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Formation of systematic joints in metamorphic rocks due to release of residual elastic strain energy, Otago Schist, New Zealand

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A B S T R A C T

We present the first regional study of joints in the Jurassic–Cretaceous Otago Schist, New Zealand. The purpose of this study was to explore the origin and mechanism of joint formation in metamorphic rocks, especially any possible association between brittle and previous ductile deformation. The Otago Schist is cut by numerous systematic joints, up to tens of metres long, at any one exposure. We measured the orientation of joints, schist foliation planes, and quartz rods/mineral lineations at 46 sites across the Otago Schist, and calculated the spherical angles between their means. In relatively high metamorphic grade schists (greenschist facies) typically one systematic joint set has developed sub-perpendicular to penetrative foliation and lineation, irrespective of foliation and lineation orientations. This relationship also holds in lower grade schists (pumpellyite–actinolite facies), but more than one joint set is occasionally present. The flanking unfoliated schist protoliths (prehnite–pumpellyite facies) contain no systematic joint sets. A Late Cretaceous age for schist joint formation is indicated on the basis of lack of joint continuation into Late Cretaceous conglomerates that unconformably overlie jointed schists, cooling history, consistent orthogonality of joints with foliation and lineation, and lack of relationship of systematic joints to late Cenozoic plate-boundary features. We propose a model for joint formation during Late Cretaceous exhumation of the schist, and suggest that the systematic joints formed due to release of residual elastic strain energy preserved in the schists from Early Cretaceous ductile deformation.

1. Introduction

The most common geological structures that are created in the upper crust are joints. Joints are surfaces, fractures or partings in a rock across which there has been no displacement (Bates and Jackson, 1987). Systematic joints form regular, planar, sub-parallel sets. They profoundly control the shape of many spectacular landforms, and play an important role in the sub-surface transport of fluids such as water, magma, contaminants and hydrocarbons (e.g., Pollard and Aydin, 1988; Gross and Eyal, 2007). Establishment of reliable relationships between joints and their cause provides important tools for inferring the loading conditions and mechanical behaviour of rocks. It is widely agreed that joints form as mode I fractures perpendicular to the least compressive principal stress, and perpendicular to the maximum extension (elongation) for a coaxial strain field (e.g., Engelder and Geiser, 1980). The main jointing mechanisms (Engelder, 1985; Bahat et al., 2005) are responses of the host rock to a regional or local stress field, effect of pore pressure and hydro-fracturing, stress relaxation due to rock uplift, and/or jointing due to material shrinking (e.g., columnar joints in basalts). These mechanisms are all different manifestations of the brittle (non-penetrative) deformation of cold rocks under low lithostatic pressure, whereas ductile (penetrative) deformation occurs in a different tectonic environment where rocks are hot and under high lithostatic pressure. Therefore, in most cases, ductilely deformed rocks should have cooled down and lithostatic pressure been relieved by exhumation before the beginning of jointing. In the literature there are numerous published studies of joints and jointing mechanisms in sedimentary and igneous rocks (e.g., Bahat et al., 2005 and references therein), but comparatively few from metamorphic rocks. A major contribution of this study is the documentation of geometric relations between joints and penetrative structural-metamorphic fabrics in metamorphic rocks.

The Otago Schist is a major Jurassic–Cretaceous metamorphic belt in New Zealand (Figs. 1 and 2) that is superimposed on the late Paleozoic–Mesozoic sediments of the Caples and Torlesse Terranes

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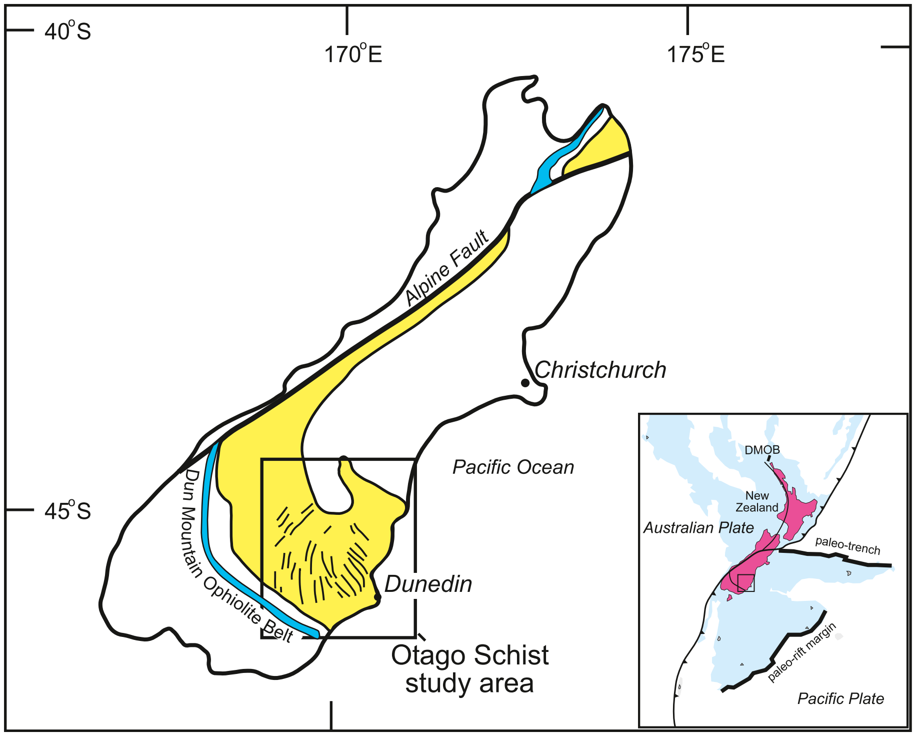

Fig. 1. Regional geological setting of the Haast Schist in New Zealand's South Island (grey area), and the location of the Otago Schist study area. Solid lines are trajectories of stretching lineations (after Mortimer, 1993). Inset: the location of the Cretaceous paleo-trench and paleo-rift margin marked on the present plate configuration. DMOB, Dun Mountain Ophiolite Belt.

(Mortimer, 1993). Joints in the Otago Schist, although common and widespread, have only rarely been mentioned in the New Zealand research literature (e.g., Paterson, 1941; Stirling, 1990; Markley and Norris, 1999). Geotechnical reports contain descriptions and measurements of rock mass fractures, including joints (e.g., Paterson, 1979; Macfarlane, 1986; Thomson, 1993), but these are restricted to individual sites and have never been synthesised or summarised. Consequently, a basic description and classification of joints and joint sets in the Otago Schist, as well as interpretations regarding their origin, time and mechanism of formation, is long overdue.

Most fractures in the Otago Schist are opening-mode features that commonly form sets of sub-vertical joints (Fig. 3a,b). Possible causes for the origin of these systematic joints include: (1) regional extension resulting from imposition of long-range plate-boundary forces either in the Late Cretaceous or the late Cenozoic; (2) influence of high pore pressure and formation of natural hydraulic fractures during the Cretaceous to Recent; (3) general stress relaxation during Late Cretaceous and/or late Cenozoic exhumation; (4) stress relaxation during Late Cretaceous exhumation controlled by inherited stresses from the ductile strain regime; (5) some combination of the above loading conditions. The purpose of this study was to investigate the origin of systematic joint formation in the Otago Schist, and use this as a case study to explore and analyse the possible association between brittle and ductile deformation in metamorphic rocks.

2. Geological setting

The Otago Schist is part of the Haast Schist belt of New Zealand and forms a wide (~150 km) structural arch in New Zealand's South Island (Fig. 1). The metamorphic grade ranges from prehnite–pumpellyite facies in the flanks to greenschist facies in the centre. The main protoliths of the Otago Schist are sandstones and mudstones of the Caples and Torlesse terranes that accumulated at the margin of Gondwanaland from the Permian to the Late Triassic, and were regionally metamorphosed from the Middle Jurassic to the Early Cretaceous (Wood, 1978; Mortimer, 1993; Bishop and Turnbull, 1996; Turnbull, 2000; Forsyth, 2001). The metamorphosed equivalents of thick-bedded sandstones and thin-bedded sequences containing sandstones and mudstones are represented by so-called "massive" and "layered" schist respectively (Brown, 1963). Regional metamorphism was accompanied by multiple ductile deformation phases (Table 1) and was associated with a peak temperature of up to 400 °C and burial depth of at least 30 km (Mortimer, 2000). The Otago Schist is divided, based on progressive development of foliation, segregation, and increasing grain size of metamorphic mica, into five textural zones: TZI, TZIIA, TZIIB, TZIII, and TZIV (Bishop, 1974; Turnbull et al., 2001). Arguably, these textural changes track bulk mesoscopic changes in anisotropy and stress- and strain-related properties in the schist more effectively than metamorphic grade alone (cf. Norris and Bishop, 1990). For example, quartz rods, up to 20 cm in length (Fig. 3i,j), are seldom observed in TZII, common in TZIII, and almost ubiquitous in TZIV. Okaya et al. (1995) have shown that laboratory measurements of *P*-wave velocities in Otago Schist are highly anisotropic with *Vp* parallel to foliation 1.2–2.0 times larger than *Vp* perpendicular to foliation. The results of unconfined compressive strength tests for geotechnical investigations have shown that the Otago Schist exhibits pronounced strength anisotropy, influenced by the orientation of the foliation with respect to the applied loading direction (Paterson, 1979; Smith and Salt, 1990). At shallow (0–20°) and steep (>70°) angles of dip the compressive strengths are relatively high. At moderate angles of dip (~65°) the compressive strengths decrease to about 30% of the strengths of shallow-dipping schist and failure occurs along the foliation planes (Read et al., 1987).

