

MINERAL DEPOSITS OF THE STILLWATER COMPLEX

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ABSTRACT

Mineral deposits of the Stillwater Complex include modest resources of Ni-Cu disseminated to massive sulfides at the base of the complex, chromitite deposits of the Ultramafic series, the platinum-group element (PGE) deposit of the Johns–Manville (J-M) Reef, and a low-grade Pt-Pd deposit known as the Picket Pin deposit—the latter two hosted in the Banded series. To date, the only deposit to be commercially mined without government support is the J-M Reef. A review of the proposed models to explain the concentration mechanisms that formed these deposits ranges from orthomagmatic interpretations involving direct precipitation from a magma to suggestions that igneous hydrothermal fluids both introduced the ore components and modified the mineral assemblage at the magmatic stage.

INTRODUCTION

There has been an explosion in layered igneous intrusion research since the publication of the Montana Bureau of Mines and Geology Special Volume 92 on the Geology of the Stillwater Complex over 35 years ago (Czamanske and Zientek, 1985). This interest follows from the fact that layered intrusions such as the Stillwater Complex preserve a record of how magmas crystallize and differentiate, but the interpretation of this record is proving to be deceptively difficult. This is particularly true for the understanding of the mechanisms that produced the impressive deposits of Ni-Cu, Cr, platinum-group elements (PGE), Ti, and V hosted in these intrusions. This report summarizes the features of the mineral resources of the Stillwater Complex and some of the current models that have been proposed for their formation. While hosting resources of Ni-Cu and Cr, it is only the Pd-Pt deposit of the Johns–Manville (J-M) Reef that has ever been successfully mined commercially, with current active mining operations by Sibanye–Stillwater.

EVOLUTION OF THOUGHT

Traditional thought on the formation of layered intrusions viewed them as the igneous equivalent of clastic sedimentary rocks. Modal layering (layering

defined by changes in the proportion of different minerals) was considered to result as crystals precipitate from the magma and either settle on top of a growing crystal + liquid mush at the floor of the magma chamber or are carried downward in density currents descending from the roof and walls. Mineral segregation was envisioned to happen owing to size and density differences between different minerals and their control on mineral settling velocity following Stoke's law. Furthermore, crystallization of magmas was largely assumed to follow a simple liquid line of descent, with perhaps some complication arising from injections of the primitive magma and mixing with a resident magma that had undergone some degree of differentiation.

However beginning in the 1960s and especially in the 1970s, evidence demonstrated that this simple model is untenable. For example, Jackson (1961) showed that coexisting olivine and chromite from chromitiferous dunites of the Stillwater Complex were not in hydraulic equilibrium. That is, although chromite is considerably denser than olivine, the olivine is so much larger that it would have a much higher settling velocity. Later work showed that plagioclase can have a neutral to positive buoyancy in many basaltic magmas (e.g., Campbell, 1978; McBirney and Noyes, 1979); this led to the apparent quandary that what is perhaps the most abundant “settled” mineral in many

intrusions instead should have floated. In addition, modal layering is observed to develop along vertical walls of magma chambers in which the effects of gravity would not be applicable, and instead suggests it is gradients in temperature and composition that drive the formation of modal layering.

Perhaps just as importantly, a number of new and improved techniques have been applied to the study of layered intrusions. These include improved precision in both bulk rock and microanalytical techniques, the increased application of a variety of isotopic systems, and sophisticated textural analysis. These studies have resulted in a variety of novel but often conflicting interpretations; what was once thought to be a rather straightforward problem of understanding how a magma crystallizes has grown increasingly uncertain and contentious. Do crystals settle, or do they settle as crystal-rich plumes of magma, or do they grow *in situ*, or are they carried into the chamber? Does the rock texture and composition evolve over time or do they reflect what was settling out from the magma? In part, this divergence of opinion arises from the open system nature of magma chambers. The question can be most broadly described now as a debate between two endmember models. The first calls on processes occurring within a crystallizing magma chamber and a growing crystal + liquid mush (including saturation in a volatile-rich fluid phase and zone-refining processes in the crystal mush). The second broadly interprets changes in the stratigraphic sequence to be the result of fresh injections of variably contaminated magmas or crystal-rich slurries that potentially could have been quite different from the magma that initially started to fill the chamber.

This debate is particularly contentious for the petrogenesis of ore deposits hosted in layered intrusions, which contain much of the world's chrome, Cu-Ni, and PGE reserves. Petrogenetic models range from orthomagmatic models, in which they are the result of magmatic crystallization, to hydromagmatic models that envision an important role for hydrothermal fluids in transporting ore components and their strong effects on changing the crystallization behavior of the magma. Studies on the Stillwater Complex and the host rocks have been particularly important in developing these models, especially for the Cr and PGE deposits, although consensus is still mostly lacking.

REGIONAL GEOLOGY

The Stillwater Complex is a late Archean (2.7 Ga; DePaolo and Wasserburg, 1979; Nunes and Tilton, 1971; Nunes, 1981; Premo and others, 1990; Wall and Scoates, 2016; Wall and others, 2018) layered ultramafic–mafic intrusion situated along the northern edge of the Beartooth Mountains. It has a maximum exposed stratigraphic thickness of 6.5 km and an exposed strike length of about 45 km (fig. 1). The Stillwater Complex and associated basement rocks are part of the Beartooth Archean terrane, one of several Archean-age fault-bound blocks that occur across Montana and Wyoming (e.g., Mogk and others, 1992). The Stillwater magma was emplaced at a depth of 6–7 km into a variety of ca. 3.27–2.74 Ga metasedimentary rock types, principally metamorphosed volcanoclastic shale, greywacke and breccia, quartzite, and banded iron formation (DePaolo and Wasserburg, 1979; Wooden and Mueller, 1988). The pre-intrusion metamorphic conditions are estimated to have been 525–550°C and 2 kbar (Helz, 1995; Labotka and Kath, 2001). Intrusion of the Stillwater magmas further heated the country rocks to produce a wide metamorphic aureole; the metapelite proximal to the complex has been thermally metamorphosed to hypersthene hornfels over a distance of 0.5–1 km from the contact, with a more distal cordierite–cummingtonite hornfels 0.5–1.5 km thick. Assemblages in the iron formation imply a peak metamorphic temperature around 825°C and pressures between 3 and 4 kbar (Labotka, 1985; Labotka and Kath, 2001). The Mouat quartz monzonite occurs along the eastern base of the complex and is exposed along the Stillwater River. Jones and others (1960) noted that that it was emplaced ~60 Ma after crystallization of the Stillwater and intrudes both the hornfels and locally the lower portion of the Stillwater Complex as a few small sills, whereas more recent dating by Wall and others (2018) implies that it is considerably older than the Stillwater Complex at 2.80 Ga. The section at the top of the intrusion is truncated by a pre-Middle Cambrian unconformity. Because of this, no evolved, magnetite-rich rocks (such as are observed in the uppermost few kilometers of the Bushveld Complex of South Africa) are seen in the exposed portion of the Stillwater Complex.

