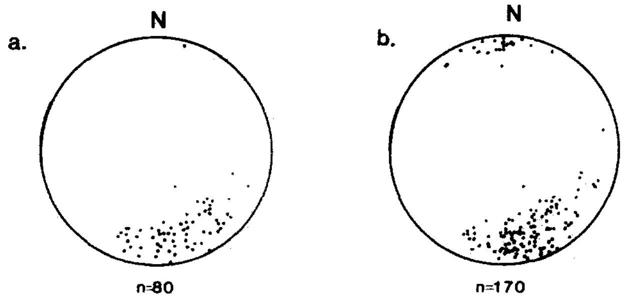

Fig. 7. Equal-area projections of representative isoclinal fold axis (a) and stretching lineation (b) orientations from the western half of Fig. 6.

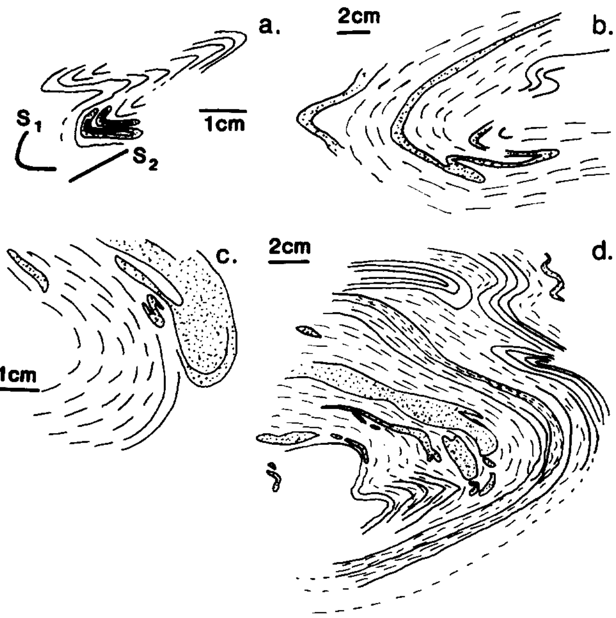

Fig. 8. Examples of isoclinally refolded isoclinal folds from the region shown in Fig. 6. Stippled pattern: quartz segregations, (a)–(c) are line drawings from photographs, (d) is a tracing of oriented slab.

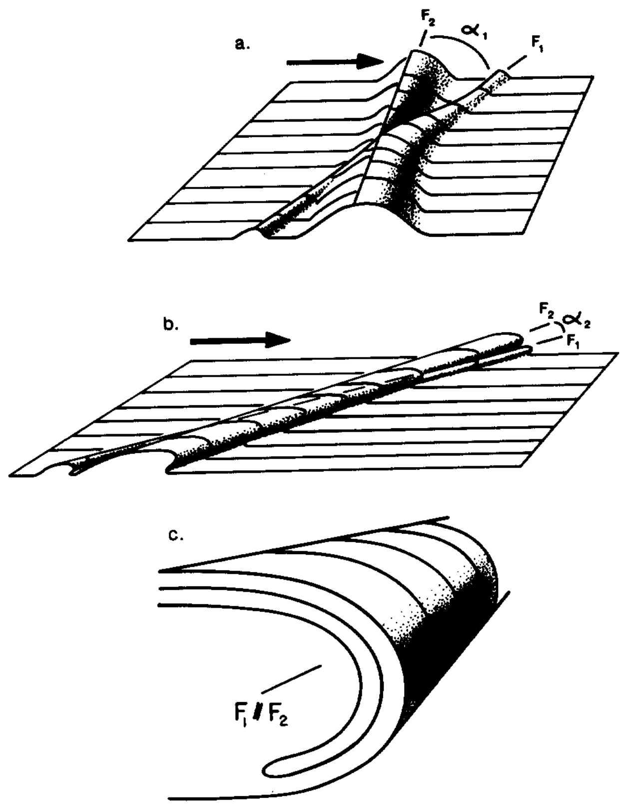

Fig. 9. (a) and (b) Schematic development of refolded folds during progressive simple shear. α is the angle between F_1 and F_2 which becomes smaller as folds rotate towards the shear direction (shown by bold arrow). (c) After high shear strains, both axes coincide. See text for discussion.

mational event occurred with the main axis of shortening oriented at right angles to that of the early event, (2) linear anisotropy was introduced during the first deformation which controlled the orientation of later folding, (3) both fold sets formed during the same deformation, with refolding and reorientation occurring simultaneously.

