Chapter 2
Regional Geological Setting of the Otago Schist

2.1 Mesozoic Accretion

The basement rocks of the South Island of New Zealand consist of a diverse set of terranes that accumulated on the Pacific margin of Gondwana during the Mesozoic (Fig. 2.1a–c; Mortimer 2004). The Otago Schist belt that hosts the Macraes mine (Fig. 2.1a, b), formed as part of this Mesozoic convergence (Fig. 2.1c). At that time, what is now the New Zealand landmass and associated submerged continental crust was attached to the edge of the Australian portion of Gondwana (Mortimer 2004). This Pacific-facing continental margin was an elongated convergent tectonic boundary with active subduction, continental arc magmatism, and terrane collision (Fig. 2.1c). The Otago Schist formed as part of a wide (>100 km) accretionary complex above the subduction zone (Fig. 2.1b, c).

The accretionary complex includes two principal metasedimentary terranes, Caples and Torlesse terranes (Fig. 2.1b, c) as well as several other subordinate terranes within the complex (Mortimer 2004; MacKenzie and Craw 2005; Cooper and Ireland 2013). Terrane accretion resulted in stacking, thickening, and pervasive deformation in the metasedimentary rocks that accumulated above the subduction zone (Fig. 2.1c). These processes led to formation of a thick metamorphic pile, which became the Otago Schist belt, primarily during the Jurassic and Early Cretaceous (Fig. 2.1a–c; Mortimer 2004; Gray and Foster 2004). Terrane boundaries, and the Caples-Torlesse terrane boundary in particular, have been deformed and metamorphosed within the metamorphic pile (Fig. 2.1b).

Subduction-related Mesozoic arc magmatism occurred in what is now the far west of the South Island, to form the Median Batholith (Fig. 2.1c; Mortimer 2004). A notable feature of the Mesozoic plate boundary is the great width of the accretionary complex, which distanced metasedimentary accretion from the subduction-related arc magmatism (Fig. 2.1c). This has resulted in clear separation of magmatic

© The Author(s) 2016

D. Craw and D. MacKenzie, *Macraes Orogenic Gold Deposit (New Zealand)*,
SpringerBriefs in World Mineral Deposits, DOI 10.1007/978-3-319-35158-2_2

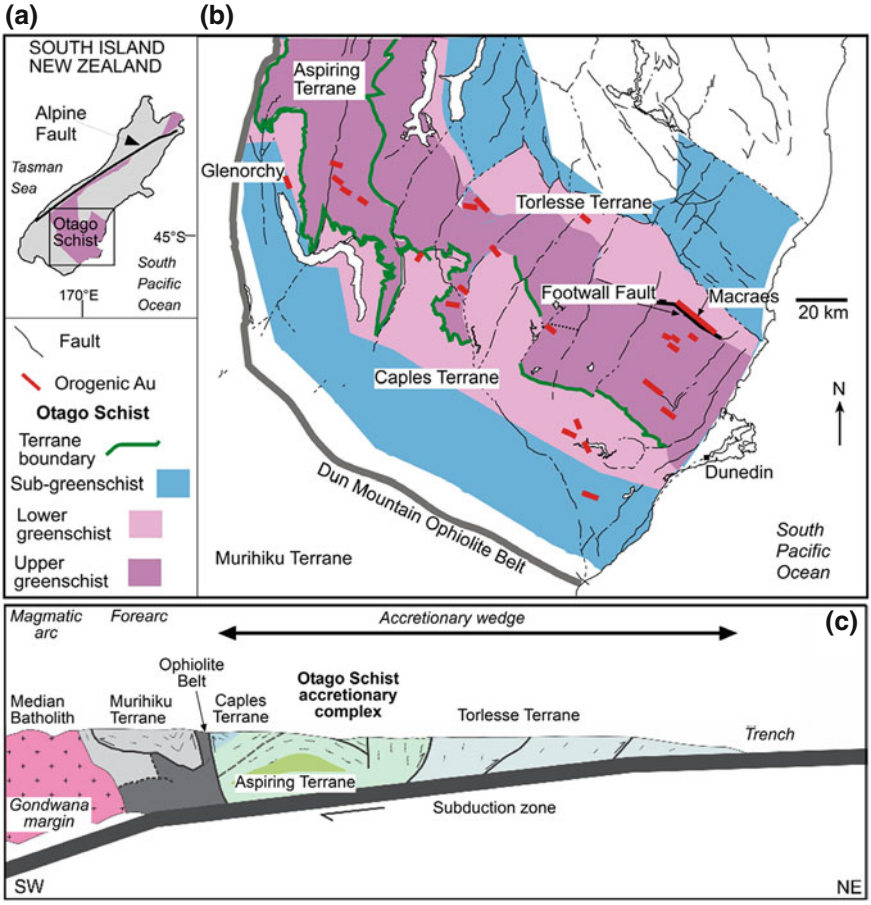


Fig. 2.1 Location of the Macraes mine within the Otago Schist as part of a Mesozoic accretionary complex. **a** Otago Schist belt in the South Island of New Zealand. **b** Metamorphic and terrane map of the Otago Schist belt. **c** Reconstruction of the Early Cretaceous plate tectonic geometry of the Pacific margin of Gondwana (modified from Mortimer et al. 2016)

processes from the accretionary complex. This lack of syn-metamorphic magmatism in the accretionary complex is a distinct feature of the Otago Schist belt that is not found in most other Phanerozoic gold-bearing metasedimentary terranes.