At least two major brittle structural events significantly modified the Stillwater Complex: (1) an older Precambrian (?) event(s) and (2) a younger (late Cretaceous to early Paleocene) collisional event (the Lar-

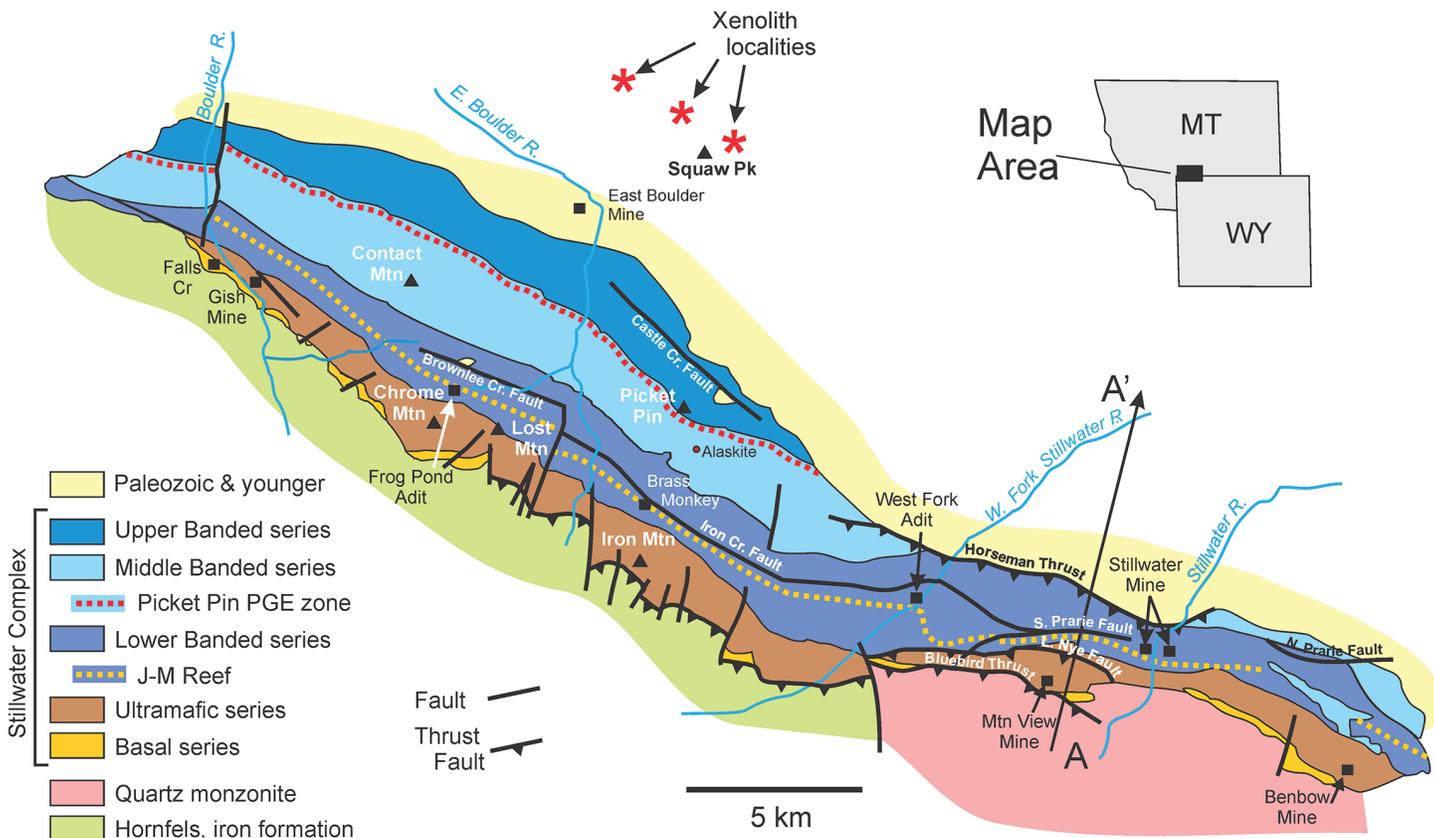


Figure 1. Geologic map of the Stillwater Complex showing major stratigraphic subdivisions, PGE reefs, and locations mentioned in the text. A–A', partial line of the cross section shown in figure 2. Redrawn after Jackson (1961) and Page and Zientek (1985), with permission.

amide Orogeny). Structural features associated with these major events have been studied through surface outcrop and underground drift mapping coupled with surface exploration drillholes and underground definition and development drillholes at both the Stillwater and East Boulder mines. The Precambrian event possibly occurred sometime in the Neoproterozoic to Mesoproterozoic. This brittle event resulted in a pronounced set of transverse faults and fractures oriented essentially north–south (30° east to 30° west of north) at high angle to west–northwest-trending layering in the complex (Geraghty, 2013), although some may be transfer structures between later Laramide thrust faults. While few of the dikes cutting the Stillwater Complex have been dated, north–south-trending diabase dikes in the eastern Beartooth Mountains have been dated at 2.0 to 2.1 Ga and dikes that trend consistently 30° west of north have been dated at 1.3 Ga (Mueller and Wooden, 1988). The dikes may have intruded a pre-existing structural grain. Thus, the prominent north–south faults and associated fractures may have had an Archean/Proterozoic time of formation. These faults are typically brecciated, commonly open, and vuggy. Rarely, some of the major regional faults have been reconstituted to chlorite schist. Some have vertical dip

when the Laramide tilting is back-rotated out, indicating a strike-slip parentage.

During the Laramide Orogeny, the Stillwater Complex and the Phanerozoic sedimentary section together were complexly deformed into Triangle Zone geometries complete with forethrusts, backthrusts, blind thrusts, frontal ramps, and lateral ramps (Geraghty, 2013; Thacker and others, 2017). The complex interaction between faults resulted in rotation and tilting of the Stillwater Complex to a moderately to steeply north- and northeast-dipping orientation, locally overturned to south (fig. 2). The tilting of the Stillwater Complex allows the originally horizontal mineralized zones to be mined while standing upright. The thrust and reverse faults associated with this deformation are typically gouge-filled, with pronounced tectonic fabric, in contrast to the brecciated Precambrian faults. The Laramide faults are mostly west–northwest trending with significant oblique slip. The uplift resulted in elevations in excess of 10,000 ft above sea level. The Beartooth plateau and Stillwater Complex have been deeply incised by the major drainages and tributaries of the Stillwater and Boulder Rivers down to elevations at the valley floor of approximately 5,000 ft.