Hypothesis (1) would require a fortuitous change in the shortening direction throughout the Seward Peninsula. Late Cretaceous E-vergent thrust-faulting has been documented further north in the Chukchi Sea and Lisburne Peninsula (Grantz *et al.* 1976, Fisher *et al.* 1982) and on the eastern shore of Norton Basin (Patton & Tailleur 1977), but there is little evidence for ductile deformation associated with these fault zones elsewhere in northern Alaska.

Hypothesis (2) was demonstrated to be viable by Cobbold & Watkinson (1981) and Watkinson & Cobbold (1981). They showed that fold axes would tend to form parallel to linear elements somewhat independently of the imposed stress field *if* the linear element was much stronger rheologically than the matrix. They used the example of a phyllite with quartz rods, which is a reasonable analogue to the pelitic schists of the Seward Peninsula. If the main axis of shortening for the refolding phase were at a high angle to foliation (thought to be roughly equivalent to the early shear plane), it would be possible to generate coaxial recumbent isoclinal folds that refolded the earlier structures. This could correspond to a nearly coaxial flattening caused by uplift of the blueschist terrane superimposed upon the earlier non-coaxial deformation.

The problem with this scenario is the requirement that the quartz rods be *much* stronger than the pelitic or feldspar–mica matrix. This is clearly true at low temperatures, but at temperatures of 450–500°C and pressures of ~10 kb, quartz readily deforms by intracrystalline slip (see section on quartz fabrics). Thus, at blueschist–greenschist-facies conditions, the rocks were probably not that rheologically anisotropic.

The third hypothesis allows for the development of two fold sets coincident in style and orientation during a monocyclic progressive non-coaxial deformation. Figure 9 shows the schematic development of isoclinal refolding of isoclinal folds with progressive simple shear. The first fold is nucleated and begins to attenuate, tighten and rotate towards the stretching direction. The second fold then nucleates, refolding the first (Fig. 9a). Both folds then continue to rotate into parallelism with the stretching direction which is itself rotating toward the shear plane. The original angular relationship between the fold sets is lost during progressive deformation. This evolution can be more precisely stated as:

$$\lim_{X \to S} X = F_1 = F_2 \qquad (\alpha \to 0) \qquad \text{(see Fig. 9).}$$

Shear sense

The deformation scenario outlined above coupled with the orientation of stretching lineations throughout the central Seward Peninsula (Fig. 10) constrains the regional shear direction to be either northwards or

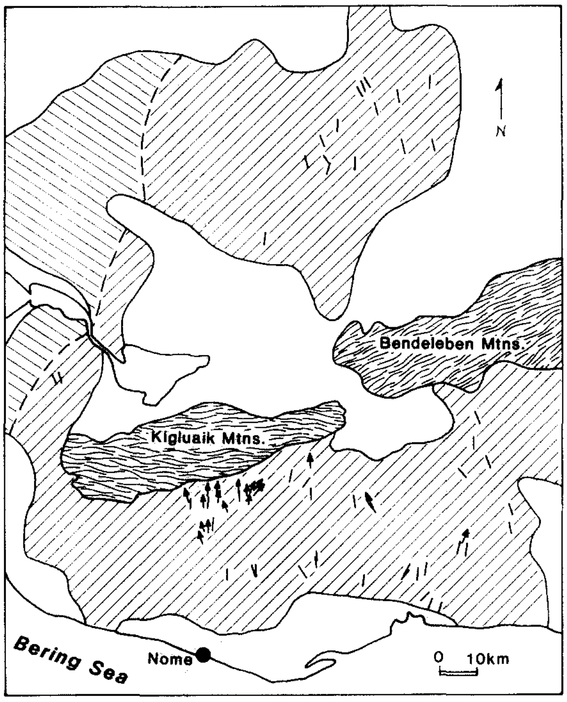

Fig. 10. Stretching lineation map for the central Seward Peninsula (see Fig. 1 for location). Arrows indicate shear sense as deduced from microstructures and quartz c-axes fabrics. Patterns are the same as for Fig. 1.