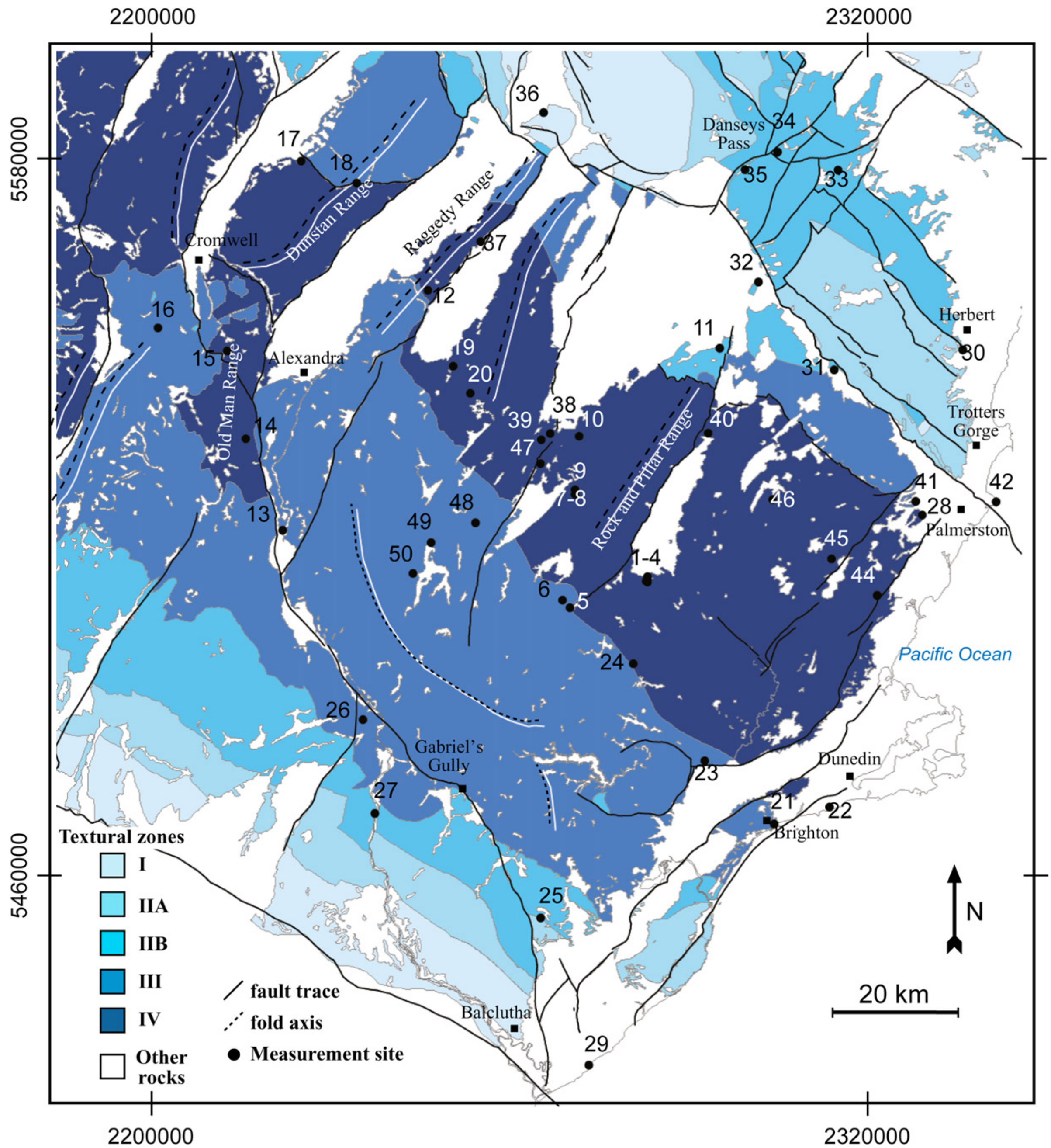

Fig. 2. Map of the Otago Schist study area coloured according to schist textural zones, and including location of the studied sites. Simplified maps of fault traces and Cenozoic fold axes in Otago are included. NE-trending faults are mainly active reverse faults. "Other rocks" represent both non-schist basement rocks (to the south west) and Late Cretaceous to Quaternary sedimentary basins (within the schist).

The earliest uplift and cooling of the Otago Schist is recorded by Middle–Late Jurassic (c. 150–170 Ma) K-Ar and Ar-Ar ages of metamorphic white mica from low grade (TZIIB-III) schists. Structurally deeper and higher metamorphic grade schists (garnet–biotite–albite zone; TZIV) give a spread of younger, Early Cretaceous, cooling ages (c. 150–110 Ma) that record a more protracted and/or episodic metamorphic and exhumation history (Adams and Gabites, 1985; Little et al., 1999; Forster and Lister, 2003; Gray and Foster, 2004). Estimations of schist exhumation rates range from an overall 0.2 mm/year (Adams and Gabites, 1985) to 0.6–1.0 mm/year after 135 Ma (Little et al., 1999). According to Forster and Lister (2003), core-complex related extension of the schists occurred from 109 to 112 Ma, but other mechanisms of schist exhumation such as ductile thinning of an accretionary wedge cannot yet be ruled out (Deckert et al., 2002, 2003). In the eastern part of Otago (Fig. 2), schists at all structural levels, textural zones and metamorphic grades were exposed at the Earth's surface by the Late Cretaceous, as revealed by a widespread Late Cretaceous (c. 85 Ma) marine unconformity in coastal Otago. The presence, in three locations, of c. 112 Ma terrestrial graben deposits containing schist clasts points to an even earlier exhumation phase (Bishop and Turnbull, 1996; Forsyth, 2001; Mortimer, 2003; Tulloch et al., 2009). In the western part of Otago near Cromwell (Fig. 2), zircon fission track ages of c. 70–120 (mainly c. 90) Ma, and apatite fission track ages of c. 35–65 Ma indicate the times at which these parts of the Otago Schist were last at c. 250 °C and c. 100 °C, respectively

Fig. 3. Photographs of structures observed in the study area. (a) Large-scale systematic joints in segregated TZIV schist. Schist foliation is dipping towards the observer (Poolburn Gorge, site 37). (b) Systematic joints in sub-horizontal, well-foliated and segregated schist tor, central Otago (Linnburn Road, site 47). (c) Closely spaced systematic joints in highly foliated, unsegregated, pelitic schist (Ramrock Road, site 45). A geological hammer 0.3 m long provides a scale. (d) Dipping (to the left), systematic joints in semischist oriented perpendicular to the steep, penetrative foliation (Danseys Pass area, site 35). Hammer length is 50 cm. (e) Torlesse greywacke dominated by short, non-systematic joints. The rocks retain their primary bedding (marked by arrows) that dips steeply to the left (Falls Dam, site 36). (f) Steep penetrative foliation (sunlit large planes) in semischist, and perpendicular systematic joint set (narrow planes in shadow). Non-systematic joints are also apparent (Danseys Pass area, site 33). (g) Two orthogonal sets of joints in semischist (Daisybank, site 11). Compass length is 90 mm. (h) Unconformity between the systematically jointed Otago Schist, and the unjointed overlying Late Cretaceous Taratu Formation (Palmerston, site 28). (i) Quartz rods in highly segregated schist. The pen is parallel to the NW-trending rods (Brighton, site 21). (j) Quartz rods in highly segregated schist; the quartz rods are oriented parallel to axes of small-scale folds. Joint planes are perpendicular to the trend of quartz rods and marked by an arrow. The black bar shows the trend of rods and pole to joint planes (Brighton, site 21). (k) Joints formed along kink folds in pelitic schist (Old Dunstan Road, site 19). Coin (in the lower centre part) is 30 mm in diameter. (l) Corrugations on quartz layer surface in highly segregated schist (Roxburgh, site 13). The corrugation striae orientation is parallel to fold hinges and quartz rods in the schist. Coin is 20 mm in diameter. (m) Set of quartz veins in highly segregated schist (Ophir, site 12) where the attitude of the quartz veins is parallel to the attitude of the systematic joints.

(Tippett and Kamp, 1993). Faulting of Cretaceous age in the schist has been demonstrated by Mutch and Wilson (1952), Bishop (1974) and Barker (2005).

New Zealand's Paleozoic–Mesozoic basement rocks, including the schist, have been deformed into a large-scale Cenozoic orocline defined by the curve of the Permian Dun Mountain Ophiolite Belt (e.g., Norris, 1979; Sutherland, 1999; Little and Mortimer, 2001; Fig. 1). In central and eastern Otago, late Cenozoic deformation along the Australian–Pacific plate boundary has resulted in uplift of basement blocks by folding and reverse faulting with predominant

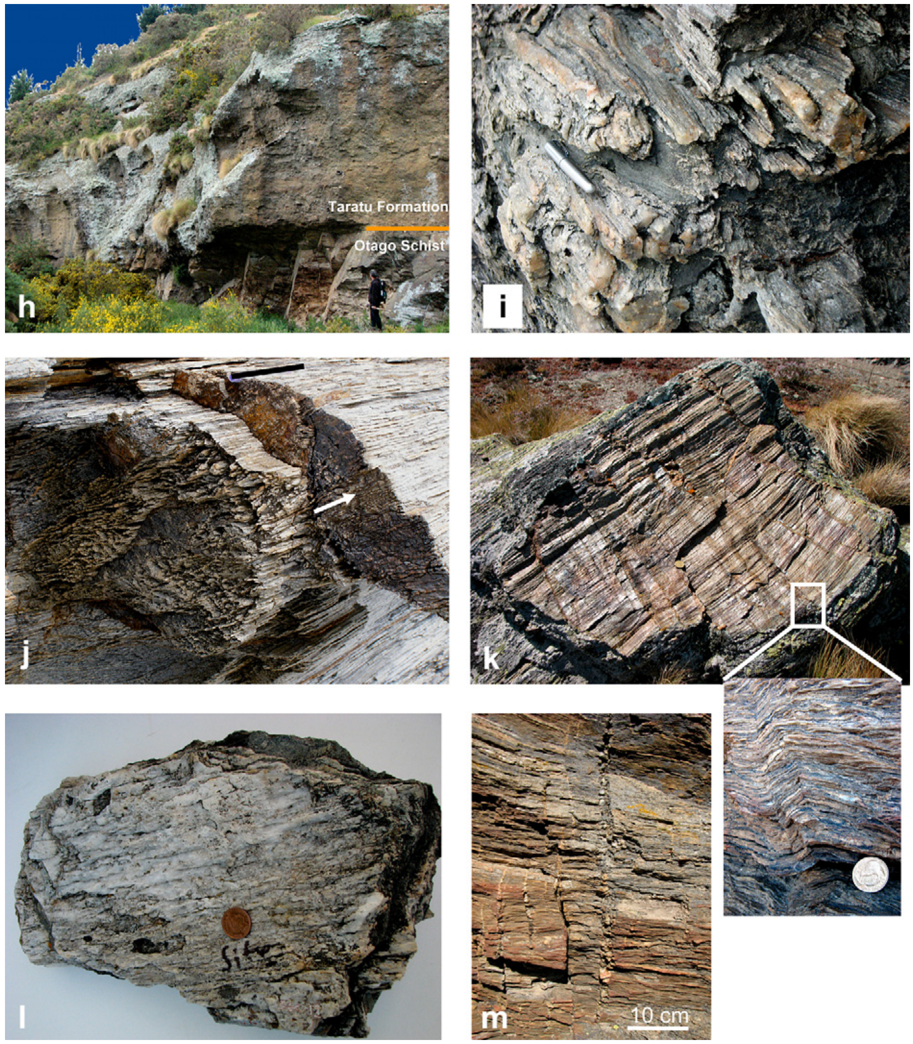

Fig. 3 (*continued*).

NE–SW strikes (Fig. 2). Based on this geological history (Table 1), jointing of the schists could have occurred during Late Cretaceous exhumation or late Cenozoic tectonism.

3. Methods

We measured the orientation of joints, schist foliation planes, quartz rods, mineral lineations, fold axes and quartz veins at 46 sites across the Otago Schist, from Brighton in the east to Cromwell in the west, and from Herbert in the north-east to Balclutha in the south (Fig. 2). The sites were selected to include different structural orientations of ductile foliation and lineation, different textural zones and different schist rock types. The latter included massive schist and layered schist (Brown, 1963), as well as rocks such as slate and TZI greywacke. In order to eliminate the influence of late Cenozoic faulting, we avoided sites within the damage zones of major faults (Bishop and Turnbull, 1996; Turnbull, 2000; Forsyth, 2001). For comparison with the schist dataset, we also measured joints and small faults in Cretaceous–Paleogene cover rocks at three sites in coastal Otago.

Wherever possible, we measured only systematic joints, i.e., smooth, planar and regularly-spaced fractures. Commonly,

Table 1
Summary of structures and inferred age in Torlesse (Rakaia) terrane (Forsyth, 2001).