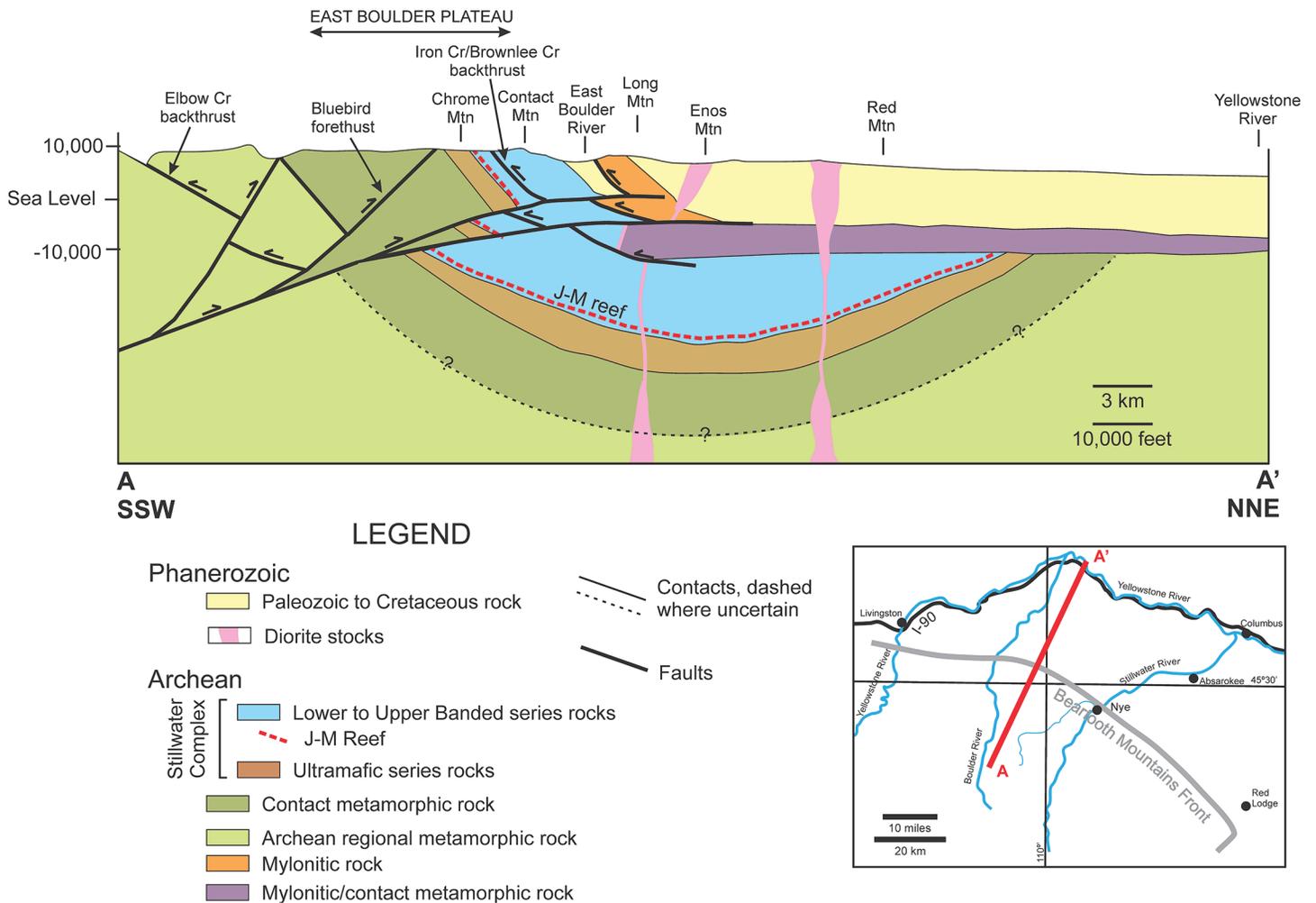


Figure 2. Generalized cross section of the Beartooth Mountains Front Triangle Zone showing the hypothetical Stillwater Complex at depth. Redrawn after Geraghty (2013).

Owing to both the earlier erosional event and especially the Laramide faulting, the original lateral extent of the Stillwater Complex is uncertain. Geophysical evidence suggests that the complex extends at depth for a distance of at least 25 km to the northeast (Kleinkopf, 1985). A buried northern extension is consistent with the presence of altered xenoliths of Stillwater affinities found in the Cretaceous volcanic rocks that erupted through the Phanerozoic sedimentary cover north of the complex (Brozdowski, 1985; DuBray and Harlan, 1998; Lopez, 2000). The Bushveld Complex of South Africa, which has a number of similarities with the Stillwater Complex in terms of rock types and thickness, has an aerial extent of >60,000 km²; this suggests the Stillwater Complex could have initially extended over a similar area.

STRATIGRAPHY OF THE STILLWATER COMPLEX

The Stillwater Complex proper has been subdivided into five major series: the Basal series, the

Ultramafic series, and the Lower, Middle, and Upper Banded series (McCallum and others, 1980; Todd and others, 1982); the last three are informally referred to as the Banded series. In addition, the Sill/Dike unit, a collection of contemporaneous igneous rocks hosted in the immediately underlying metasedimentary rocks, is commonly included as part of the Stillwater stratigraphy (Zientek, 1983). Various subdivisions of Stillwater Complex stratigraphy have been proposed by different authors based on changes in the modal mineralogy, some of which are shown in figure 3. This paper follows the terminology of McCallum and others (1980) as modified by Boudreau (2016). The economic platinumiferous J-M Reef lies ~400 m above the base of the Lower Banded series.

Sill/Dike Unit

Beneath the Stillwater Complex proper is a complex assemblage of broadly contemporaneous gabbro-norites and norites occurring as sills, dikes, and podiform to pipelike igneous bodies that are associated