2.2 Otago Schist Lithology and Structure

The schist protoliths on both sides of the belt are dominated by greywacke (lithic sandstone) with interlayered argillites in thick turbidite sequences that have poorly developed stratigraphy (Fig. 2.2a; Mackinnon 1983; Mortimer 2004). The Torlesse

Terrane rocks (Fig. 2.2a) are generally more quartzofeldspathic than the volcanogenic Caples Terrane rocks, but these distinctions are subtle. Hence, the whole schist belt is made up of a compositionally uniform rock mass that differs mainly in metamorphic grade and degree of recrystallisation, and primary lithological variations are subordinate. Minor interlayered mafic horizons (1–100 m scale) occur throughout the belt, and these are more abundant in the Aspiring Terrane that is exposed on the western side of the belt (Fig. 2.1b). These mafic horizons do not make up more than 10 % of the schist pile locally, and are almost completely absent from the lower greenschist facies rocks near to the Macraes mine.

Polyphase recumbent folding, with associated fold axial surface cleavage development, has created pervasive flat-lying foliation that dominates much of the schist belt (Fig. 2.2b). Subsequent differential exhumation has formed a broad antiformal structure in the flat-lying foliation, with the highest grade rocks (upper greenschist facies) in the core of the antiform (Fig. 2.1b, c). Metamorphic grade decreases from the core to the northeast and southwest, where original sedimentary features are recognisable in the low grade (sub-greenschist facies) portions of the metasedimentary terranes. Boundaries between metamorphic zones are now extensively faulted, disrupting what were originally more continuous metamorphic transitions on both sides of the belt (Mortimer 2000; Henne and Craw 2012).

Greenschist facies rocks are fully recrystallised, so that no bedding or clastic textures are preserved. Syn-metamorphic quartzofeldpathic segregations, including

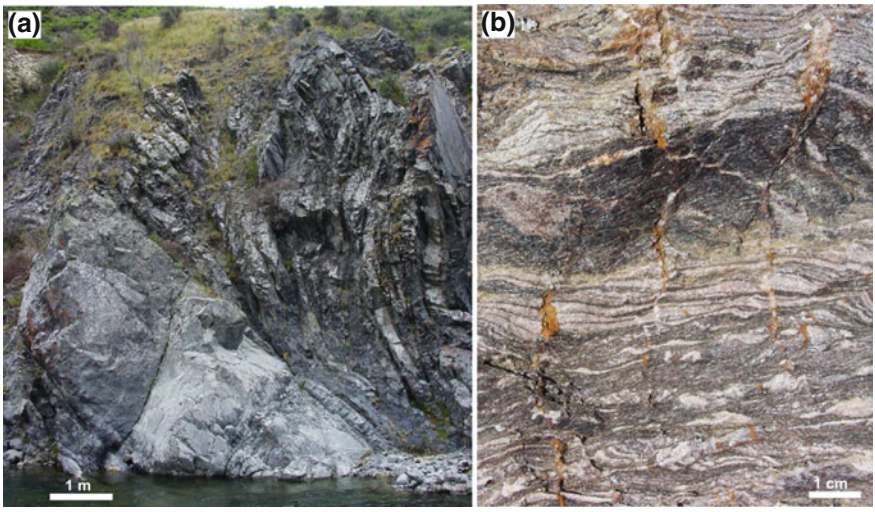


Fig. 2.2 Otago Schist structure and lithology. **a** Outcrop of low grade (sub-greenschist facies) Torlesse Terrane protolith dominated by steeply-dipping bedding that has been folded and steepened. Paler and resistant beds are greywacke (sandstone); recessive darker beds are argillite. **b** Close view of lower greenschist facies schist that is dominated by pervasive flat-lying foliation and metamorphic segregations. Remnants of more and less micaceous layers (*dark* vs. *light*, respectively) reflect original lithological variations in the protolith

some veins, have developed in greenschist facies rocks and dominate the rock fabric at the 1–10 cm scale (Fig. 2.2b). Segregations are separated by micaceous laminae containing quartz and muscovite (Fig. 2.2b). The quartzofeldspathic segregations have either formed parallel to the pervasive foliation or have been rotated into near-parallelism with that fabric during progressive ductile deformation. The development of these segregations has further obscured the subtle lithological variations within the schist, although more and less micaceous horizons are still recognisable on the 0.1–10 m scale (Fig. 2.2b).