southwards. Out of a suite of 25 oriented thin sections cut perpendicular to foliation and parallel to the stretching lineation, 18 show unequivocal evidence of top-to-the-north sense of shear. Indicators of shear sense include snowball garnets, asymmetric strain shadows around garnet and albite (both σ and δ type as defined by Passchier and Simpson 1986), S–C fabrics, preferred orientation of flattened, elongate quartz aggregates and sigmoidal phengite ('mica fish') (Figs. 3 and 4). Of the six samples that show no consistent shear sense, a few show conjugate shear bands that may indicate postmetamorphic coaxial flattening (Platt & Vissers 1980). One sample (AB86–9) showed unequivocal top-to-the-south shear. This was determined on the basis of postmetamorphic shear bands and preferred orientation of elongate flattened quartz aggregates. The reason for this discrepancy in shear sense is not clear. One possibility is that these postmetamorphic structures are related to normal movement along the Tumit Creek Fault (Fig. 6).

Examination of thin sections cut perpendicular to the stretching lineation and foliation show little evidence of monoclinic fabrics and no consistent shear sense.

Quartz fabrics

In addition to analysis of microscopic shear indicators, five samples of quartzite were selected for quartz c-axis fabric study. c-Axes were measured by standard universal-stage techniques.

Pure quartzites are rare in the metasedimentary pile on the Seward Peninsula and are generally quite thin. Care was taken only to sample monomineralic layers (>95% quartz) so as to minimize the possible heterogeneous reorienting effects on other minerals (i.e. white mica and feldspar). Samples were taken largely from the centers of layers >15 cm in thickness. It is thought that the thicker layers are more likely to have developed fabrics reflective of the bulk strain rather than some local strain field.

The results of the quartz fabric measurements are shown in Fig. 11. Four of the five samples show a well

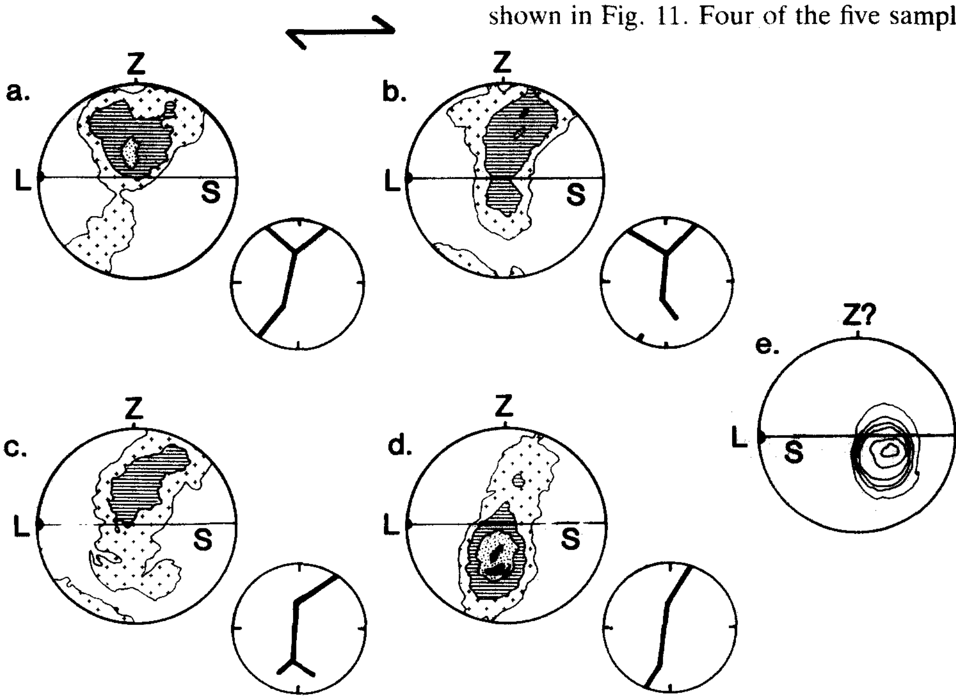

Fig. 11. Quartz c-axis plots and associated fabric 'skeletons' for samples cut parallel to the stretching lineation (L) and perpendicular to foliation (S). Z is the approximate inferred compression direction of the finite-strain ellipsoid. Stretching lineation is inferred to represent the shortening direction of the finite strain (X). (a)–(d) Plots all indicate top-to-the-north sense of shear. n = 100 for all plots; contour intervals are 1, 2, 4 and 6% of points per 1% area for (a)–(d) and 0.5, 2, 4, 6, 10 and 11% of points per 1% area for (e). See text for further discussion.