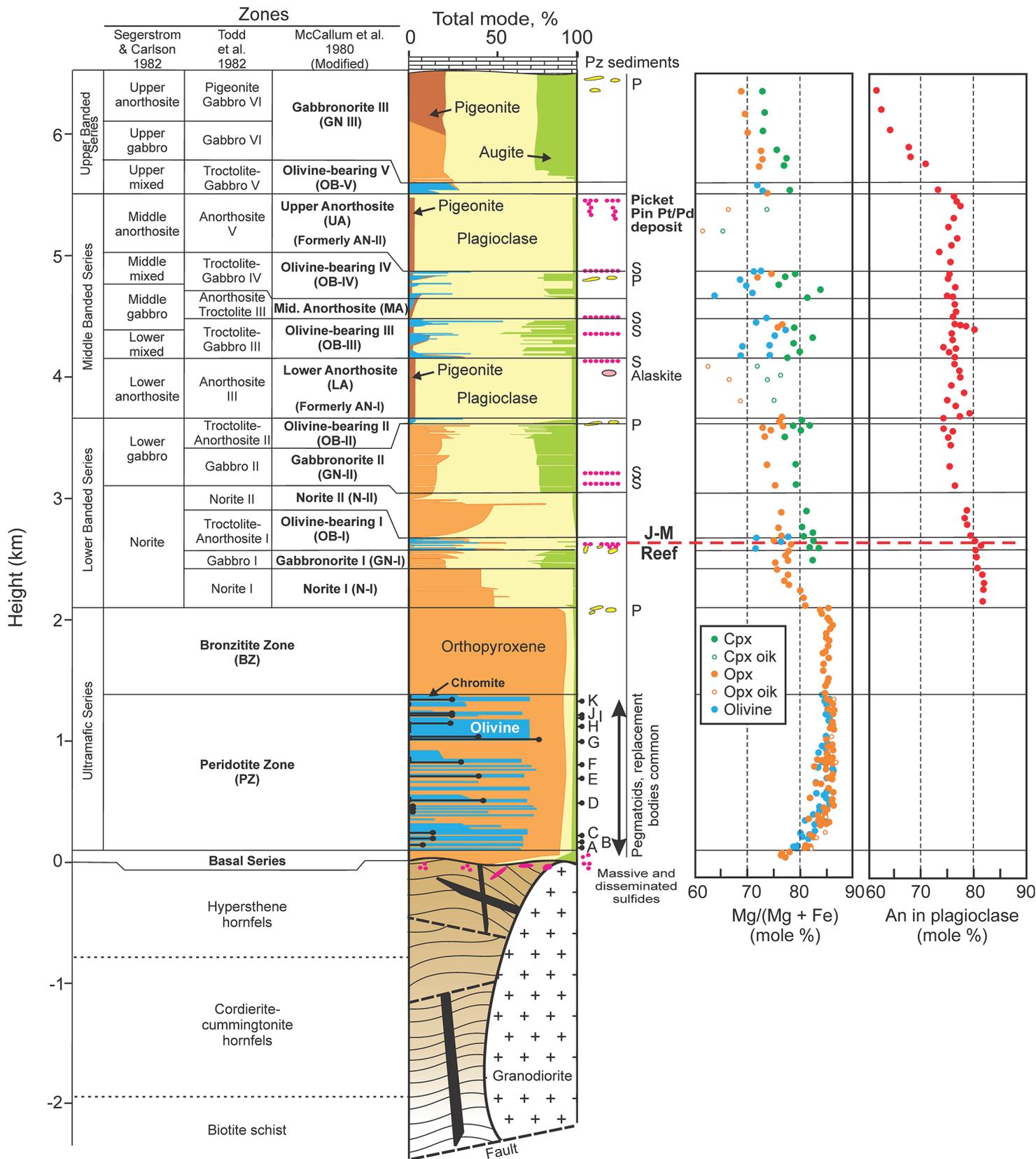


Figure 3. Major mineral modes in the Stillwater Complex and mineral compositional trends as a function of stratigraphic height. Modal olivine in poikilitic and granular harzburgite from the Ultramafic series is in part schematic, based on average modes. Also shown are the locations of the A through K chromitites (labeled “A”–“K”), zones where pegmatoids and sulfides are common (labeled “P” and “S,” respectively), the alaskite body described by Czamanske and others (1991), and the locations of the J-M Reef and the Picket Pin PGE zone (labeled). The lower, Middle, and Upper Banded series are informally known as the Banded series. Also shown are three of the many subdivisions of the Banded series proposed over the years; that used in this report is that of McCallum and others (1985) as modified by Boudreau (2016). After McCallum and others (1985), Raedeke and McCallum (1985), and Raedeke and others (1985) with permission.

with small (typically a few meters in longest dimension) podiform bodies of massive sulfide, hosted in the metasedimentary rocks. The frequency of igneous rocks and sulfides increases towards the complex. In addition, the lower parts of the Stillwater Complex can contain large xenoliths of country rock. Because of this it can be difficult to determine when one is in the metamorphic aureole or in the Basal series in drill core, and some workers include the sill/dike unit with the overlying Basal series. Some of the sill/dike rocks have been proposed to represent the parental Stillwater magma based on geology, age, and composition (e.g., Premo and others, 1990; Helz, 1985, 1995).

Basal Series (60–250 m thick)

The Basal series defines the lowermost rocks of the Stillwater Complex proper. The lower contact of the Basal series is defined at the base of the first laterally continuous norite or orthopyroxenite, although it may not be locally present owing to later faulting. The norite (with subordinate anorthosite, gabbro, and peridotite) grades upward by an unsystematic decrease in plagioclase mode to an orthopyroxenite (Page, 1979; Zientek, 1983).

Ultramafic Series (1,500–1700 m thick)

The base of the Ultramafic series is placed at the first significant appearance of olivine, and the top at the base of the norite (plagioclase + orthopyroxene) that defines the base of the Lower Banded series. It is divided into a lower Peridotite zone and an overlying Bronzite zone. The Peridotite zone is itself composed of a cyclic repetition of nominally peridotite/poikilitic harzburgite → granular harzburgite → orthopyroxenite that repeats 21 times at Mountain View (Raedeke and McCallum, 1984; Cooper, 1997). Chromitite layers occur near the base of many cyclic units and are labeled A–K upward from the bottom (Jones and others, 1960; Jackson, 1961, 1968).

The Bronzite zone at the top of the Ultramafic series is a generally uniform orthopyroxenite with interstitial plagioclase and augite, minor chromite and quartz, and rare phlogopite, apatite, and sulfides. It does have some modal layering near the bottom and top, and marked orthopyroxene size graduations (from the typical 2–3 mm to >1 cm in size) can occur locally throughout (Raedeke and McCallum, 1984).

Lower, Middle and Upper Banded Series (>4,500 m thick)

These three units constitute all rocks above the UL-

tramafic series in which plagioclase is a major modal mineral. The lower and upper contacts of the Middle Banded series are defined by the presence of two thick (to 600 m) anorthosite units. Based on the major rock type or the appearance of olivine as noted in figure 2, these three units are further subdivided into various zones and subzones. Olivine-bearing zone I (OB-I) of the Lower Banded series contains the J-M Reef, an economic concentration of platinum-group element (PGE) mineralization associated with sulfide-enriched rocks that is currently being mined by the Sibanye–Stillwater.

MINING HISTORY

The various mineral deposits of the Stillwater Complex have long attracted interest from the mining community. This history, largely for the J-M Reef, is summarized in table 1, and details of the earlier exploration and mining history can be found in Page and others (1985b). Except for the current operations along the J-M Reef, the only major mining was in the chromitites of the Ultramafic series done under government contract in the 1940s (in the Benbow area) and again in the 1950s to early 1960s (in the Mt. View area).

BASAL MASSIVE NI-CU SULFIDES

Subeconomic amounts of Fe-Ni-Cu sulfides, present as disseminated to meter-sized bodies of massive sulfide (fig. 4), are common in the sill/dike unit at the base of the Stillwater Complex and into the lowermost units of the Stillwater Complex. They are particularly well exposed along Verdigris Creek in the Mountain View area (e.g., Page and others, 1985a; fig. 5). The associated igneous rocks are typically no more than several meters in maximum dimension and can have finer-grained (chilled) margins. Norites and gabbronorites with a diabasic texture and typically poor in sulfides are most commonly associated with the hornfels, whereas a mafic norite in which sulfides make up from 2 to 40% of the rock occurs proximal to the Stillwater Complex (Page, 1979; Zientek, 1983). Sulfide is also concentrated in the lowermost portion of the Basal series but decreases markedly once one moves into the upper part of the Basal Bronzite and the lower Peridotite zone. The bulk rock Cu + Ni content of the sulfide-bearing rocks averages 0.5 wt.% with a Ni/Cu ratio of ~1.0. Zientek and others (1986) note that average bulk rock Pt + Pd + Rh = 20 ppb for all mineralized rock types that include the Basal series rocks.

Table 1. Mining history of the Stillwater Complex.