2.3 Uplift, Exhumation, and Extension

Early-formed parts of the accretionary complex were metamorphosed and underwent uplift and exhumation while the more distal parts of the complex were still accreting. Hence, debris eroded from the uplifted portions contributed to the accreting sedimentary pile in the Jurassic and Cretaceous (Mortimer 2004; Cooper and Ireland 2013). At the same time, topographically uplifted parts of the evolving schist belt were undergoing extension while the deeper parts were still being compressively deformed (Mortimer et al. 2016). In the currently-preserved schist belt, it was the lower-grade flanks that were exhumed first, in the late Jurassic and Early Cretaceous (Gray and Foster 2004). The rocks that now constitute the higher-grade core of the belt continued to undergo ductile deformation into the middle Cretaceous, after what are now the lower grade flanks of the belt had been exhumed (Gray and Foster 2004).

Localised extension in the shallower parts of the accretionary complex in the Early Cretaceous evolved to more regional scale extension as the plate tectonic setting changed in the middle and Late Cretaceous. During this time, subduction ceased and extensional breakup of Gondwana became the dominant tectonic process (Mortimer 2004). The regional extension culminated in the separation of New Zealand continental crust, that now included the accretionary Otago Schist, from the Australian continent in the Late Cretaceous (Mortimer 2004). This regional extension led to development of a network of normal faults across the exhumed Otago Schist belt, with extensive erosion of faulted topography resulting in localised thick sedimentary accumulations in the middle Cretaceous (Craw 2010). Some of the normal faults were shallow-dipping, and these contributed to exhumation of the higher-grade core to form the antiformal structure that now dominates the geometry of the belt (Fig. 2.1b; Deckert et al. 2002).

Regional extension continued through the early Cenozoic, and the exhumed schist belt subsided beneath the sea until the late Oligocene or early Miocene (Landis et al. 2008). Reconfiguration of the plate tectonic setting in the SW Pacific Ocean at that time was marked in the South Island by development of the transpressional Alpine Fault (Fig. 2.1a). Regional deformation associated with this new plate boundary caused reactivation of many of the normal faults as reverse structures. This deformation resulted in uplift and renewed erosion across the schist belt.

The uplift and erosion is most pronounced near to the Alpine Fault, where a range of high mountains, the Southern Alps, has formed on the eastern side of that fault. The rejuvenated topography becomes progressively more subdued farther east, where much of the terrain is dominated by an exhumed and deformed Late Cretaceous-middle Cenozoic basement unconformity surface (Landis et al. 2008). In the vicinity of the Macraes mine, the topography is dominated by this unconformity surface that forms broad uplands incised by narrow steep streams (Fig. 1.4).

2.4 Hyde-Macraes Shear Zone

The Hyde-Macraes Shear Zone, which hosts the Macraes mine, is a major regional structure within the eastern part of the schist belt (Figs. 2.1b, 2.3a–c). The shear zone formed in lower greenschist facies schists as they were being uplifted through the brittle-ductile transition (Craw et al. 1999). Hence, early development of the shear zone was dominated by ductile features, including folding, mineral recrystallisation and reorientation, and mylonitic deformation (Craw et al. 1999; Petrie et al. 2005). The ductile features were progressively overprinted by more brittle compressive structures, including cataclasites and fault gouges, (Craw et al. 1999). The shear zone is a shallow-dipping thrust structure that is sub-parallel to the shallow-dipping metamorphic foliation on the northeastern limb of the regional schist antiform (Fig. 2.3c).

The earliest deformation in the shear zone occurred under lower greenschist facies metamorphic conditions, with recrystallised mineralogy similar to that of the host rocks. This deformation was accompanied by hydrothermal fluid flow and gold mineralisation. Fluid flow was pervasive, along grain boundaries and microshears, through large volumes of rock (Fig. 2.3b, c). The structural upper limit of this shear-hosted fluid flow is as yet poorly defined, although the shear zone is variably mineralised over several hundred metres thickness in places (Fig. 2.3b). As the rocks became more brittle, fluid flow and mineralisation processes continued, with progressively lower temperature mineralogy prevailing, and fluid flow became more focussed in cataclastic zones and fractures, still over a large thickness of rock.

The structural base of the shear zone is now defined by a prominent but narrow post-mineralisation normal fault, the Footwall Fault (Fig. 2.3a–c). This extensional structure has juxtaposed lower greenschist facies rocks, which host the mineralised shear zone, on to unmineralised upper greenschist facies rocks (Fig. 2.3a–c). The Footwall Fault is part of the regional set of extensional structures that was initiated in the middle Cretaceous, and has contributed to exhumation of the higher-grade schist core (Deckert et al. 2002; Craw 2010). Because of this post-mineralisation truncation of the lower part of the Hyde-Macraes Shear Zone, it is not possible to determine the original shear zone thickness.