developed monoclinic fabric. The fabric 'skeletons' (Fig. 11) resemble the type-I crossed-girdles developed in the computer models of Lister & Hobbs (1980), with the exception that one of the arms is poorly developed in some of the samples, see for comparison the fabrics of Bouchez & Pêcher (1976) and in fig. 12 of Lister & Williams (1979).

Asymmetric quartz *c*-axis fabrics, coupled with the evidence presented above for one episode of non-coaxial deformation, allows determination of shear sense (cf. Lister & Hobbs 1980). The fabrics shown in Fig. 11 (a)–(d) are consistent with the other microstructural indicators in that they demonstrate top-to-the-north sense of shear.

One of the quartz *c*-axis fabrics obtained does not show a well developed asymmetric pattern (Fig. 11e). Instead it shows a well developed maximum centred well away from the inferred shortening direction. This could represent a coaxial flattening fabric with a strongly developed single maximum around Z as shown in the simulations of Lister & Hobbs (1980), although the discrepancy between the inferred shortening direction and the position of the maximum is not well understood. This *c*-axis pattern illustrates the danger in relying solely on crystallographic fabrics to determine shear sense and bulk deformation.

REGIONAL TECTONICS

The regional pattern of stretching lineations and shear sense indicates blueschist formation took place within a top-to-the-north shear zone of crustal dimensions (8000 km^2 area by >4.5 km depth). High pressure (~12 kbar) parageneses imposed on a coherent sequence of miogeoclinal sediments points to metamorphism related to collisional tectonics (A-type subduction as defined by Bally 1975).

The above fits in well with the tectonic model for evolution of the Brooks Range first proposed by Roeder & Mull (1978) and recently greatly revised and expanded by Box (1985). In the model proposed by Box (1985), a N-facing island arc (the Yukon-Koyukuk) collided with a rifted continental margin. Thrusting associated with the collision started in late Jurassic in the western Brooks Range (Mayfield *et al.* 1983) and occurred progressively later to the east. The earliest phase of thrusting coincides rather well with the mid-Jurassic radiometric ages for the blueschist metamorphism on the Seward Peninsula as determined by Armstrong *et al.* (1986).

A possible scenario for blueschist development on the Seward Peninsula is shown in Fig. 12. Here the Seward Peninsula blueshists formed as a direct consequence of the collision of the Yukon-Koyukuk basin with the northern Alaska continental margin. The crustal-scale shear zone responsible for the development of fabrics in the Nome Group blueschists changes character at higher crustal levels, where it becomes the early-formed portion of the thin-skinned fold and thrust belt of the Brooks Range.

Stretching fabrics and plate motions

The relationship between stretching lineations and relative plate motion was examined in a paper by Shackleton & Ries (1984). A number of recent authors