1883	Nickel-Copper deposits discovered in what will come to be known as the Stillwater Complex. May have been as early as 1863 by Skookum Joe Anderson.
1890	Chromite deposits discovered.
1930	Stillwater Complex named by Princeton University investigators (principally Ed Sampson).
1936	Platinum and palladium-bearing minerals discovered in rocks from the Stillwater Complex. Professor Arthur Buddington of Princeton had speculated on the Merensky Reef analog being present in the Stillwater Complex.
1941–	
1943	Chromite mined in the Benbow area under government contract.
1953–	
1961	Chromite mined in the Mt. View area under government contract.
1967	Initiation of Johns–Manville exploration program on Stillwater Complex. Keith Conn was looking for the Merensky Reef equivalent platinum group metal zone. Reconnaissance by Stan Ellingwood and Bob Mann. Limited sampling by YBRA students. One sample ran about 1 oz per ton.
1973	The platinum/palladium Zone of Interest (ZOI horizon) was discovered through trenching and drilling of the Camp Zone, southwest of Brass Monkey Exploration Camp. Stan Todd's work on the detailed stratigraphy of the ZOI horizon was very instrumental in the discovery.
1974–	
1976	Manville conducts test mining on the ZOI at West Fork adit.
1979	Chevron Minerals Company joins Manville in a joint venture exploration program on the ZOI. This partnership was named Stillwater PGM Resources (SPGMR).
1979–	
1981	Anaconda Minerals Company conducts test mining on the Howland Reef (ZOI) at the Minneapolis adit.
1980–	
1981	The joint venture conducts test mining on the ZOI at Frog Pond adit.
1982	The Zone of Interest is officially named the J-M Reef, after Johns–Manville Corporation.
1983	Anaconda Minerals Company joins Manville and Chevron in a tri-venture on the J-M Reef. This tri-venture is named Stillwater Mining Company.
1983–	
1984	The tri-venture, SPGMR, conducts test mining of the J-M Reef at the Minneapolis adit.
1985	SPGMR initiates a study on the feasibility of a mine start-up. Mining of the J-M Reef begins in October 1985 and has continued unabated through the current time (2018). LAC minerals buys out Anaconda's portion of the tri-venture and joins Manville and Chevron.
1986–	
1987	The go-ahead is given by the partners to start up Stillwater mine. Construction of a concentrator begins. The concentrator is commissioned in March 1987.
1989	Chevron buys out Lac Mineral's portion of the tri-venture.
1990	Smelter is constructed in Columbus.
1994	Stillwater Mining Company IPO; Chevron sells out; Manville retains a Royalty. Number 1 Shaft is started.
1996	Number 1 Shaft is completed; Base Metals Refinery is constructed in Columbus.
1999–	
2002	Development of East Boulder mine. 2002, East Boulder mine begins production.
2003	Norilsk Nickel, a Russian Nickel mining company and the world's largest nickel miner, buys a majority (56%) interest in Stillwater Mining Company.
2010	Norilsk Nickel sells its majority interest out on the open market.
2017	Sibanye Gold, a South African gold and PGM miner, buys Stillwater Mining Company outright. Names the merged companies Sibanye–Stillwater.
2018	Sibanye–Stillwater announces its intent to buy Lonmin, the world's 3rd largest platinum producer. Deal is expected to close by mid-year.

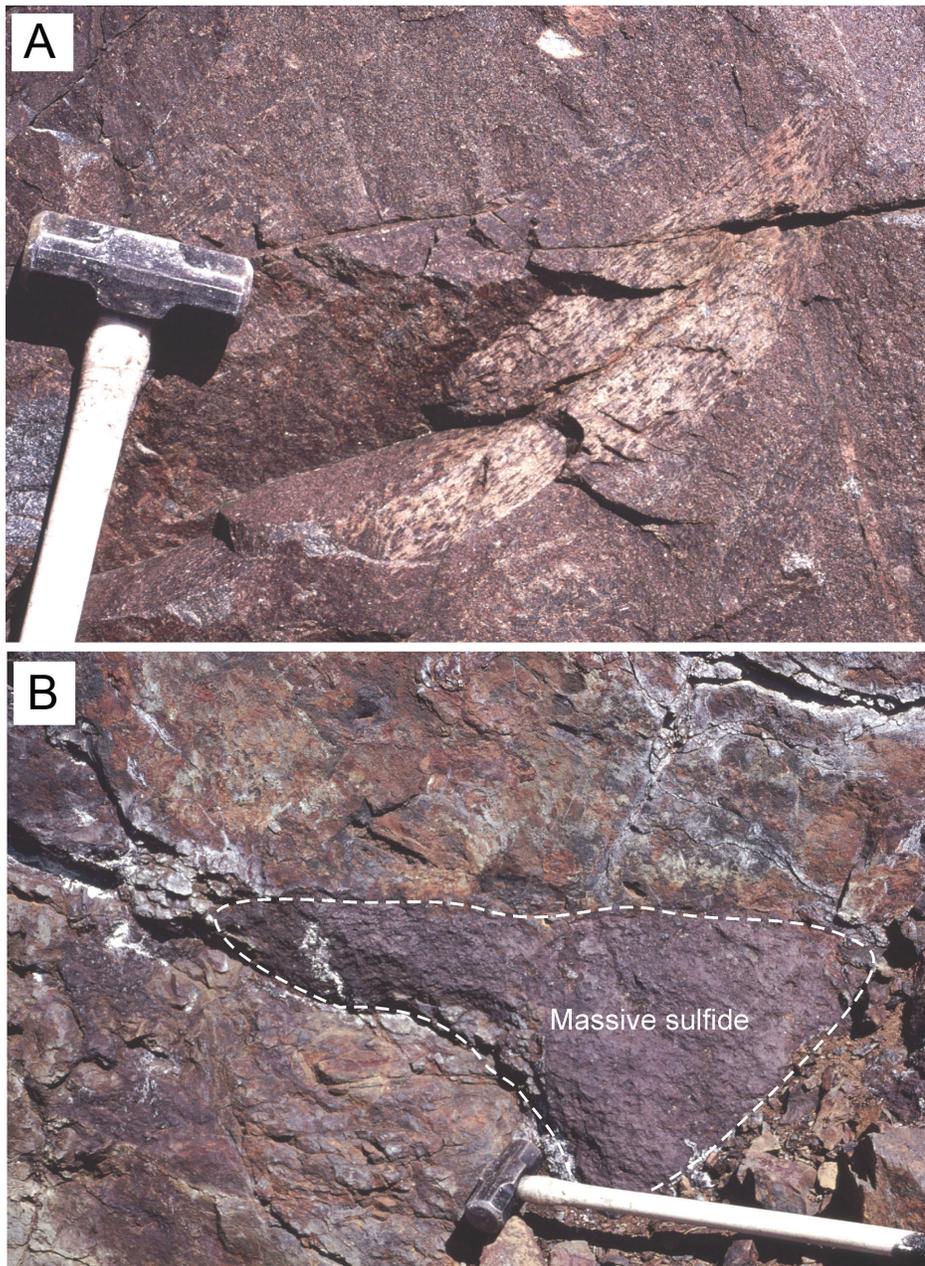


Figure 4. (A) Fragmental diamict texture in hornfels beneath the Stillwater Complex interpreted as relict fragmentary sedimentary texture (breccia). (B) A typical example of a small body of massive sulfide. Both photos from the Verdigris Creek area at Mountain View.