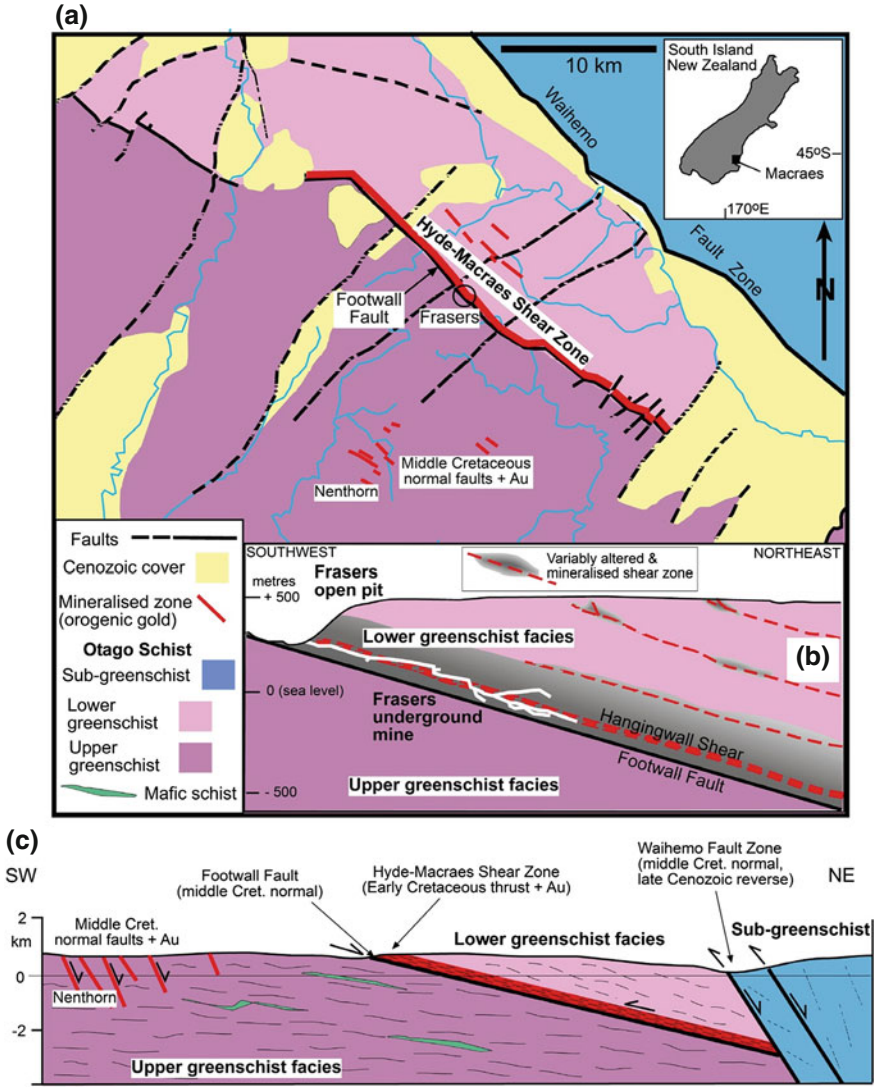


Fig. 2.3 Geological setting of the Macraes mine in the Hyde-Macraes Shear Zone. **a** Geological map of the Macraes area. **b** Cross section through the Frasers open pit and underground mine. **c** Regional cross section across the Hyde-Macraes Shear Zone near to the Frasers open pit

2.5 Mesozoic Geochronology

There were two distinct pulses of orogenic gold mineralisation (Mortensen et al. 2010) during the Cretaceous evolution of the Otago Schist belt: Early Cretaceous (135–140 Ma) and middle Cretaceous (110–100 Ma). The earlier event, which

included Macraes, was intimately linked to the latter stages of metamorphism of the schist, and the later event was linked to regional extensional processes (Fig. 2.4a). The lower greenschist facies rocks on the flanks of the schist belt were being uplifted through the brittle-ductile transition while the upper greenschist facies core of the belt remained ductile and was still undergoing high grade metamorphism (Fig. 2.4a).

The relative timing of the specific structural features relevant to formation and deformation of the Hyde-Macraes Shear Zone has been determined radiometrically on available rocks, and these data are summarised in Fig. 2.4b. The lower greenschist facies host rocks for the shear zone were exhumed and cooled below argon and strontium closure temperatures (between 300 and 400 °C) in the Late Jurassic (160–140 Ma). Widespread recumbent ductile folds of early-formed foliation in these rocks were formed before this exhumation and cooling, and pre-date the formation of the shear zone. Rocks of the shear zone itself recrystallised and cooled by ~130 Ma, and the shear zone structures were superimposed on the pre-mineralisation ductile folds (Fig. 2.4b). In particular, precise $^{40}Ar/^{39}Ar$ dating of hydrothermal muscovite from the Macraes deposit suggests that mineralisation occurred at 135–136 Ma (Mortensen et al. 2010). These geochronological results are in accord with structural and mineralogical inferences that suggest that the shear zone formed in the latter stages of exhumation of the hosting lower greenschist facies rocks.