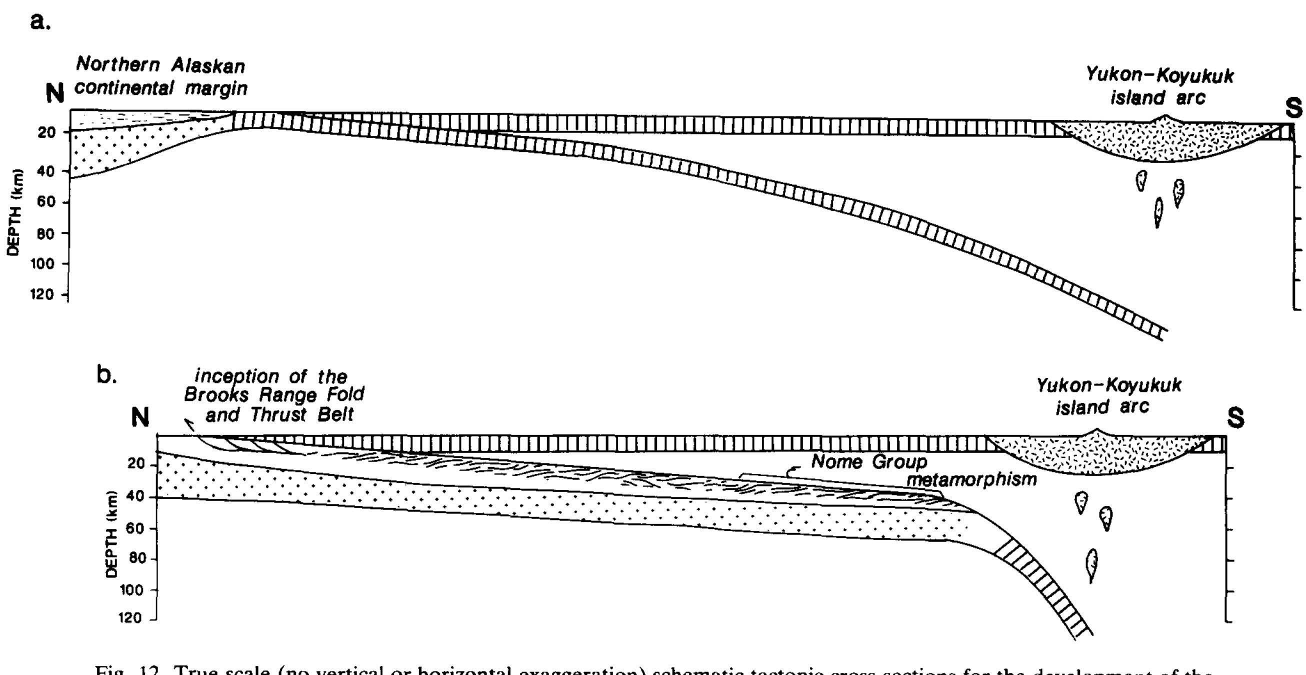

Fig. 12. True scale (no vertical or horizontal exaggeration) schematic tectonic cross-sections for the development of the Nome Group blueschist terrane and the inception of the Brooks Range fold and thrust belt. (a) Pre-collisional cross-section. Crosses—continental crust; dashes—Paleozoic passive margin sedimentary sequence; vertical lines—oceanic crust; hatched pattern—Yukon-Koyukuk island arc. (b) Middle Jurassic (*ca* 140–165 Ma), early collisional phase of the Brookian orogeny. The wavy pattern corresponds to a crustal-scale shear zone of which the Nome Group blueschists (N.G.) are but a portion. This shear zone merges upward to form the earliest formed portion of the thin-skinned Brooks Range fold and thrust belt. See text for discussion.

(Baird & Dewey 1986, Brunel 1986, Choukroune *et al.* 1986, Brown 1986) have compared the orientation of stretching lineations with known or inferred plate directions with varying degrees of success. In the Western Alps, for example, the metamorphic fabrics related to eoalpine eclogite-facies metamorphism may be coincident with northward convergence of the Apulian plate relative to the European plate (Baird & Dewey 1986, Choukroune *et al.* 1986). Later metamorphic fabrics represent nappe emplacement direction and should not be used to infer plate motions.

It could be the case that in collisional orogenies, the earliest formed fabrics, developed as one plate begins to underthrust another, reflect the original relative plate motions. These fabrics may only be preserved in the high-pressure metamorphic rocks formed synkinematically. Fabrics associated with later, lower P–T conditions (i.e. Barrovian sequences) may not then represent relative plate motions, but instead be indicative of local thrusting directions (cf. Brunel 1986).

In northern Alaska, Tertiary remagnetization in the Yukon-Koyukuk basin has all but obliterated earlier paleomagnetic information (Harris 1985), so the relative plate motion between the Yukon-Koyukuk island arc and the Brooks range is not known. The convergence direction between these plates, as indicated by the regionally consistent stretching lineations on the Seward Peninsula was nearly due north. In the absence of paleomagnetic confirmation, stretching lineations from early-formed blueschist-facies rocks could yield a first approximation as to relative plate motions. Without independent confirmation, however, it is dangerous to deduce plate motions based solely on the orientation of stretching lineations.