Classic interpretations suggested that the initial intrusion of the Stillwater magma was sulfide-saturated and that sulfide ponded in depressions at the base of the complex. However, the podiform nature of the mineralization and experimental evidence that sulfide saturation has a negative pressure dependence such that magmas rising from the mantle should be strongly S-undersaturated have required other interpretations. Based on the low PGE tenor and similar S isotope composition, Zientek and Ripley (1990) suggested that S was derived from the iron formation and assimilated into a limited volume of magma at the base of the

complex. It is noted, however, that the iron formation occurs nearest the complex only along the middle sections and is otherwise rare or not associated with the better-mineralized footwall regions, nor is it evident that the sill/dike compositions are consistent with assimilated iron formation (Helz, 1985). Vaniman and others (1980) noted that the contact metamorphosed iron formation has sulfides (pyrite, pyrrhotite, chalcopyrite) associated with the olivine-bearing metamorphic assemblage and associated with fractures. They suggested that Cu and S in the iron formation were introduced by fluids moving downward from the Stillwater Complex. In contrast, Frost (1982) suggested that the metamorphic assemblage was self-buffering and that it was unlikely that the assemblage was controlled by an external fluid. A recent study by Ripley and others (2017) suggested that up to 50% of the sulfur in some samples of the Basal series was of crustal origin, based in part on mass independent sulfur isotope variations, $\Delta 33S$. Benson and others (2020) suggested that volatile fluid generated by dehydration of the country rock would be analogous to dehydration of descending lithospheric slabs at subduction zones and that these fluids would tend to move upward and into the base of the intrusion, which may have moved country rock ore components as well.

CHROMITE DEPOSITS OF THE ULTRAMAFIC SERIES

The chromite deposits of the Stillwater Complex have long been the subject of extensive study (e.g., Peoples and Howland, 1940; Jackson, 1961, 1963; Page, 1971; Foose and Nicholson, 1990; Talkington and Lipin, 1986; Loferski and others, 1990; Campbell and Murck, 1993; Waters and Boudreau, 1996; Cooper, 1997; Horan and others, 2001; Spandler and others, 2005; Pagé and others, 2011; Lenaz and oth-

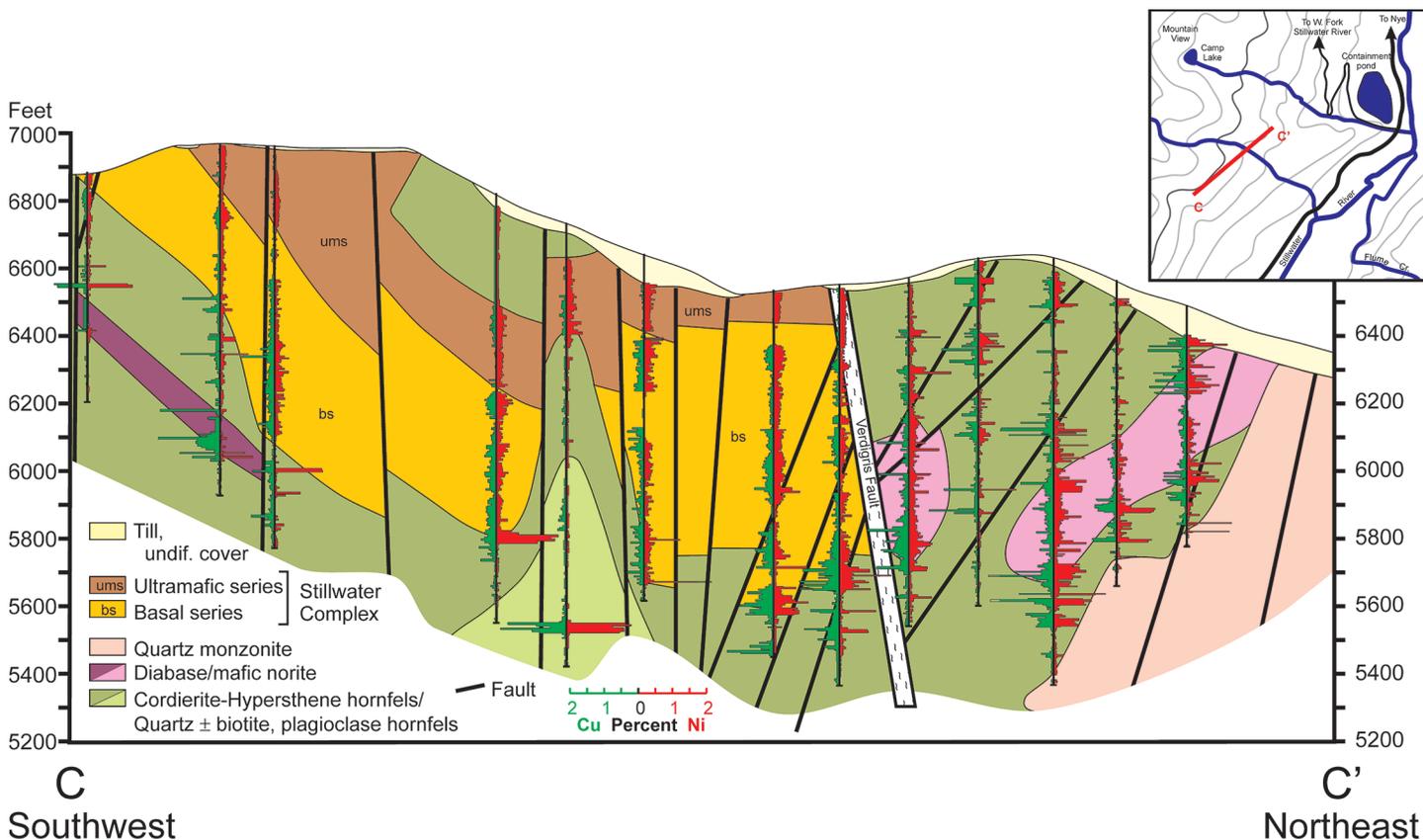


Figure 5. Section showing the geology and copper–nickel variation in drillholes through the lower units of the Stillwater Complex and underlying metasedimentary rocks in the Verdigris Creek area, Stillwater River valley. Note that the Cu–Ni values decrease as one moves into the lower parts of the Ultramafic series. Section redrawn after Zientek and others (1989).

ers, 2012; Barnes and others, 2016). In brief, layers of massive and disseminated chromite occur in the lower parts of many but not all of the cyclic units the Peridotite zone. The main chromitite seams are denoted as A (lowermost) through K (uppermost), with the G and H having the thickest development of chromitite. This labeling can be complicated by the fact that some zones can have multiple chromitites, such that the G and H at Mountain View actually have a G, G', H, and H' chromitite (fig. 6). Furthermore, even these sub-units can be made up of numerous smaller interlayered chromitite/dunite–harzburgite rocks. Modal layering in chromite-bearing rocks ranges from poor to well layered, with the latter occurring on scales from ~meter thick down to laterally extensive chromitite layers only a grain or two thick (fig. 7). Minor chromitite seams also occur in Olivine-bearing zone I of the Banded series (Waters and Boudreau, 1996; Boudreau, 2016). The chromitites and host rocks can also have subeconomic concentrations of PGE-bearing sulfides. Relatively high PGE concentrations are associated with the A and B chromite seams near the base of the Ultramafic series (e.g., Zientek and others, 2002).