Phase	Structures	Inferred age
0	Bedding	Carboniferous–Permian–Triassic
1	Transposed bedding, schistosity, mineral segregation, steeply plunging folds, tight isoclinal folds	Jurassic (early regional metamorphism)
2	Recumbent folding, nappe sheets and high strain zones, stretching lineation, veins, foliation, schistosity, quartz rods	Jurassic (late regional metamorphism and terrane juxtaposition)
3	Brittle–ductile fault zones, mainly normal faulting, kink folds, mineralisation and veins, joints, striations on quartz veins and/or quartz rods	Cretaceous (uplift and exhumation of schist prior to peneplanation, reduction of the lithostatic pressure)
4	Broad folds, mainly reverse faults, kinks, chevron folds, joints	Late Cenozoic (regional uplift and crustal shortening across Otago due to tectonic events along the Australian–Pacific plate boundary)

systematic joints occur as sets of sub-parallel joints over tens of metres. In contrast, curved and non-parallel fractures are regarded as non-systematic joints; these apparently do not form regular joint sets and were measured only when systematic joints appeared absent. It should be emphasised that the purpose of this study was not to measure *all* joints (systematic and non-systematic) at exposures in order to quantify joint spacing and densities (e.g., Stirling, 1990; Thomson, 1993). For structural measurements we used a *Breithaupt Kassel* compass, estimating the accuracy of the measured orientations to $\pm 2^{\circ}$. In order to calculate the angle between the joints and the various metamorphic structures in the schists, we measured, at the same site, the orientations of quartz rods, mineral lineations, and foliation, irrespective of D1 or D2 generation. Wherever possible, we measured over 20 joints, 15 rods/lineations and five foliations at each site. At several sites where no single systematic joint set was obvious, we represented the population by measuring 35–65 representative joints. We determined schist textural zone according to the mica grain size (Turnbull et al., 2001), and metamorphic grade from published 1:250 000 scale maps (Bishop and Turnbull, 1996; Turnbull, 2000; Forsyth, 2001).

Data were plotted and analysed using stereographic plots and Fisher statistics incorporated in the Stereonet 1.2 program by R. Allmendinger, Cornell University (see http://www.geo.cornell.edu/geology/faculty/RWA/programs.html). For the present analysis two spherical angles, ψ and θ, were defined for each site: ψ is the angle between the mean pole to a set of joint planes and the mean pole to schist foliation, and θ is the angle between the mean pole to a set of joint planes and the mean orientation of a population of quartz rods/mineral lineations.

At site 29 (Fig. 2), measurement of small reverse faults in a sandstone unit enabled a kinematic analysis of the fault data using the FaultKin program (Allmendinger et al., 2001). P and T axes were calculated for each fault based on fault orientation, trend of striae and the sense of motion. The mean P and T axes provide a first-order approximation of the infinitesimal maximum shortening and maximum extension directions, respectively, for this fault population.

Structural measurements for each site are summarised in the Supplementary Material, Item 1. All raw data have been entered into the GNS Science QMAP database (http://www.gns.cri.nz/research/qmap).

4. Results

4.1. Joint, vein, and corrugation striae characteristics

Systematic joints can be seen in many, but not all, exposures of Otago Schist. Systematic joints in the Otago Schist are smooth and seldom have fractographic features (fracture surface morphologies). They occur in a wide variety of lithologies, textural zones, and metamorphic grades, and have two main types of appearance:

(i) Most joints, individually up to tens of metres high and long, that cross road/rail cuttings and/or schist tors from end to end (Fig. 3a,b). The joints consist of smooth, planar surfaces that are typically spaced 1–10 m apart. They are commonly observed in coarse-grained, medium to highly segregated schists of textural zones III and IV. These rocks are characterised by prominent layering of dark mica-rich and white quartz–plagioclase-rich layers and commonly contain isoclinal folds and refolded folds.

(ii) Other joints up to several metres (but commonly less) high and long that are confined within an individual schist tor or other exposure. These joints consist of smooth surfaces, which are either planar or slightly curved, and are closely spaced (<1 m) (Fig. 3c). They are commonly observed in fine-grained pelitic schists (slates and phyllites) of textural zones IIA and IIB. These schists, while containing ductile folds, are usually unsegregated (rarely consist of quartz or quartz–plagioclase layers).

Another, more rare, set of joints accompany kink-folds in well-foliated, thin-layered schists (Fig. 3k). Their dimensions and spacing are dictated by that of the kinks. There is evidence for several development stages of jointing along the kink-folds (Fig. 3k) from the initial stage, in which a short joint was formed along a small part of the kink, to a through-going joint resulting from coalescence of several, small, kink-related joints along the entire kink. An even more rare set of joints, recorded only at site 25, are closely spaced, sub-horizontal fractures in fine-grained, low textural zone schist (Supplementary Material, Item 1). These "sheet joints" are differentiated both from the primary bedding where visible, and from the foliation in TZIIB schists, based on sporadic occurrence of fractographic features.

The majority of joints in Otago Schist do not contain mineral infilling. However, a few quartz-filled joints, up to 10 mm thick, were observed in segregated schists of textural zones III and IV (e.g., site 21) (Fig. 3j). The appearance of these quartz-filled joints is different from cross-cutting quartz veins observed at sites 12, 13 and 16 (Fig. 3m). The latter "gash" veins are much shorter (<1 m), have an elliptical cross section, form a set of closely spaced (<0.1 m) features and occasionally accompany kink traces.

Corrugation striae are 0.5–5 mm amplitude and wavelength grooves that form at the contact between quartz–plagioclase layers and mica layers in highly segregated schists and quartz rods (e.g., Roxburgh and Brighton, sites 13 and 21; Fig. 3i,j). They are generally parallel to other dominant linear fabrics in schist outcrops e.g., quartz rods and mineral lineations. Quartz rods, mineral lineations, schist foliation and fold hinges are well-studied features in Otago Schist and their characteristics are described elsewhere (e.g., Craw, 1985; Mortimer, 1993).

4.2. Orientations of schist structures

We measured the orientation of 1230 joints and 742 quartz rods/mineral lineations. Stereographic projections of all joint and lineation orientations are presented separately for each site in Fig. 4, with the means in Fig. 5. The trends of the quartz rods and mineral lineations in the Otago Schist vary considerably from site to site (Fig. 4). The areal distribution of lineation orientations obtained in the present study is very similar to that observed in previous studies (Fig. 1), in which trend changes of c. 80° across Otago have been noted (Mortimer, 1993; Bishop and Turnbull, 1996; Turnbull, 2000; Forsyth, 2001). The joint strikes also vary across Otago by c. 80° (Fig. 4). The projected mean poles to joint planes are distributed along the primitive circle (Fig. 5a), indicating sub-vertical joint planes with various direction of strikes. This strike distribution is similar to the trend distribution of the quartz rods and mineral lineations in the schist (Fig. 5b). Notably, the scatter of joints in single sites is larger than that of quartz rods and mineral lineations (Supplementary Material, Item 1).

Fig. 6 shows the areal distribution of both joint and lineation orientations in the Otago Schist. The orientations of the dominant joint sets in the Otago Schist are presented here for the first time and, hence, are described in more detail. The general strike of the joint sets in the eastern part of the study area, from Brighton to Palmerston (e.g., sites 21, 28, 41, 44), is ENE–WSW. At one site (site 31), a NNW–SSE-trending joint set is also observed. In the northern

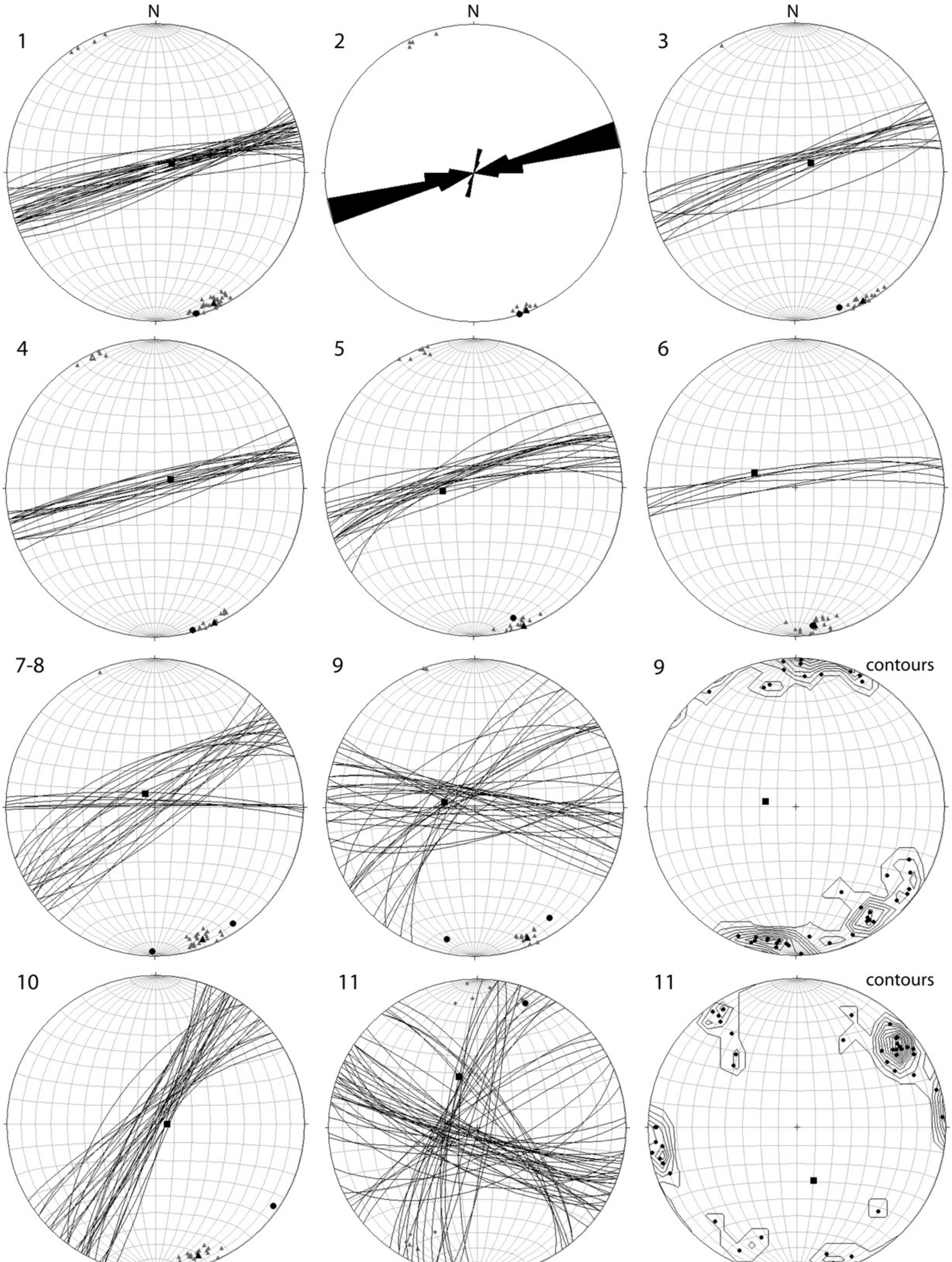

Fig. 4. Lower hemisphere stereographic projection of the structural data collected in the Otago Schist. Sites are denoted by numbers. Black and grey great circles represent joint and quartz vein planes, respectively; solid triangles and diamonds represent trends of quartz rods and mineral lineations, respectively; open triangles represent corrugation striae of quartz layers; open diamonds represent fold hinges. Large solid circles, triangles, diamonds, and rectangles represent mean pole of joints, quartz rods, mineral lineations, and foliation, respectively. Integrated rose diagrams represent joints/veins where only their strikes were measured. Contour intervals in sites 9 and 11 are 2% per 1% area. See Supplementary Material, Item 1 for the number of measurements and statistical data.