In contrast, the underlying upper greenschist facies rocks did not cool below argon closure temperatures until at least 120 Ma (Gray and Foster 2004). These rocks remained ductile, with isoclinal folding and associated muscovite recrystallisation, after the Hyde-Macraes Shear Zone had formed at shallower structural levels (Fig. 2.4b). These relatively young ductile fold structures were superimposed on at least one generation of recumbent ductile folds that are temporally equivalent to the pre-mineralisation folds in the lower greenschist facies rocks (Fig. 2.4b). The young ductile folds in the upper greenschist facies rocks are at least partially the deeper-level expression of the same compressional deformation that formed the Hyde-Macraes Shear Zone as a thrust structure near to the brittle-ductile transition in the Early-middle Cretaceous.

The Footwall Fault and associated extensional structures in the Macraes area were initiated at ~112 Ma (Tulloch et al. 2009; Craw 2010), after cessation of mineralisation in the Hyde-Macraes Shear Zone. Steeper gold-bearing normal faults in the upper greenschist facies core of the schist belt, including the Macraes area (Fig. 2.3a, c), formed in a second pulse of gold mineralisation in the middle to Late Cretaceous (as young as ~100 Ma; Fig. 2.4b, Mortensen et al. 2010). These faults are entirely brittle extensional structures, and are genetically unrelated to the hydrothermal system of the much older Hyde-Macraes Shear Zone (Fig. 2.4a).

The Early Cretaceous gold mineralisation at Macraes was coeval with formation of a regional scale swarm of extensional quartz veins in the Glenorchy area (Fig. 2.1a; Mortensen et al. 2010). Some of these veins fill normal faults and have been mineralised, primarily with scheelite and minor gold (Hay and Craw 1993; Begbie and Sibson 2006). Host rocks are lower greenschist facies and