Chromitite Formation Models

As noted above, Jackson (1961) was perhaps the first to note that the crystal grain size distributions (CSD) of olivine and chromite in rocks where both occur are similar to CSD's seen in clastic sedimentary rocks, but that otherwise these two minerals are not “hydraulically equivalent.” That is, although the chromite is denser, the greater size of the olivine implies a faster settling velocity. Waters and Boudreau (1996) suggested that this discrepancy could be accounted for by the process of crystal aging, during which larger grains grow at the expense of smaller grains owing to surface energy considerations. Boudreau (1994, 1995) and Boudreau and McBirney (1997) suggested that surface energy is a function of the surrounding minerals such that chromite has a lower free energy when surrounded by other chromite crystals rather than silicate minerals. This can lead to phase segregation over long cooling times to produce well-defined fine-scale modal layering, broadly similar to processes that form metamorphic banding.

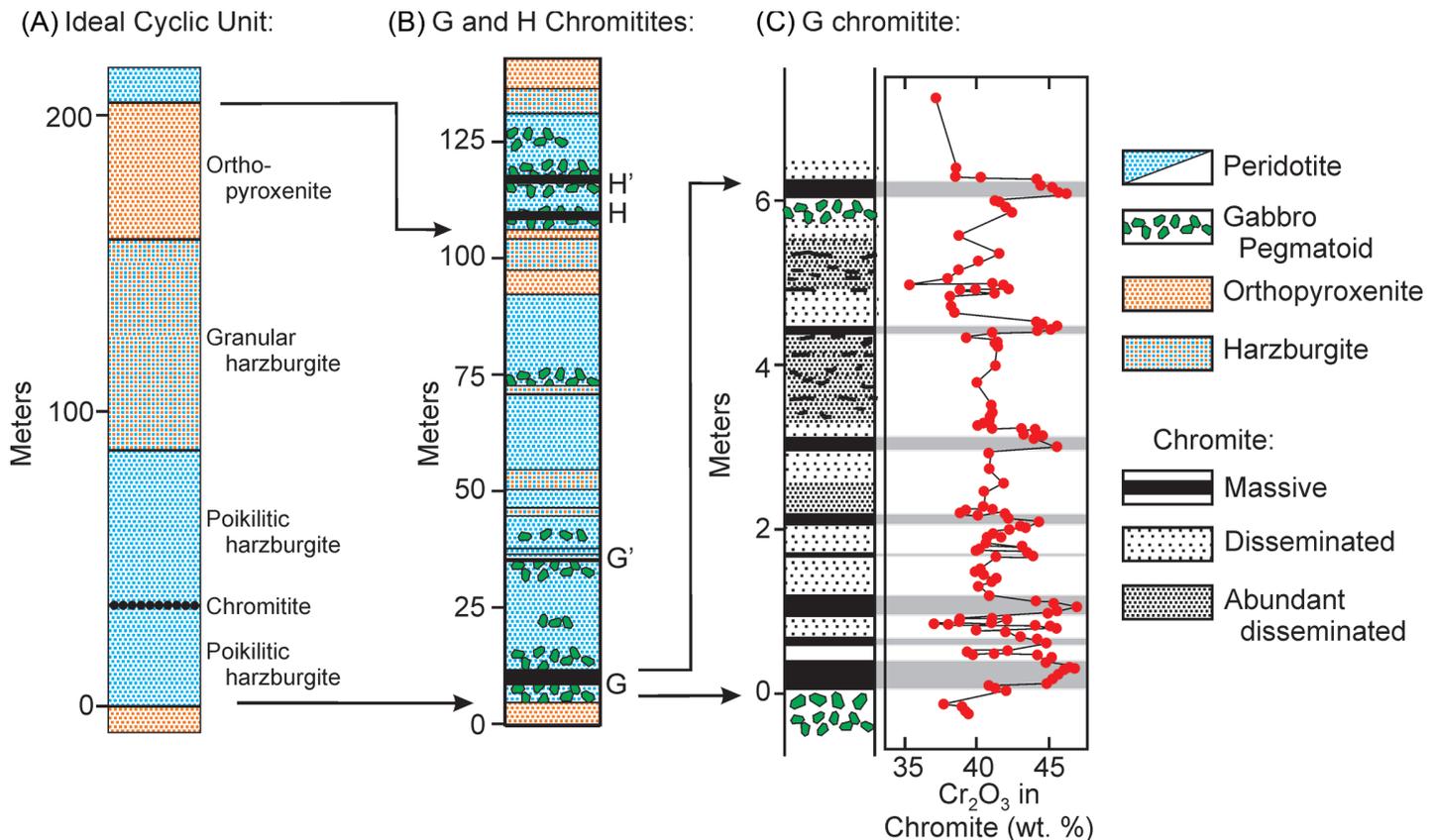


Figure 6. A detail of the complexity of the G chromite unit of the Peridotite zone. (A) "Ideal" Peridotite zone cyclic unit (after Jackson, 1961). In reality, the contact between the poikilitic harzburgite and the overlying granular harzburgite is irregular (fig. 7A) and the chromitite can be variably developed, as shown. (B) The section containing the G and H chromitite units as exposed at Mountain View. (C) Detail of the G chromitite. Redrawn after Campbell and Murck (1993), with permission.

It has otherwise been a longstanding petrologic problem of how to produce meter-size layers of essentially monomineralic chromitites. If precipitated directly from a magma, these thick chromitites require an extensive period of crystallization of chromite alone, a mineral which otherwise would only crystallize in cotectic proportions of about 1%. In part the problem is that olivine and chromite can be produced by a number of mechanisms, ranging from those that are strictly magmatic to those that envision a role by volatile fluids. These mechanisms also apply to the formation of the J-M Reef, where again the question arises as to why olivine and chromite reappear in the stratigraphic section. These models can be summarized as follows:

(1) Direct precipitation from a mantle melt. Olivine and chromite can precipitate directly from a primitive mantle liquid saturated in one or both of these two minerals. This is commonly observed in boninites, for example (e.g., Crawford and others, 1989). A variation on this is the suggestion that chromitites formed from the injection of chromite-rich slurries (e.g., Voordouw and others, 2009), but this simply moves the problem to somewhere off-stage.

(2) Magma mixing. The most commonly accepted model involves the injection of primitive magma and mixing with fractionated magma resident in the magma chamber that drives the hybrid magma into the chromite-only field (fig. 8A; e.g., Irvine, 1977; Campbell and Murck, 1993). Similarly, contamination with a silica-rich composition (e.g., melted roof rock) can also move the liquid into the chromite-only field (e.g., Irvine, 1975; Spandler and others, 2005).

(3) Pressure changes. Pressure changes on the magma chamber cause shifts in the chromite cotectic boundary such that an increase in the pressure on the magma moves the cotectic to crystallize the dense chromite alone (e.g., Loferski and others, 1990; Lipin, 1993). A problem is whether or not a large sill in which the roof is essentially floating on the sill can support the development of significant overpressures. Also, increasing pressure tends to shrink the olivine phase field. Although this allows chromite alone to crystallize, it would not be consistent with the change from cotectic crystallization required to make the underlying granular harzburgite to peritectic reaction relationships between olivine and orthopyroxene to produce the poikilitic harzburgite that hosts the chromitites.