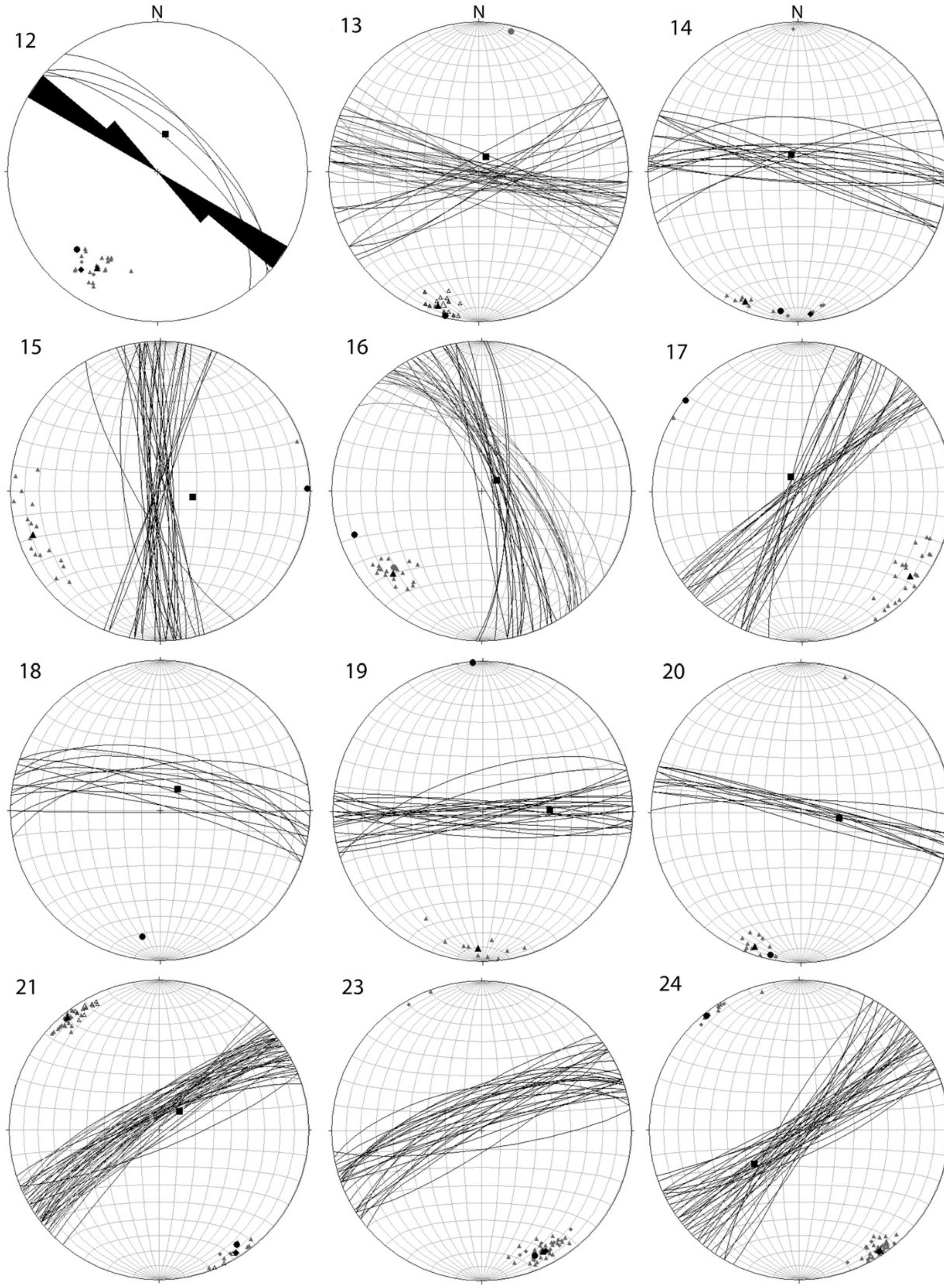

Fig. 4 (*continued*).

part of the study area near Danseys Pass, NE–SW striking joints were measured (e.g., sites 33–35). The central part of the study area is characterised by ENE–WSW striking joints (e.g., sites 1–8, 38, 45–49). Another, less dominant joint set observed in this area is striking NW–SE (e.g., 39). In the western part of the study area the orientations of the joints vary from range to range. In the Raggedy Range, the joints strike NW–SE (sites 12 and 37), whereas in the Old Man Range they strike NNW–SSE (e.g., sites 15 and 16) and E–W (site

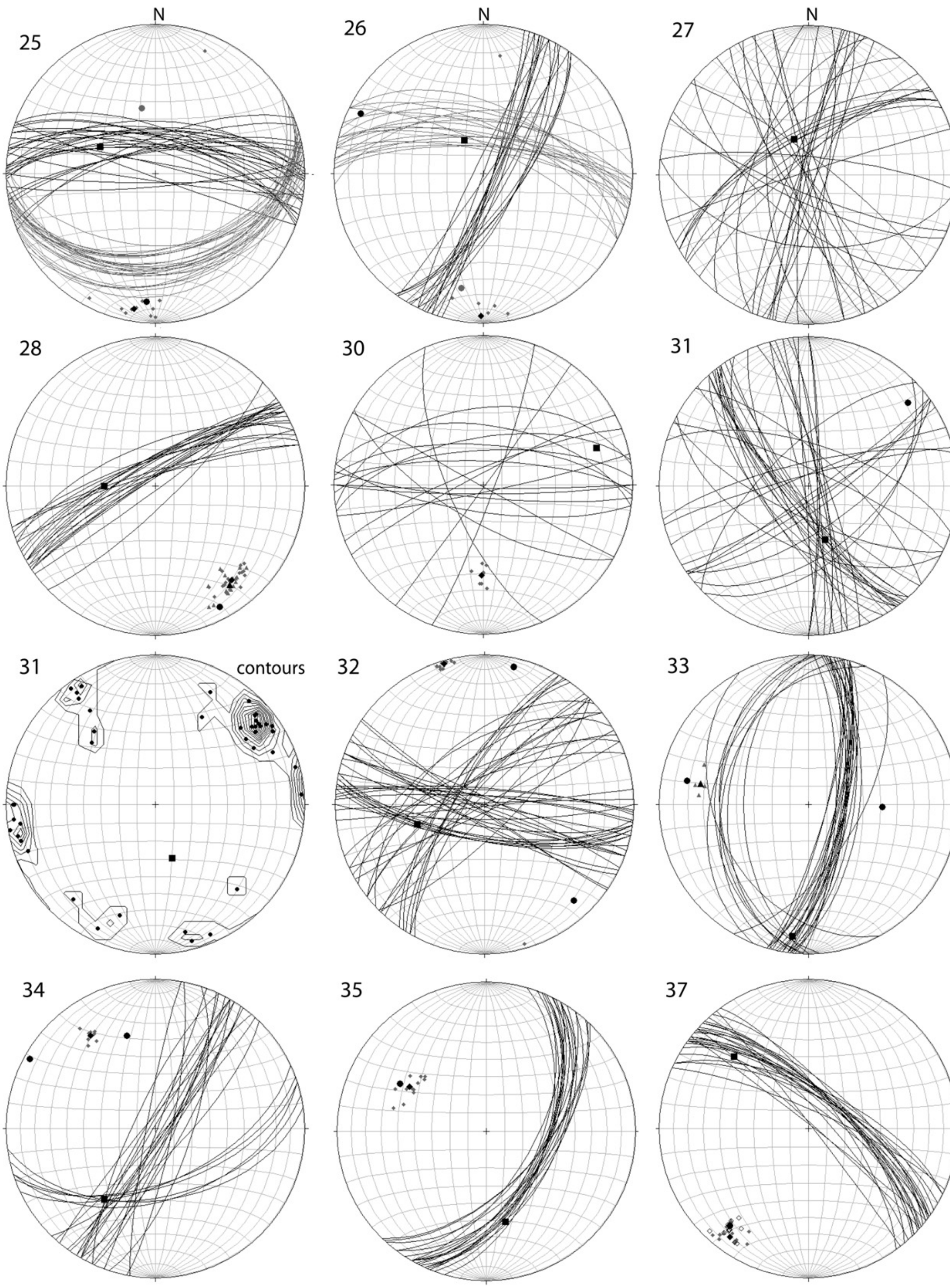

Fig. 4 (*continued*).

14). At a few sites we were able to measure sets of quartz-filled joints and veins that are commonly sub-parallel to the respective systematic joint set measured at the same site (Fig. 4, sites 12, 13, and 21).

4.3. Spherical angle ψ of joints

The frequencies of ψ, the calculated angle between the mean pole to a joint set and the mean pole to foliation at the same site are