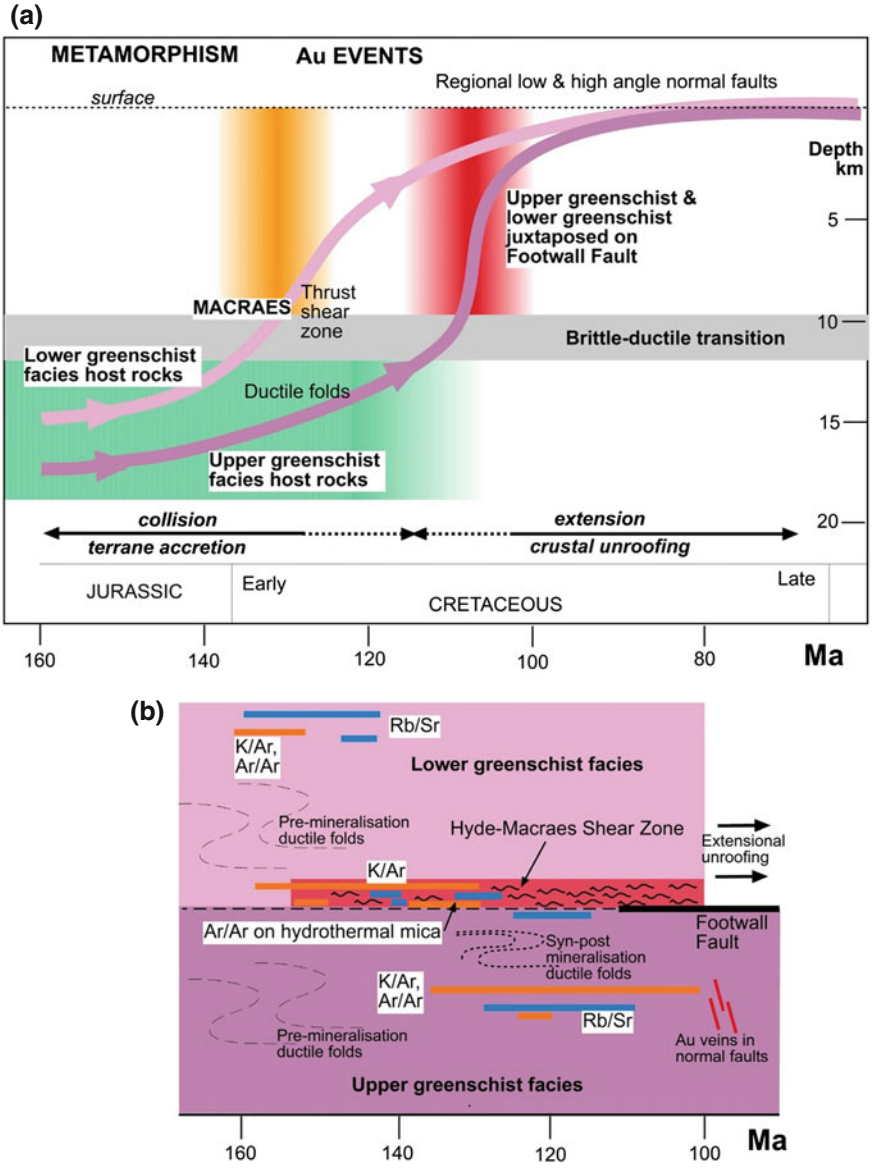


Fig. 2.4 Schematic representations of relative timing of geological events and orogenic gold mineralisation in the Otago Schist belt (modified from Mortensen et al. 2010). **a** The early pulse of mineralisation was associated with the latter stages of greenschist facies metamorphism, and the later stage of mineralisation was associated with regional extension The Macraes deposit formed during differential uplift of the lower and upper greenschist facies rocks. **b** Summary of geochronology associated with the Hyde-Macraes Shear Zone and the relative ages of various structures in the surrounding rocks (from Adams and Graham 1997; Craw 2002; Gray and Foster 2004; Mortensen et al. 2010)

sub-greenschists facies Caples Terrane schists that are broadly similar to those at Macraes. Mineralisation was a late metamorphic process, as at Macraes, but vein emplacement occurred at a higher structural level above the brittle-ductile transition.

2.6 Cenozoic Geological History of Macraes Area

Tectonic extension in the Macraes area continued through the Early Cenozoic, resulting in subsidence and erosion to a low relief land mass in the area. Numerous normal faults formed an orthogonal grid of northwest and northeast striking structures that disrupted the basement structures (Fig. 2.1). In particular, a set of northeast striking structures cut across the Hyde-Macraes Shear Zone, locally offsetting it on a scale of tens to hundreds of metres (Fig. 2.5a, b). These cross-faults may have been partly controlled by pre-existing structures in the shear zone itself.

A thin veneer of fluvial sediments accumulated locally, primarily in faulted depressions. Some of these sediments contained placer gold derived from erosion of orogenic deposits in the basement, possibly including the Macraes deposit. Groundwater-driven alteration of the basement rocks underlying the sediments resulted in a clay-altered and variably oxidised zone that extended up to 100 m below the basal sediment unconformity. This altered zone was readily eroded as on-going fault-related deformation exposed the basement on fault scarps. Quartz-rich sediments and placer gold were progressively recycled into younger deposits. Remnants of quartz-rich Eocene fluvial sediments are still preserved near the Macraes mine, and represent the end result of recycling processes that started in the middle Cretaceous (Craw 2010).

The groundwater-driven basement alteration beneath the sedimentary veneer caused oxidation and localised supergene enrichment of gold in orogenic deposits in the basement, including the Macraes deposit (Craw et al. 2015). The supergene gold has coarser particle size (Fig. 2.5c) than the primary hydrothermal gold at Macraes, which is typically micron scale particles enclosed in sulphide minerals. The supergene gold is also highly irregular in shape, and is distinctly different from the flakes of placer gold that have been recycled from older sediments (Fig. 2.5c).

Tectonic and topographic rejuvenation of the Macraes area in the Late Cenozoic occurred mainly via reactivation of the pre-existing normal faults, including the cross-faults that cut the Hyde-Macraes Shear Zone. These faults were reactivated as reverse faults, again with offsets on a scale of tens to hundreds of metres, including the prominent Macraes Fault (Fig. 2.5a, b). Alkaline basaltic volcanism accompanied the onset of late Cenozoic rejuvenation of the area, mainly in the early Miocene. Lava flows associated with this volcanism contribute to the thin veneer of young cover rocks, and their relative competence have protected the veneer from erosion in places. Some open pit mining has cut through the volcanic veneer to mineralised rocks below.