CONCLUSIONS

The Seward Peninsula blueschist terrane represents a former passive continental margin that was underthrust as a coherent package beneath an advancing N-facing island arc in the middle to late Jurassic. Geobarometry suggests depths of burial on the order of 35–40 km. Synmetamorphic fabrics developed in the Nome Group protoliths are remarkably consistent over much of the central Seward Peninsula and are consistent with their development in a crustal-scale ductile shear zone.

The parallels between these blueschists and the eoalpine blueschists–eclogites of Western Europe are striking. Since both blueschist terranes are inferred to have formed in a collisional orogeny, this result is not particularly surprising. The structural evolution associated with the earliest orogenic phase provide us with a generalized model with which to evaluate the early tectonometamorphic history of other mountain belts where this early phase is cryptic.

Acknowledgements—This research represents a portion of the authors Ph.D. dissertation at the University of Washington. Support was provided by N.S.F. grants EAR 8218471 and 8507757 to B. W. Evans, a G.S.A. Grant-in-aid of graduate research and the Corporation Fund of the University of Washington. I am indebted to A. B. Till of the U.S.G.S. for her logistical assistance, S. P. Thurston for discussion of Nome Group structures and J. E. Lieberman for his assistance in the field. An early version of this manuscript was reviewed by A. B. Till, D. S. Cowan and B. W. Evans. The manuscript was improved through reviews by P. Choukroune, P. J. Hudleston and an anonymous reviewer.

REFERENCES

Armstrong, R. L., Harakal, J. E., Forbes, R. B., Evans, B. W. & Thurston, S. P. 1986. Rb–Sr and K–Ar study of metamorphic rocks of the Seward Peninsula and Southern Brooks Range, Alaska. In: *Blueschists and Eclogites* (edited by Evans, B. W. & Brown, E. H.). *Mem. geol. Soc. Am.* **164**, 185–203.

Bally, A. W. 1975. A geodynamic scenario for hydrocarbon occurrences. *Proc. Ninth World Petrol. Congr., Tokyo*, Vol. 2. Applied Science, Essex, 33–44.

Baird, A. W. & Dewey, J. F. 1986. Structural evolution in thrust belts and relative plate motion: the upper Pennine Piemont zone of the Internal Alps, southwest Switzerland and northwest Italy. *Tectonics* **5**, 375–387.

Bell, T. H. 1978. Progressive deformation and reorientation of fold axes in a ductile mylonite zone: the Woodroffe Thrust. *Tectonophysics* **44**, 285–320.

Bouchez, J.-L. & Pêcher, A. 1976. Plasticité de quartz et sens de cisaillement dans les quartzites du Grand Chevauchement Central Himalayen. *Bull. Soc. géol. Fr.* **6**, 1375–1384.

Box, S. E. 1985. Early Cretaceous orogenic belt in northwestern Alaska: internal organization, lateral extent, and tectonic interpretation. In: *Tectonostratigraphic Terranes of the Circumpacific Region* (edited by Howell, D. G.). *Circumpacific Council for Energy and Mineral Resources, Earth Sci. Ser.* **1**, 137–146.

Brown, E. H. 1986. Geology of the Shuksan Suite, North Cascades, Washington, U.S.A, In: *Blueschists and Eclogites* (edited by Evans, B. W. & Brown, E. H.). *Mem. geol. Soc. Am.* **164**, 143–154.

Brunel, M. 1986. Ductile thrusting in the Himalayas: shear sense criteria and stretching lineations. *Tectonics* **5**, 247–265.

Bryant, B. & Reed, J. C. Jr. 1969. Significance of lineation and minor folds near major thrust faults in the southern Appalachians and the British and Norwegian Caledonides. *Geol. Mag.* **106**, 412–429.

Cannat, M. 1985. Quartz microstructures and fabrics in the Island of Groix (Brittany, France). *J. Struct. Geol.* **7**, 555–562.

Choukroune, P., Ballevre, M., Cobbold, P., Gautier, Y., Olivier, M. & Vuichard, J.-P. 1986. Deformation and motion in the western Alpine arc. *Tectonics* **5**, 215–226.