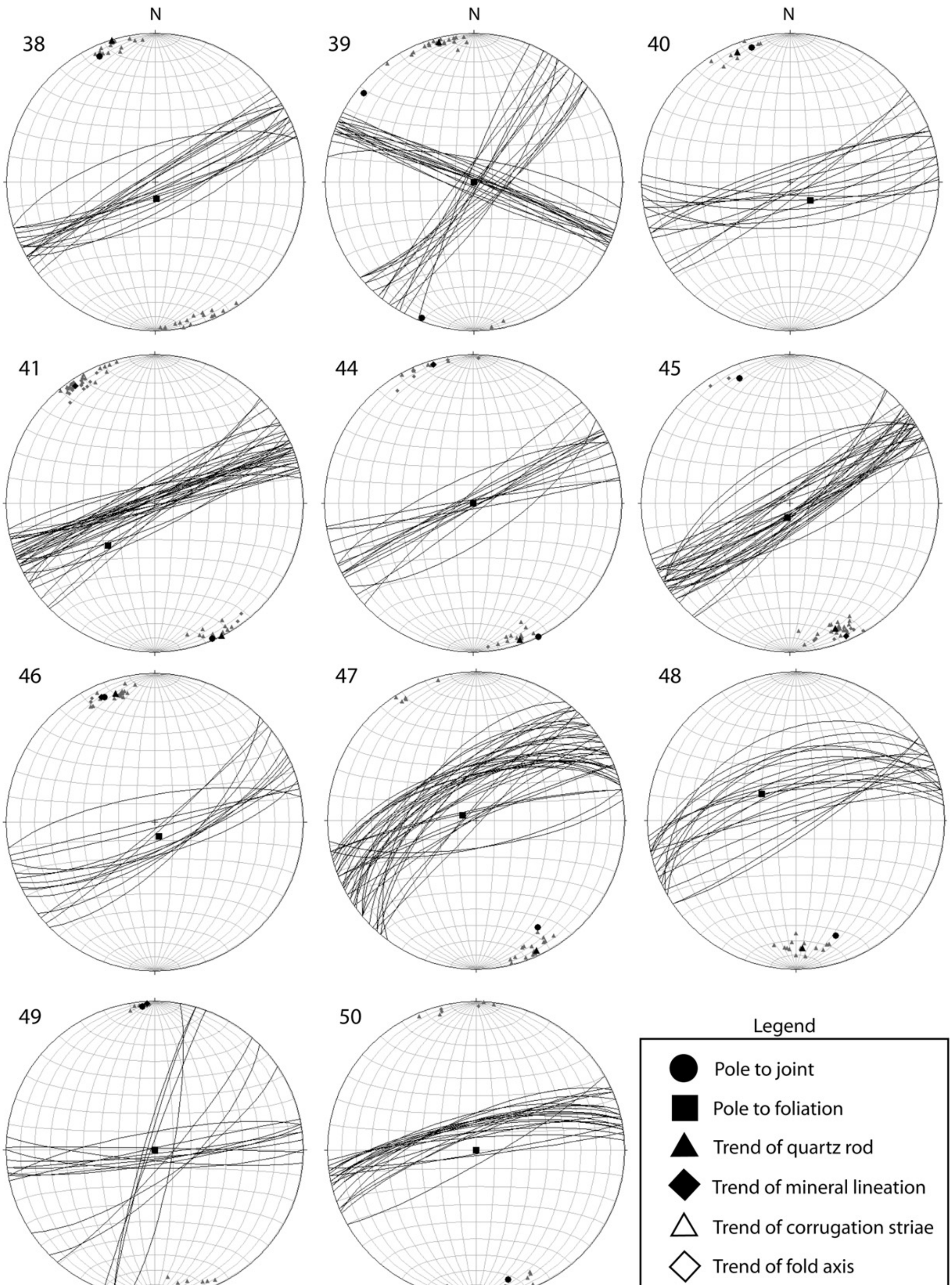

Fig. 4 (*continued*).

presented in Fig. 7a. The majority of the calculated ψ (36 out of 42 angles) are greater than 80°, indicating sub-perpendicular relationships between the two planes. The peak frequency of ψ is between 85 and 90° and the mean is 85.9° (3.5° standard deviation). Hence, although the absolute dips and dip directions of the joints and the foliations vary across the Otago Schist (Fig. 4) the spherical angle ψ remains high with little scatter (Fig. 7a). The joints are thus sub-perpendicular to the foliation, independent of

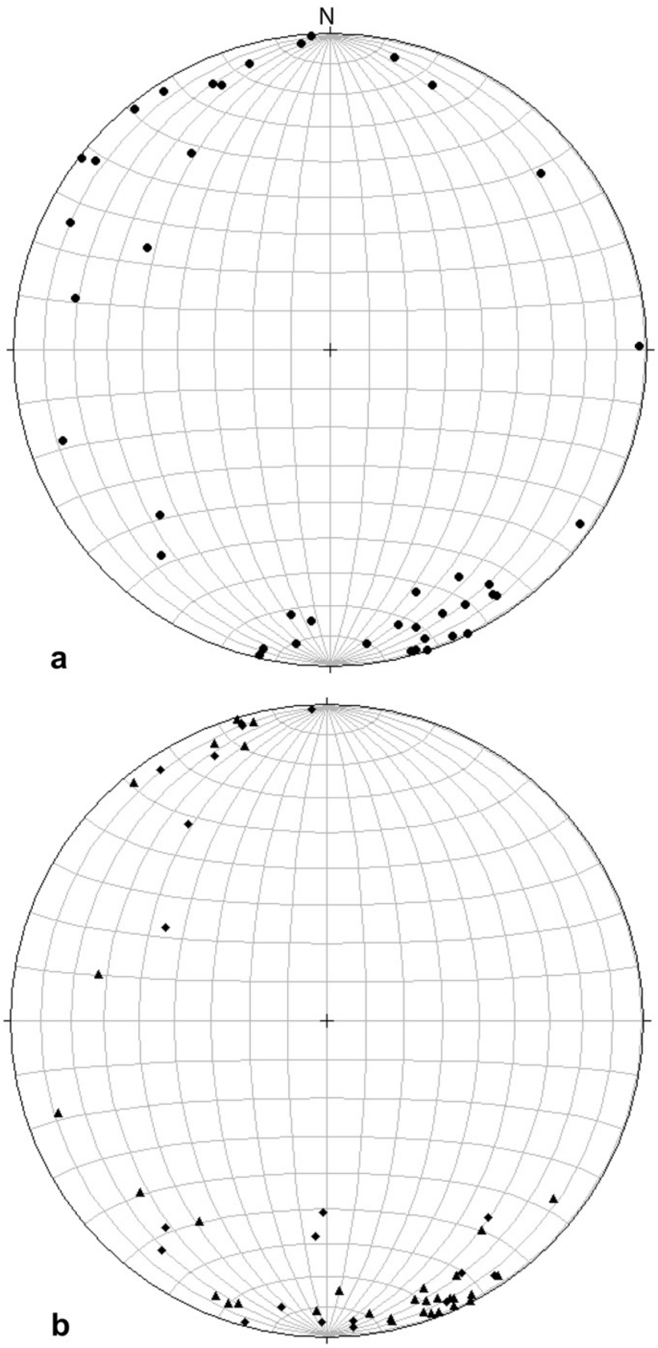

Fig. 5. Lower hemisphere stereographic projection of (a) means to joint poles (circles) from 42 sites, and (b) mean trends of quartz rods (triangles) from 33 sites, and mean trends of lineations from 20 sites (diamonds) in the Otago Schist. Legend as for Fig. 4.

whether the foliation is sub-horizontal (most sites, e.g., Fig. 3i) or very inclined (Fig. 3d,f). For example, the high spherical angle holds even when the foliation dips as much as ~80° (Danseys Pass; Fig. 4, site 33).

4.4. Spherical angle θ of joints

The frequencies of θ, the calculated angle between the mean pole to a set of joint planes and the mean trend of either L1 or L2 quartz rods and mineral lineations, are presented in Fig. 7b. The majority of the calculated θ (42 out of 52 angles) are smaller than 15°. The peak frequency of θ for quartz rods is between 5 and 10°, and the mean is 11.0° (7.6° standard deviation). The peak frequency of θ for mineral lineations is between 10 and 15°, and the mean is 11.3° (7.9° standard deviation). Hence, although the orientations of the joints and the lineations vary considerably (Figs. 5 and 6) the spherical angle θ shows a limited range of values and deviations (Fig. 7b). Some of the higher θ values in Fig. 7b arise from rare situations where two sets of joints, bisected by the lineation, are present (e.g., sites 32, 39). Notably, low values of θ are maintained even when the quartz rods/mineral lineations are inclined. For example, the plunge of the mineral lineations along Danseys Pass is 40°, (Fig. 4, site 35) but the spherical angle θ is still less than 10° (Supplementary Material, Item 1).

4.5. Relations between jointing and schist textural zones

We observed a relationship between the schist textural zone, and the type of jointing as follows: (a) unfoliated indurated greywacke sandstones (TZI) show non-systematic short joints that are scattered in all directions with no relation to bedding (Fig. 3e, Forsyth, 2001; Fig. 17). (b) Semi-schists (TZIIA, B) are frequently characterised by several joint set orientations (Fig. 3g), one of which may be perpendicular to the foliation and lineation, and two of which may be bisected by the lineation trend. (c) Schists belonging to TZIII and IV commonly have just one set of systematic joints, which is oriented sub-perpendicular to both schist foliation and lineation.

The relations between the spherical angles θ and ψ and schist textural zone is presented in Fig. 8a,b. The mean θ is relatively low in schists of zone III (~10°) and zone IV (~12°), and is higher (~15°) in schists of zone II. Likewise, the relations between θ and schist metamorphic grade (Fig. 8c,d) indicate that mean θ is relatively low (~10°) in schists of high metamorphic grade, and is higher (~15°) in schists of low metamorphic grade. The angle ψ is independent of the schist textural zone, showing a constant value of ~85° and a similar standard deviation of 4–5°. Thus joints are almost always perpendicular to foliation, regardless of grade.

4.6. Jointing and faulting in sedimentary rocks

Joints and small faults were studied in Late Cretaceous–Miocene strata at Tunnel Beach (site 22), Shag Point (site 41) and Wangaloa (site 29). The stereographic projections of the collected data are presented in Fig. 9. The Early Miocene marine calcareous sandstone (Otakou Group, Tunnel Beach) is crossed by ~20 m high joints whose WNW–ESE strikes indicate a least compressive principal stress of ~020–200° (Fig. 9a). The Late Cretaceous coarse quartz sandstone of Taratu Formation at Shag Point is cut by two sets of systematic joints. The younger set strikes WSW–ENE similar to that at Tunnel Beach, and the older set strikes WNW–ESE (Fig. 9b). A kinematic analysis of small-scale faults in Paleocene Taratu sandstone (Lindqvist and Douglas, 1987; our site 29) indicates that the instantaneous shortening direction is ~290–110°. The fault solution is that of almost pure thrust motion along NNE–SSW striking fault planes (Fig. 9c).

In Gabriel's Gully, Trotter's Gorge and near Palmerston (site 28), basal Cretaceous quartz-rich conglomerates unconformably overlie systematically jointed schists (Fig. 3h). However, at each of these sites, no evidence for continuation of the underlying systematic joints into the conglomerates was found, and the indurated conglomerate is not fractured. These observations have implications for the time at which the joints formed in the schist (see below).