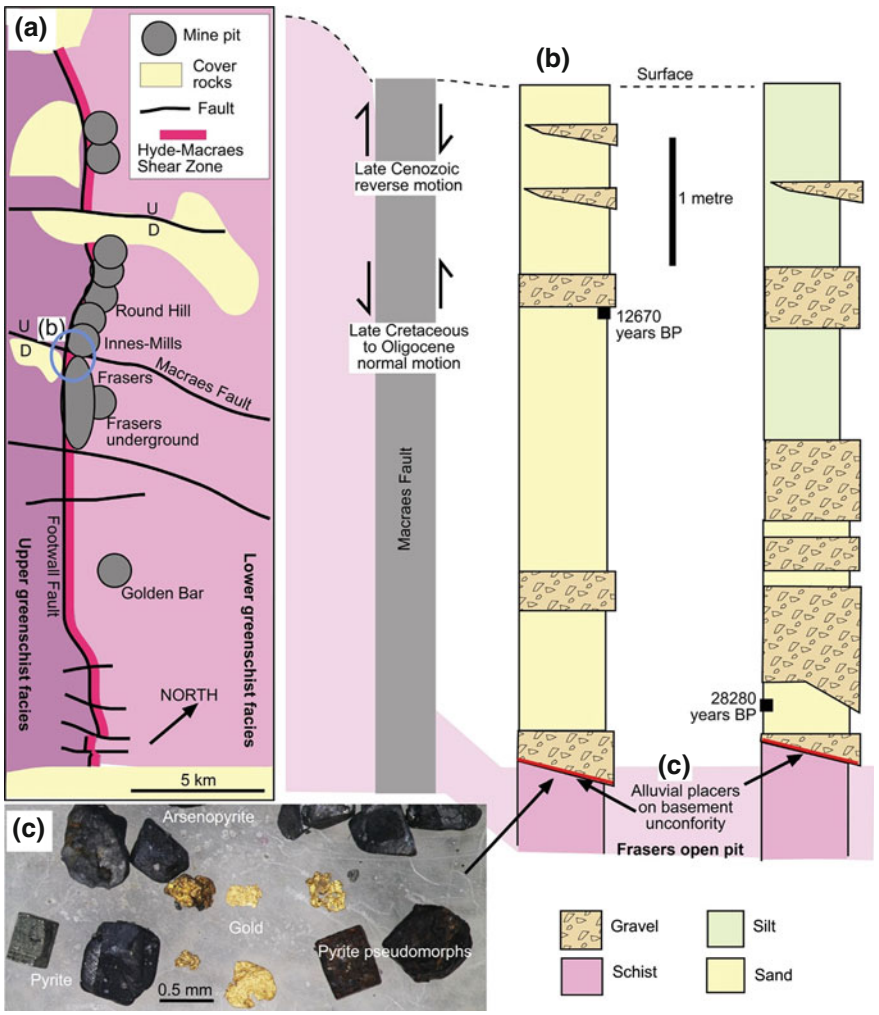


Fig. 2.5 Cross-faults cutting the Hyde-Macraes Shear Zone. **a** Sketch map showing the principal cross-faults, including the Macraes Fault. **b** Schematic section across the Macraes Fault, with the Late Pleistocene sediments (carbon dates indicated) that were on top of Frasers Pit, on the downthrown side of the fault (modified from Craw and Chappell 1999). **c** Detrital gold, pyrite, and arsenopyrite from the base of the Late Pleistocene sediments in (**b**). Gold flake at bottom centre is recycled from Eocene placers, whereas the other irregular gold particles are of supergene origin in the Hyde-Macraes Shear Zone

Cross-fault reactivation in the Macraes area was a Late Pleistocene effect, as indicated by radiocarbon dating of sediments that were deposited on what is now the Frasers open pit (Fig. 2.5b). These sediments were locally derived, from the rising reversed fault scarp nearby, and include some supergene gold, pyrite, and arsenopyrite, and their oxidised pseudomorphs, as well as some flaky gold derived

from nearby Eocene sediments. Placer gold in the Late Pleistocene sediments (Fig. 2.5b, c) was mined for a short period as Frasers open pit was being excavated in the late 1990s. Similar gold-bearing sediments on the downthrown side of the Macraes Fault were mined historically immediately to the southwest of what is now Frasers open pit (Fig. 2.5a).

2.7 Absence of Syn-mineralisation Magmatism

There is no field evidence for magmatic intrusion into the metasediments that host the Hyde-Macraes Shear Zone during the Late Jurassic to Early Cretaceous metamorphism of the Otago Schist belt. The shear zone was immediately underlain by a thick pile of upper greenschist facies rocks that were exhumed by the Footwall Fault, and there are no intrusive rocks in this higher grade core of the schist belt in the Macraes area. Further, the Aspiring Terrane underlies at least some of the upper greenschist facies schist package and this has been exhumed to deeper levels than most of the schist belt by Cenozoic uplift and erosion in the western mountains (Fig. 2.1). Even this deeper structural level has no evidence for igneous intrusion into the schist belt. Xenoliths in Miocene alkaline basalts erupted in the Macraes area consist of metamorphic and mantle rocks, but no igneous rocks from intrusions at depth.

The first evidence for post-metamorphic magmatism in the accretionary complex, in the middle Cretaceous, occurs as small remnants of silicic tuff and ignimbrite (112 Ma) associated with regional extensional structures related to the Footwall Fault (Tulloch et al. 2009) that were emplaced >20 million years after the shear zone was mineralised (Mortensen et al. 2010). Magmatism that was approximately coeval with Early Cretaceous shear zone mineralisation did occur in the Median Batholith to the west of the accretionary complex (Tulloch et al. 2009), but the nearest intrusions were >150 km away from Macraes, in entirely different tectonic setting in different terranes (Fig. 2.1).