Cobbold, P. R. & Quinquis, H. 1980. Development of sheath folds in shear regimes. *J. Struct. Geol.* **2**, 119–126.

Cobbold, P. R. & Watkinson, A. J. 1981. Bending anisotropy: a mechanical constraint on the orientation of fold axes in an isotropic medium. *Tectonophysics* **72**, T1–T10.

Dillon, J. T., Pessel, G. H., Chen, J. H. & Veach, N. C. 1980. Middle Paleozoic magmatism and orogenesis in the Brooks Range, Alaska. *Geology* **4**, 338–343.

Escher, A. & Watterson, J. 1974. Stretching fabrics, folds and crustal shortening. *Tectonophysics* **22**, 223–231.

Evans, B. W. & Patrick, B. E. 1987. Phengite (3T) in high-pressure metamorphosed granitic orthogneisses, Seward Peninsula, Alaska. *Can. Mineralogist* **25**, 141–158.

Fisher, M. A., Patton, W. W. Jr. & Holmes, M. L. 1982. Geology of Norton Basin and continental shelf beneath northwestern Bering Sea. *Bull. Am. Ass. Petrol. Geol.* **66**, 255–285.

Forbes, R. B., Evans, B. W. & Pollock, S. 1981. The Nome Group blueschist terrane: a possible extension of the Brooks Range schist belt. *Geol. Soc. Am. Abs. w. Prog.* **13**, 56.

Forbes, R. B., Evans, B. W. & Thurston, S. P. 1984. Regional progressive high-pressure metamorphism, Seward Peninsula, Alaska. *J. Met. Geol.* **2**, 43–54.

Frey, M., Hunziger, J. C., Frank, W., Bocquet, J., Dal Piaz, G. V., Jäger, E. & Niggli, E. 1974. Alpine metamorphism of the Alps: a review. *Schweiz. miner. petrogr. Mitt.* **54**, 247–290.

Gillet, Ph., Choukroune, P., Balleure, M. & Davy, Ph. 1986. Thickening history of the Western Alps. *Earth Planet. Sci. Lett.* **78**, 44–52.

Goldsmith, J. R. & Newton, R. C. 1969. P–T–X relations in the system $CaCO_3$–$MgCO_3$ at high temperatures and pressures. *Am. J. Sci.* **267A**, 160–190.

Grantz, A., Holmes, M. L. & Kosoki, B. A. 1976. Geologic framework of the Alaskan continental terrace in the Chukchi and Beaufort Seas. *Can. Soc. Petrol. Geol. Mem.* **4**, 669–699.

Harris, R. A. 1985. Paleomagnetism, geochronology and paleotemperature of the Yukon-Koyukuk province, Alaska. *Eos* **66**, 1103.

Lieberman, J. E. 1986. Structural and metamorphic continuity of the Kigluaik and Nome Groups, Seward Pen. Alaska: pressure–temperature–time path implications. *Geol. Soc. Am. Abs. w. Prog.* **18**, 673.

Lister, G. S. & Hobbs, B. E. 1980. The simulation of fabric development and its application to quartzite: the influence of deformation history. *J. Struct. Geol.* **2**, 355–370.

Lister, G. S. & Williams, P. F. 1979. Fabric development in shear zones: theoretical controls and observed phenomena. *J. Struct. Geol.* **1**, 283–297.

Mayfield, C. F., Tailleur, I. L. & Ellersieck, I. 1983. Stratigraphy, structure and palinspastic synthesis of the western Brooks Range, northwestern Alaska. *U.S. Geol. Surv. Open-File Report* 83-779.

Meneilly, A. W. & Storey, B. C. 1986. Ductile thrusting within subduction complex rocks on Signey Island, South Orkney Islands. *J. Struct. Geol.* **8**, 457–472.

Minnigh, L. D. 1979. Structural analysis of sheath folds in metachert from the Western Italian Alps. *J. Struct. Geol.* **1**, 275–282.

Passchier, C. W. & Simpson, C. 1986. Porphyroclast systems as kinematic indicators. *J. Struct. Geol.* **8**, 831–843.

Patrick, B. E. 1987. Petrological and structural studies in the Seward Peninsula blueschist terrane, Alaska. Unpublished Ph.D. thesis, University of Washington.