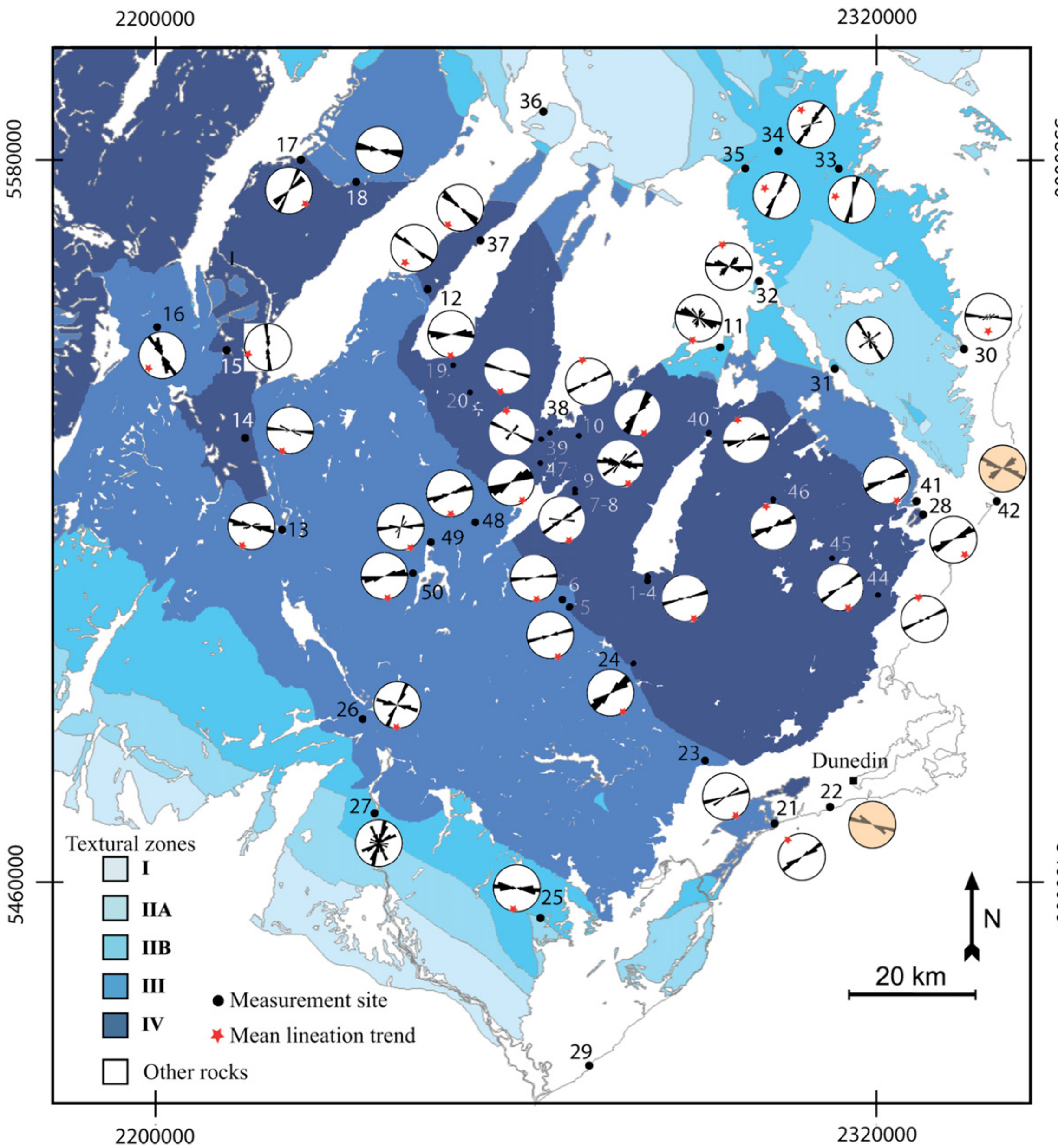

Fig. 6. Map of the Otago Schist showing rose diagrams of the joint strikes and mean trend of quartz rods/mineral lineations (red stars) for each studied schist site. Rose diagrams with pink background show orientations of joints in sedimentary rocks. Textural zones as in Fig. 2.

5. Discussion

The Otago Schist contains numerous joints that help to control the shape of the schist tors, stream orientations and landslide locations. Until now, there has only been passing mention of joints in Otago Schist, which provisionally connected their formation to local late Cenozoic tectonics (e.g., Stirling, 1990). However, our new results indicate that the angular relations between systematic joint orientations and that of the metamorphic fabric are predictable across our entire 20 000 km^2 Otago study area, and that a Late Cretaceous timing of systematic jointing can be constrained by field observations. Below we discuss these aspects and develop a conceptual model for joint formation in the schists resulting from release of lineation-related residual elastic strain energy during Late Cretaceous exhumation, rather than to any long-range tectonic forces or abnormal pore pressure at any other time.

5.1. Sub-perpendicular relation between joints and metamorphic fabrics

The sub-perpendicular relationship between joints and foliation (angle ψ) could result from rock anisotropy, whereby joints preferentially form perpendicular to discontinuities in the rock. Such relationships have been mainly observed in sedimentary rocks, where joints form perpendicular to bedding or existing fractures (e.g., Gross, 1993). Moreover, foliation-normal joints have been reported from other places where metamorphic rocks occur, e.g., Saghand region, central Iran (Verdel et al., 2007) and Eastern External Rif, Morocco (Azdimousa et al., 2007). However, such perpendicular relationships between two planes (i.e., foliation and joint) would not necessarily pose restrictions on the strike of the joints. In the entire area of the Otago Schist, the small spherical angle θ (between joint poles and lineations) strongly indicates

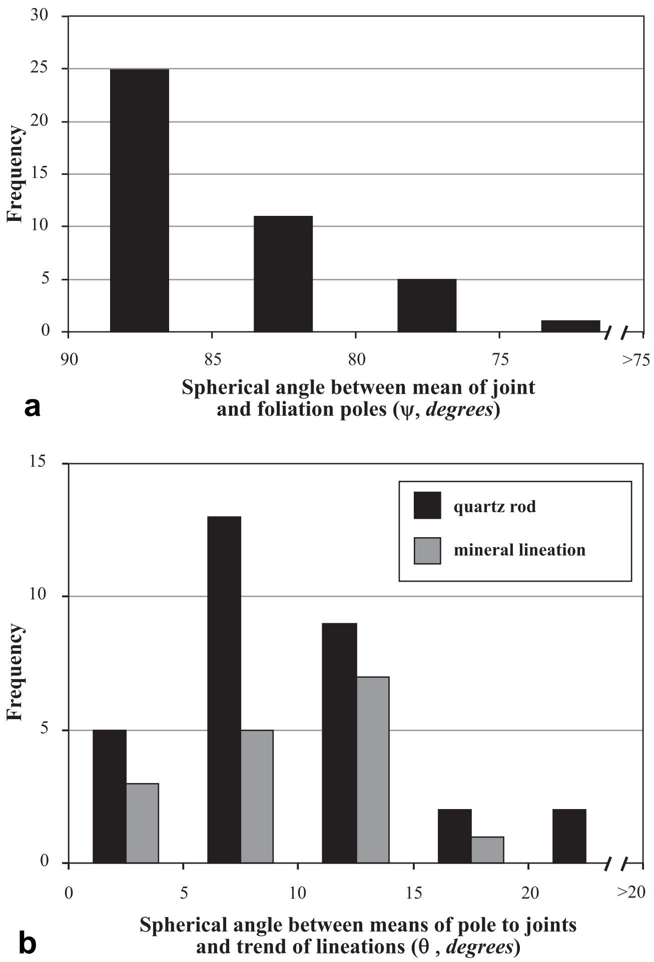

Fig. 7. Histograms of calculated spherical angles between mean pole to joints and (a) mean pole to foliations ($n = 42$, black bars); (b) mean trend of quartz rods ($n = 33$, black bars), and mineral lineations ($n = 18$, grey bars).

a structural association between the brittle (joint) and ductile (schist lineations) deformation. This association is strengthened by the fact that θ remains small despite the c. 80° variation in the trends of lineations in the Otago study area (Fig. 11).

The above observation helps to eliminate a late Cenozoic stress field origin for the lineation-perpendicular systematic joints in the Otago Schist. In several locations throughout Otago the Cretaceous lineation trajectories are parallel to the late NE-trending Cenozoic fold axes (e.g., Dunstan and Raggedy ranges). This might lead to the interpretation that SE-striking joints in these ranges (sites 12, 37) are classic "*ac*" joints sub-perpendicular to late Cenozoic fold axes. However, there is consistently a 30° mismatch in the orientation of these structures next to the axis of the Rock and Pillar Range (sites 1–6, 40). Moreover, throughout most of eastern Otago (e.g., east of the Rock and Pillar Range), where prominent late Cenozoic folds are absent, the spherical angle θ is still small (e.g., at sites 28, 41, 44–46), indicating no control of the joints by late Cenozoic deformation. A similar angular relation between joints and lineations in the Otago Schist has been observed in other studies where all joints (both systematic and non-systematic) were measured (Paterson, 1941; Stirling, 1990; QMAP database). Examination of Fig. 2 of Paterson (1941) and Fig. 8 of Stirling (1990) reveals that, out of all their measured joint sets, it is the systematic joints (as measured in the same areas for this study) that are the dominant sets (Fig. 2). Other, less systematic, joints in the Otago Schist could indeed be the result of late Cenozoic folding, as inferred by Markley and Norris (1999).

The regionally curving trends of joint orientations with low θ values in the Otago Schist (Fig. 10) allow us to reject a model for the formation of the Otago Schist systematic joints by a Late Cretaceous regional stress field. This is because the trend of the joints varies across Otago and bears no consistent relation to the Gondwana margin paleo-trench (Wood, 1978; Mortimer, 1993), the Zealandia–Antarctica breakup margin (Tulloch et al., 2009), or to known Cretaceous faults in Otago (Figs. 1 and 10; see also Bishop and Turnbull, 1996; Turnbull, 2000; Forsyth, 2001; Barker, 2005). A palinspastic correction of the Cenozoic orocline (Norris, 1979) is not needed in this analysis because our study area is opposite the straight part of the Dun mountain ophiolite belt (Fig. 1). While pore pressure might play a role in assisting fracturing in the Otago Schist, the majority of the systematic joints are barren, lack typical fractographic features typical of hydro-fracturing, and therefore show no supporting evidence for fluid overpressuring.

To better understand the structural association between joints and lineations as revealed by consistently very low values of θ, it is worth discussing what the ductile lineations in the Otago Schist represent. In one interpretation, the quartz rods formed parallel to the maximum regional extension direction (e.g., Mortimer, 1993). In another interpretation they formed initially by folding of the quartz-rich layers and isolation of the hinges by thinning of the fold limbs, or transportation of quartz from limbs and recrystallisation along fold hinges (e.g., Craw, 1985; Cox, 1991). With increasing deformation, these quartz rods/hinges are expected to rotate towards the maximum extension direction. Hence, in both interpretations, for the purposes of joint formation, the quartz rod orientations represent a direction of close to the maximum ductile extension. The aforementioned studies interpreted the schist to have formed during crustal thickening in an accretionary orogen. However, in recent years, it has been recognised that the planar and linear fabrics in the high grade (TZIII and TZIV) parts of the schist could have formed in an extensional ductile regime (Deckert et al., 2002; Forster and Lister, 2003; Mortimer, 2003; Gray and Foster, 2004). In this scenario, the quartz rods and other mineral lineations in TZIII and IV schist approximate the finite extension direction during ductile thinning.

The consistent low values of θ lead us to propose a mechanism for joint formation in which elastic strain energy related to ductilely-formed lineations, was not totally released. Although most accumulated ductile strain would be visco-plastic and not influence joint formation, Kunz et al. (2009) show that residual elastic strain can be preserved in deformed minerals during the ductile deformation. We suggest that residual elastic strain remained in the schists after transition of crust from a ductile to a brittle regime was subsequently released by jointing. Norris and Bishop (1990) showed that schists of higher textural grade were more highly strained and minerals are more preferably oriented (Turner, 1938). As deformation continues, preferred orientation of the elastically and plastically strained minerals increases, and, consequently, a higher correlation between the lineation and joint orientations might be expected in higher grade rocks. Although the present data show weak dependence between θ and schist textural zones and metamorphic grades (Fig. 8), there is a gradual focusing of preferred joint orientation with the evolving schist textural zones (see Section 4.5). The fractures in greywackes belonging to TZI are short, and the joint orientations are scattered in all directions. In semi-schists belonging to TZII (A, B), rock fracturing results in a few joint sets, whereas in schists belonging to TZIII and IV commonly one set of systematic joints, oriented sub-perpendicular to lineation, is formed.