References

Adams CJ, Graham IJ (1997) Age of metamorphism of Otago Schist in eastern Otago and determination of protoliths from initial strontium isotope characteristics: NZ J Geol Geophys 40: 275-286

Begbie MJ, Sibson RH (2006) Structural controls on the development of fault-hosted scheelite-gold mineralization in Glenorchy, northwest Otago: In Geology and Exploration of New Zealand Mineral Deposits, Christie AB, Brathwaite RL (Eds) Aust Inst Min Metall Monog 25: 291–298

Cooper AF, Ireland T (2013) Cretaceous sedimentation and metamorphism of the western Alpine Schist protoliths associated with the Pounamu Ultramafic Belt, Westland, New Zealand. NZ J Geol Geophys 56: 188-199

Craw D (2002) Geochemistry of late metamorphic hydrothermal alteration and graphitisation of host rock, Macraes gold mine, Otago Schist, New Zealand. Chem Geol 191: 257-275

Craw D (2010) Delayed accumulation of placers during exhumation of orogenic gold in southern New Zealand. Ore Geol Rev 37: 224-235

Craw D, Chappell DA (1999) Evolution and sulphide mineral occurrences of an incipient nonmarine sedimentary basin, New Zealand. Sed Geol 129: 37-50

Craw D, Windle SJ, Angus PV (1999) Gold mineralisation without quartz veins in a ductile-brittle shear zone, Macraes Mine, Otago Schist, New Zealand. Mineral Deposita 34: 382-394.

Craw D, MacKenzie DJ, Grieve P (2015) Supergene gold mobility in orogenic gold deposits, Otago Schist, New Zealand. NZ J Geol Geophys 58: 123-136

Deckert H, Ring U, Mortimer N (2002) Tectonic significance of Cretaceous bivergent extensional shear zones in the Torlesse accretionary wedge, central Otago Schist, New Zealand. NZ J Geol Geophys 45: 537-547

Gray DR, Foster DA (2004) $^{40}Ar/^{39}Ar$ thermochronologic constraints on deformation, metamorphism and cooling/exhumation of a Mesozoic accretionary wedge, Otago Schist, New Zealand. Tectonophysics 385: 181-210

Hay, R. and Craw, D. (1993). Syn-metamorphic gold mineralisation, Invincible Vein, NW Otago Schist, New Zealand. Mineral Deposita 28: 90-98

Henne A, Craw D (2012) Synmetamorphic carbon mobility and graphite enrichment in metaturbidites as a precursor to orogenic gold mineralization, Otago Schist, New Zealand. Mineral Deposita 47: 781-797

Landis CA, Campbell HJ, Begg JG, Mildenhall DC, Paterson AM, Trewick SA (2008) The Waipounamu erosion surface: Questioning the antiquity of the New Zealand land surface and terrestrial fauna and flora, GeolMag 145: 173-197

MacKenzie DJ, Craw D (2005) Structural and lithological continuity and discontinuity in the Otago Schist, Central Otago, New Zealand. NZ J Geol Geophys 48: 279-293

Mackinnon TC (1983) Origin of Torlesse and related rocks, South Island, New Zealand. Geol Soc America Bull 94: 967-985

Mortensen JK, Craw D, MacKenzie DJ, Gabites JE, Ullrich T (2010) Age and origin of orogenic gold mineralisation in the Otago Schist belt, South Island, New Zealand: Constraints from lead isotope and $^{40}Ar/^{39}Ar$ dating studies. Econ Geol 105: 777-793

Mortimer N (2000) Metamorphic discontinuities in orogenic belts: example of the garnet-biotite-albite zone in the Otago Schist, New Zealand. Int J Earth Sci 89: 295-306

Mortimer N (2004) New Zealand's geological foundations. Gondwana Res 7: 261-272

Mortimer N, Craw D, MacKenzie DJ, Mortensen J, Ring U, Pitcairn I (2016) Cretaceous tectonics and regional gold mineralisation in the Otago Schist, New Zealand. Aust Inst Min Metall Monog (in press)

Petrie BS, Craw D, Ryan CG (2005) Geological controls on refractory ore in an orogenic gold deposit, Macraes mine, New Zealand. Mineral Deposita 40: 45-58

Tulloch AJ, Ramezani J, Mortimer N, Mortensen J, van den Bogaard P, Maas R (2009) Cretaceous felsic volcanism in New Zealand and Lord Howe Rise (Zealandia) as a precursor to final Gondwana breakup. Geol Soc London Special Paper 321: 89-118

http://www.springer.com/978-3-319-35157-5

Macraes Orogenic Gold Deposit (New Zealand)
Origin and Development of a World Class Gold Mine
Craw, D.; MacKenzie, D.
2016, XII, 127 p. 83 illus. in color., Softcover
ISBN: 978-3-319-35157-5