Patrick, B. E. 1986. Relationship between the Seward Peninsula Blueschists and the Brooks Range orogeny: evidence from regionally consistent stretching lineations. *Geol. Soc. Am. Abs. w. Prog.* **18**, 169.

Patrick, B. E., Evans, B. W., Dumoulin, J. A. & Harris, A. G. 1985. A comparison of carbonate mineral and conodont color alteration index thermometry, Seward Peninsula, Alaska. *Geol. Soc. Am. Abs. w. Prog.* **17**, 399.

Patton, W. W., Jr. & Tailleur, I. R. 1977. Evidence in the Bering Strait region for differential movement between North America and Eurasia. *Bull. geol. Soc. Am.* **88**, 1298–1304.

Platt, J. P. & Vissers, R. L. M. 1980. Extensional structures in anisotropic rocks. *J. Struct. Geol.* **2**, 397–410.

Pollock, S. M. 1982. Structure, petrology and metamorphic history of the Nome Group blueschist terrane, Salmon lake area, Seward Peninsula, Alaska. Unpublished M.S. thesis, University of Washington.

Ramsay, J. G. 1967. *Folding and Fracturing of Rocks.* McGraw-Hill, New York.

Rice, J. M. 1977. Contact metamorphism of impure dolomitic limestone in the Boulder aureole, Montana. *Contr. Miner. Petrol.* **59**, 237–259.

Roeder, D. & Mull, C. G. 1978. Tectonics of the Brooks Range ophiolites, Alaska. *Bull. Am. Ass. Petrol. Geol.* **62**, 1696–1713.

Sainsbury, C. L. 1972. Geologic map of the Teller Quadrangle, western Seward Peninsula, Alaska. *U.S. Geol. Surv. misc. field inv. map* I-685.

Sainsbury, C. L., Coleman, R. G. & Kachadoorian, R. 1970. Blueschist and related greenschist facies rocks of the Seward Peninsula, Alaska. *U.S. Geol. Surv. Prof. Pap.* **700-B**, 33–42.

Shackleton, R. M. & Ries, A. C. 1984. The relation between regionally consistent stretching lineations and plate motions. *J. Struct. Geol.* **6**, 111–117.

Thurston, S. P. 1985. Structure, petrology and metamorphic history of the Nome Group blueschist terrane, Salmon lake area, Seward Peninsula, Alaska. *Bull. geol. Soc. Am.* **96**, 600–617.

Till, A. B. 1983. Granulite, peridotite and blueschist: Precambrian to Mesozoic history of the Seward Peninsula. In: *Proceedings of the* 1982 *Symposium on Western Alaska Geology and Resource Potential. J. Alaska geog. Soc.* **2**, 59–66.

Till, A. B., Dumoulin, J. A., Gamble, B., Kaufman, D. & Carroll, P. I. 1986. Preliminary geologic map and fossil data, Solomon, Bendeleben and Southern Kotzebue quadrangles, Seward Peninsula Alaska. *U.S. Geol. Surv. Open-File Report* 86-276,

Quinquis, H., Audren, Cl., Burg, J. P. & Cobbold, P. R. 1978. Intense progressive simple shear in the Ile de Groix blueschists and compatibility with subduction or obduction. *Nature, Lond.* **273**, 43–45.

Warburton, J. 1986. The ophiolite-bearing Schistes lustrés nappe in Alpine Corsica: a model for the emplacement of ophiolites that have suffered HP\LT metamorphism. In: *Blueschists and Eclogites* (edited by Evans, B. W. & Brown, E. H.). *Mem. geol. Soc. Am.* **164**, 313–331.

Watkinson, A. J. & Cobbold, P. R. 1981. Axial directions of folds with linear/planar fabrics. *J. Struct. Geol.* **3**, 211–217.

Williams, G. D. 1978. Rotation of contemporary folds into the X direction during overthrust processes in Lakesfjord, Finnmark. *Tectonophysics* **48**, 29–40.

Yokoyama, K., Brothers, R. N. & Black, P. M. 1986. Regional eclogite facies in the high-pressure metamorphic belt of New Caledonia. In: *Blueschists and Eclogites* (edited by Evans, B. W. & Brown, E. H.). *Mem. geol. Soc. Am.* **164**, 407–423.