At several sites of TZII (sites 31, 32) and TZIV schists (sites 9, 39), a high θ angle and more than one set of joints were measured. In these cases, the schist lineation sometimes bisects the acute angle between the joints (as also observed in other metamorphic

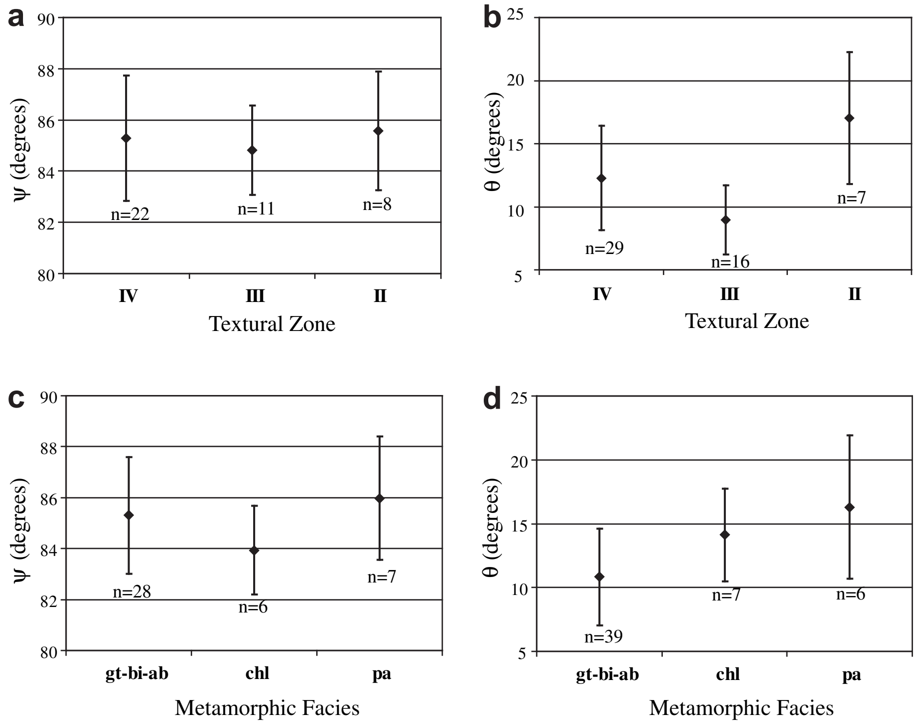

Fig. 8. (a) Relations between schist textural zones and ψ; (b) relations between schist textural zones and θ; (c) relations between metamorphic facies and ψ; (d) relations between the metamorphic facies and θ. Metamorphic facies; pa, pumpellyite–actinolite; chl, greenschist facies chlorite zone; gt-bi-ab, greenschist facies garnet–biotite–albite zone. Bars are one standard deviation.

terrains; e.g., Azdimousa et al., 2007). This weaker structural association between schist lineations and joints could result from one or both of two possibilities: (i) stress perturbation by the Late Cretaceous regional stress field (intrinsic stress relaxation versus extrinsic regional stress field), and/or (ii) weaker residual elastic strain that enabled scatter in joint orientation.

5.2. Timing of joint formation

The WNW–ESE direction of the joints in Miocene calcareous sandstones at Tunnel Beach (site 22, Fig. 6) indicates a least compressive principal stress of c. 020–200° that was acquired after the Early Miocene. This direction is compatible with the direction of the older joint set in Late Cretaceous sandstones at Shag Point (site 42, Fig. 6), as well as with the WNW–ESE shortening direction, inferred from kinematic analysis of small faults in the Paleocene sandstone at Wangaloa (site 29, Fig. 2). A maximum WNW–ESE compressive principal stress (and compatible least compressive principal stress of NNE–SSW) were also obtained from studies of Miocene dikes near the Alpine Fault (Cooper et al., 1987), focal plane solutions of recent earthquakes (Leitner et al., 2001), plate tectonic modelling (Hillis and Reynolds, 2000), and interpretation

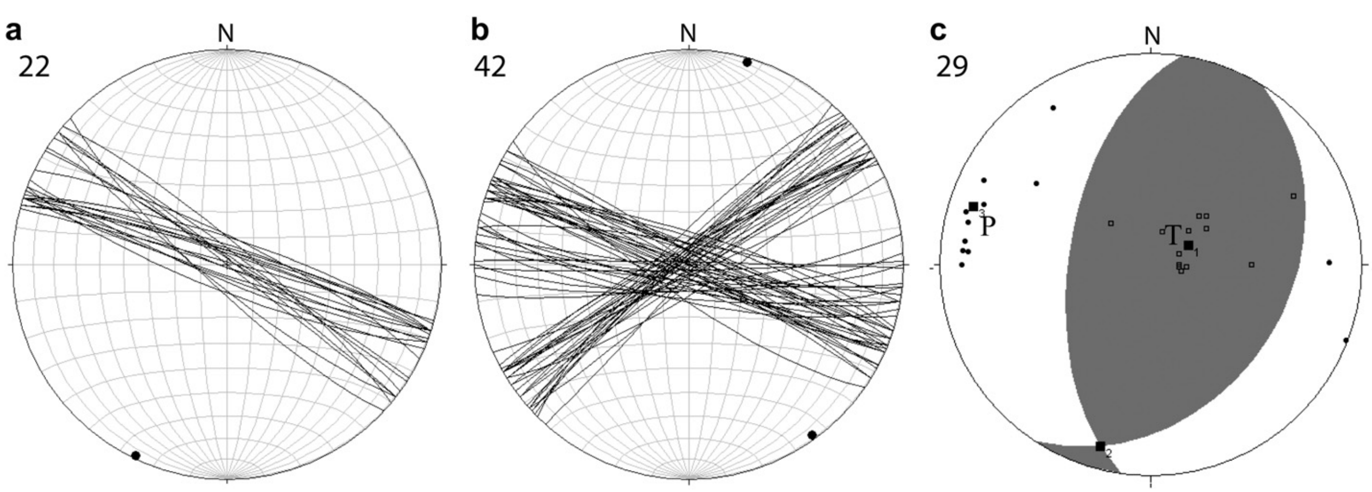

Fig. 9. Lower hemisphere stereographic projection of structural data collected in sedimentary rocks along the Otago coast. (a) Joint orientations in Early Miocene calcareous sandstone, Otakou Group (Tunnel Beach, site 22; mean pole 1/206, $\alpha_{95} = 4.3$, $n = 22$). (b) Joint orientations in Cretaceous coarse quartz sandstone, Taratu Formation (Shag Point, site 42; mean pole of the old Set, A, 2/016, $\alpha_{95} = 3.5$, $n = 39$; mean pole of the young Set, B, 2/144, $\alpha_{95} = 2.5$, $n = 30$). (c) Kinematic analysis of 14 fault data with reverse sense of motion collected at Wangaloa, southern Otago (Paleocene Wangaloa Formation, site 29). P-axis and T-axis are axes of maximum shortening and maximum extension, respectively, and are marked by solid rectangles and the letters P and T. Fault plane solution is of thrust mechanism, striking NNE–SSW and dipping either ESE or WNW.

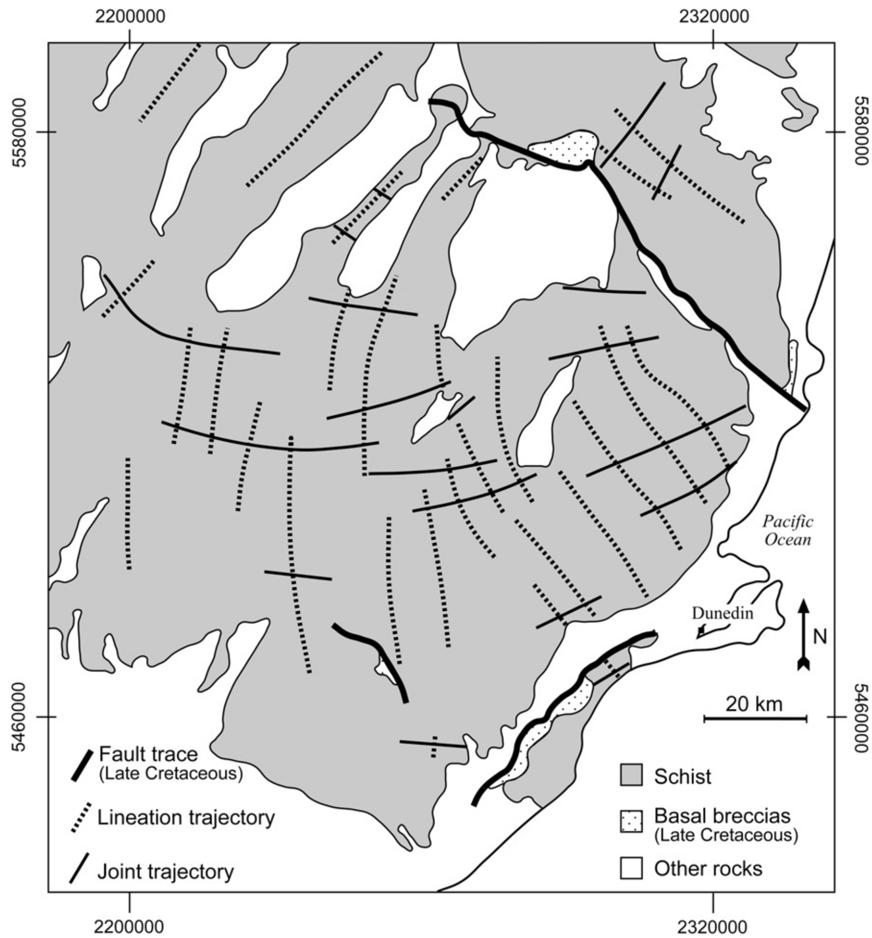

Fig. 10. Summary of Cretaceous region-wide structures in Otago. Trajectories of Late Cretaceous systematic joints (this study) are everywhere developed sub-perpendicular to trajectories of Early Cretaceous stretching lineations (Mortimer, 1993). The joint and lineation patterns are generalised and schematic, and hence are not presented in all exposures.

of GPS data (Beavan et al., 1999; Wallace et al., 2007). Hence, the expected strike of vertical joint formation across the entire Otago Schist (and indeed much of the South Island) during late Cenozoic tectonics is mainly WNW–ESE.

However, a Miocene 020–200° extension direction for development of the Otago Schist systematic joint sets in Fig. 6 is not supported because of the following: (1) WNW–ESE joints were observed only in a few sites, mainly in the Raggedy Range (sites 19, 20, 50, Fig. 6). Even in these sites the low values of θ (less than 11°; Supplementary Material, Item 1) suggests an association between the ductile and brittle deformation. (2) The 40° difference in the respective joint orientations between site 22 (Tunnel Beach, Miocene sandstone) and site 21 (Brighton, Otago Schist TZIII) is too large considering their close location, only a few kilometres apart (Fig. 6). (3) Some fossil tors are covered by Miocene sediments (Raeside, 1949), and many of the tor terminations consist of joint surfaces (Fig. 3b), suggesting that the joints formed before the Miocene sedimentation. (4) Late Cretaceous basal conglomerates and breccias that unconformably overlie the schists are not fractured. For example, a spectacular exposure near Palmerston (site 28, Fig. 2) shows that the systematic joints in the schists have no continuation in the overlying conglomerate of the Taratu Formation, even though the conglomerate is indurated and has a similar stiffness to the schist (Fig. 3h). Hence, it is our interpretation that the Otago Schist systematic joints formed before the development of the Late Cretaceous unconformities (certainly before the c. 85 Ma Taratu Formation, and probably also before the c. 112 Ma Kyeburn-Horse Range Formation; Forsyth, 2001; Tulloch et al., 2009), and are not due to Miocene to Recent deformation. This interpretation is consistent with the clustering of Otago Schist zircon fission track ages at $\sim$90 Ma (Tippett and Kamp, 1993) whose annealing temperatures of $\sim$250 °C indicate a brittle regime at geologically normal strain rates. A Late Cretaceous age of brittle deformation is further supported by the occurrence of $\sim$96 Ma pseudotachylyte veins in normal faults in TZIII schist near Alexandra (Barker, 2005).

The systematic joints in the Otago Schist formed after the rocks passed the brittle–ductile transition during Late Cretaceous exhumation. In many metamorphic terrains this transition is associated with kink folding (e.g., Carreras, 1997). Locally, jointing in the Otago Schist is kinematically associated with kink folds and quartz veining (Fig. 3k,m) indicating activity of hydrothermal systems during deformation; some joints have been inflated and filled with quartz. In most of these sites, the values of θ of the quartz-filled veins are quite similar to those of the systematic joints (e.g., site 12).

5.3. Mechanical aspects of schist jointing

The brittle mechanical behaviour of Otago Schist depends on the rock and stress anisotropies of the schist. Rock anisotropy resulted from fabric developed in the protoliths (i.e., formation of foliation, segregation and lineation), whereas any stress anisotropy would develop due to directional residual elastic strain preserved in the rocks during metamorphism. The formation of joints is strongly related to these intrinsic properties of the schist, and possibly also on the extrinsic loading (stress) conditions prevailing during exhumation. Nevertheless, joints would form only when the tension exceeded the tensile strength of the schist. Behrestaghi et al. (1996), Brosch et al. (2000), and Singh et al. (2001) show that schist is strongest parallel to foliation, and that the maximum tensile strength can be 3–4 times greater than the minimum tensile strength. These rock mechanics experiments might suggest an explanation for the formation of the less common 'sheet joints' parallel to foliation, rather than to the commonly observed foliation-perpendicular systematic joints in Otago Schist. Vernik and Zoback (1992) and Brosch et al. (2000) found no statistically significant difference between the tensile strengths parallel to and perpendicular to the lineation. Hence these rock mechanics experiments provide no explanation for the observed lineation-perpendicular systematic joints in Otago Schist.

Quartz rods are common in TZIII and TZIV schists and form a large fraction of the schist volume. From a mechanical perspective, rod-rich schists are composite-like materials (i.e., multiphase materials), in which the stiff quartz rods are aligned within a softer mica-bearing matrix (Fig. 3i). Assuming that both the matrix and the quartz rods behave elastically, and that the quartz rods are continuous fibers within the matrix, then the mechanical properties of the composite schist depend on the tensile stresses of the matrix (σ_m) and the rods (σ_r), the volume fraction of matrix (V_m) and rods (V_r), the type of rod arrangement and their relative alignment with respect to the direction of the applied tension (Matthews and Rawlings, 2000). In longitudinal loading, the tensile strength of the composite material (σ_c) is equal to the sum of two products that signifies the relative contribution of the matrix/fibers to the bulk stress (i.e., $\sigma_c = \sigma_m.V_m + \sigma_r.V_r$). In transverse loading, $\sigma_c = \sigma_m$. Representative values of the tensile strength of quartz (approximated by either quartzite or fused quartz) are ~25–50 MPa (Bieniawski, 1984; Matthews and Rawlings, 2000) and that of biotite–quartz–mica schists 10–15 MPa (Singh et al., 2001; measured parallel to foliation). The volume fraction of quartz rods in rod-rich TZIV Otago Schist is up to 0.5 (e.g., Fig. 3j). For $\sigma_m = 15$ MPa, $\sigma_r = 50$ MPa, $V_m = 0.5$ and $V_r = 0.5$, the tensile strength of the schist parallel to the quartz rods is ~1.5–3 times greater than the strength perpendicular to the rods. Hence, the quartz rods reinforce schist strength parallel to the rods, and result in strength anisotropy in the plane of foliation. Based on this analysis, the existence of the quartz rods should inhibit the formation of joints perpendicular to them, in contrast to what is observed in the Otago Schist where joints are oriented perpendicular to the quartz rods. Noticeably, the tensile strength of the schist, either parallel or perpendicular to the quartz rods (30 MPa and 15 MPa, respectively), is lower than the 50 MPa of residual elastic stress calculated in deformed natural quartz based on direct strain measurements (Kunz et al., 2009). Another possibility is that quartz rods in the schist might behave as "Eshelby inclusions" (Eshelby, 1957), i.e., stiff elastic inclusions enclosed in a compliant elastic matrix, which tend to concentrate the applied stress and fracture while the matrix is preserved intact (e.g., Eidelman and Reches, 1992). However, individual joints continuously cut the matrix and the numerous embedded quartz rods, and show no preference of forming only within the rods.

The mechanical aspects considered above indicate that the physical anisotropy of the schist did not enhance, and actually would have been expected to inhibit the formation of joints perpendicular to the schist fabrics. Hence, we suggest a control on joint formation by stress anisotropy developed due to directional residual elastic strain preserved in the rocks following ductile deformation (Fig. 11). A few other workers (e.g., Price, 1966; Engelder, 1985) have previously proposed that joints arise during uplift of a region from release of residual elastic strain energy

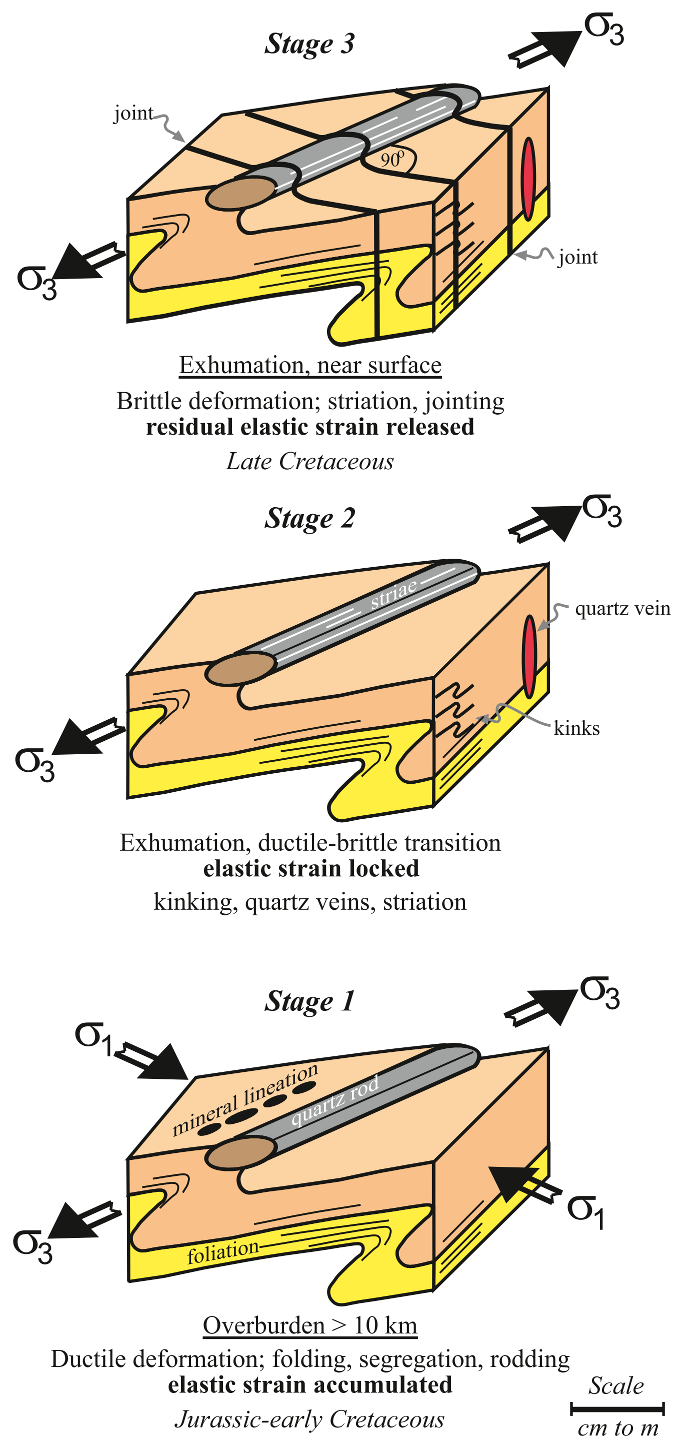

Fig. 11. Schematic diagrams showing conceptual model for schist deformation and systematic joint set formation in the Otago Schist. σ_1, maximum compressive principal stress (maximum shortening); σ_3, least compressive principal stress (maximum extension).

preserved in rocks. Residual elastic strains related to Mesozoic and possibly Precambrian tectonic events have even been measured (Eisbacher and Bielenstein, 1971; Friedman, 1972).

5.4. Conceptual model for schist jointing

Our model spans the time of metamorphism to the end of jointing (Fig. 11). Ductile folding during the metamorphism resulted in, among other features, isoclinal folds and refolded folds (Fig. 11, stage 1). Along parts of the fold hinges quartz crystallised as elongated rods. The trend of the quartz rods represents syn-metamorphic extension along the least compressive principal stress (maximum extension) directions. Corrugations on quartz segregations and flattened veins are kinematically linked with this penetrative strain. As metamorphism waned, just before the rocks ceased to deform in a totally ductile manner, local kink folds were formed (Fig. 11, stage 2; note these are different from the F4 kinks of Craw (1985) which are clearly related to late Cenozoic deformation around faults).

As the schist rock mass was exhumed into the upper crust, lithostatic pressure declined, the rocks cooled, and the schist rheology changed from ductile to brittle. At this stage the residual elastic strain energy could be released by tensile fracturing of the schists. First, small-scale kink folds (Fig. 11, stage 2) were formed, along with quartz veins in regions of local hydrothermal systems. Consequently (Fig. 11, stage 3), the large-scale systematic joint sets were formed along with small-scale joints along kink folds. The orientation of the joints did not develop at random, because it was dictated by the strain (stress) anisotropy preserved in the schists. Petrofabric studies (e.g., Turner, 1938; Cooper, 1995; Maxelon, 1999) show that quartz preferred orientations are widely preserved across the Otago Schist, thus demonstrating that preserved residual elastic strain in the schist mass is feasible.

6. Conclusions

Systematic joints in the Otago Schist are common in the greenschist facies, less common in pumpellyite–actinolite facies, and absent in the flanking prehnite–pumpellyite facies greywackes. Most commonly, the major systematic set of joints is present in the schists and it is sub-perpendicular to foliation, and to prominent lineations. Sometimes two sets of joints are developed with lineation bisecting the acute angle between the joints. These angular relationships hold irrespective of the dip and trend of foliation and lineation. The systematic joints developed in the Late Cretaceous, based on lack of joint continuation into Late Cretaceous conglomerates that unconformably overlie jointed schists, consistent orthogonality of joints with foliation and lineation, cooling history, and lack of relationship of systematic joints to late Cenozoic plate-boundary features. Mechanical aspects of schist jointing indicate that the physical anisotropy of the schist alone did not enhance the formation of joints perpendicular to the schist fabric. We therefore conclude that the systematic joints have formed during exhumation of the schist due to release of residual elastic strain energy inherited from earlier Early Cretaceous ductile deformation.

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Appendix. Supplementary data

Supplementary data associated with this article can be found in the online version, at doi:10.1016/j.jsg.2009.12.003

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