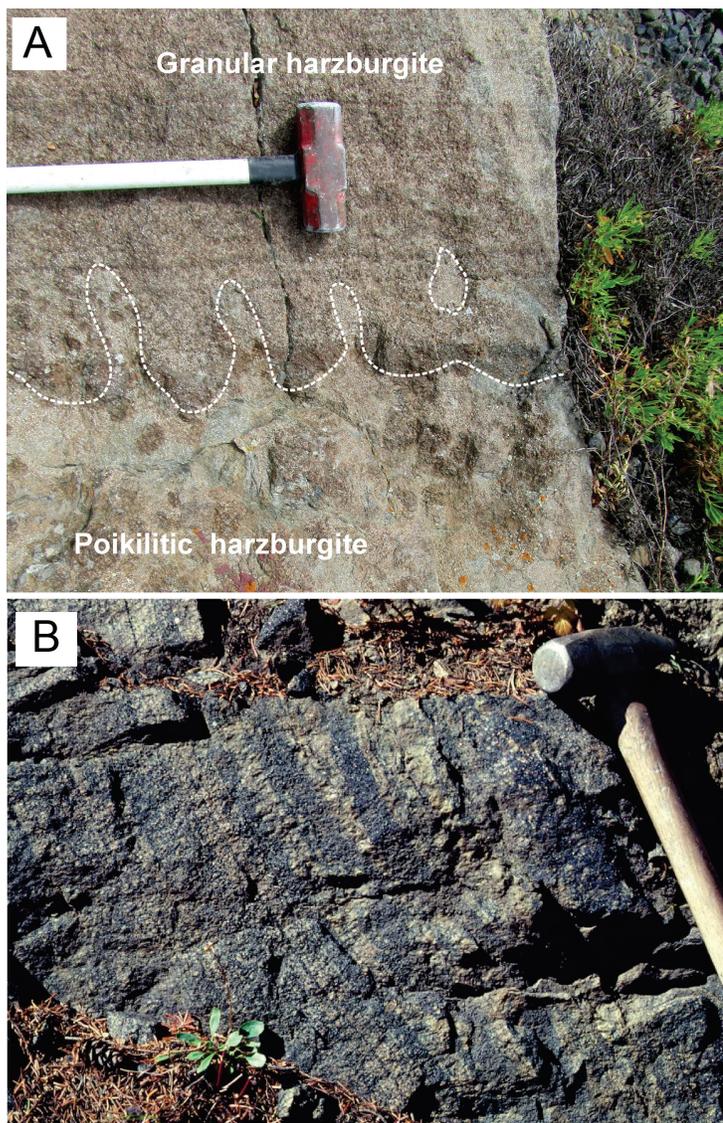
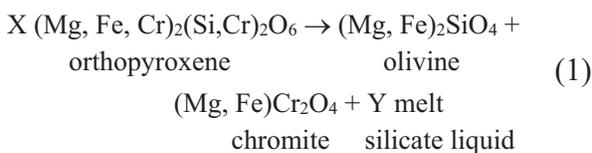


Figure 7. Rock of the Peridotite zone of the Ultramafic series. (A) Photo of poikilitic harzburgite developed as “fingers” intruding the overlying granular harzburgite indicative of a replacement origin for the poikilitic harzburgite. (B) Example of interlayered chromitite (dark layers) and olivine-rich rock (dunite) in the G chromitite unit.

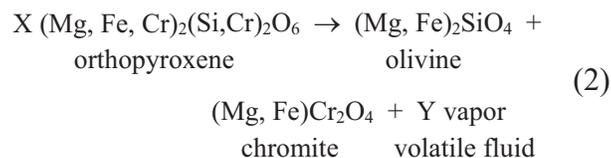
(4) Addition of volatiles. An increase in H_2O and/or other volatiles can also shift the chromite cotectic such as to drive the magma to crystallize chromite alone (e.g., Ford and others, 1972; Mathez and Kinzler, 2017). This can also shift the olivine–pyroxene boundary such that a cotectic boundary between these two minerals becomes a peritectic reaction (fig. 8B).

(5) Incongruent melting. Both minerals can be produced by incongruent melting of Cr-bearing pyroxene via a model reaction (e.g., Kushiro, 1975):



This melting can be either by direct heating (by injection of a hotter magma, for example) or by hydration melting, similar to the production of arc magmas by hydration of mantle source rocks by fluids coming off a subducting lithospheric slab. Hydration melting of a chromite \pm orthopyroxene \pm olivine mush preferentially leaves behind chromite as a residual (restitute) mineral (e.g., Brearly and Scarfe, 1986; Armitage, 1992).

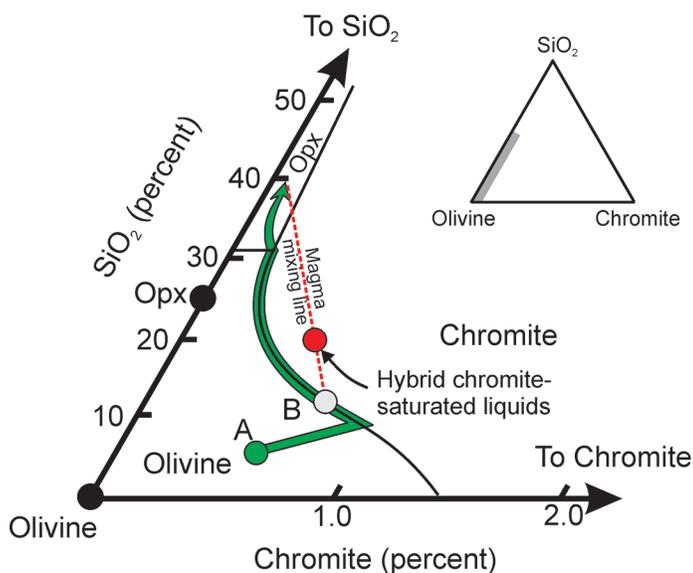
(6) Metasomatic silica loss. Olivine and chromite can be formed by loss of silica to a silica-undersaturated volatile fluid in a model reaction similar to reaction (1) above:



The hydration models 5 and 6 have not been extensively called upon to explain the Stillwater chromitites outside of the smaller chromite seams that occur, for example, at the margin of the pegmatoids noted above and the minor chromite that occurs in Olivine-bearing zone I (e.g., Boudreau, 1999). However, several lines of evidence are consistent with such a process operating in the main chromitite seams (Boudreau, 20160). These include: (1) the increase in Cl in apatite and the halite inclusions in olivine are consistent with a Cl-rich vapor moving upward through the crystal pile. (2) Polyphase inclusions in chromite, interpreted to have crystallized from melt inclusions, are rich in volatiles (Spandler and others, 2005), implying that the chromite grew from hydrated magmas. (3) Injection of S-bearing vapor can lead to a local saturation of the magma in sulfide, resulting in elevated PGE concentrations. (4) The increase in H_2O can change the olivine–orthopyroxene boundary from an equilibrium cotectic boundary to a peritectic relationship. This would suggest that the poikilitic harzburgite, the typical host to the chromitite seams, crystallized from a magma with higher H_2O concentration to produce a reaction relationship between olivine and the orthopyroxene oikocrysts. In contrast, the granular harzburgite crystallized from a drier liquid in which olivine and orthopyroxene crystallized happily together along a cotectic.

The models presented are not necessarily independent of each other. For example, one could imagine that a conventional magma-mixing event pushes the

(A) Magma Mixing/Assimilation:



(B) Magma Hydration:

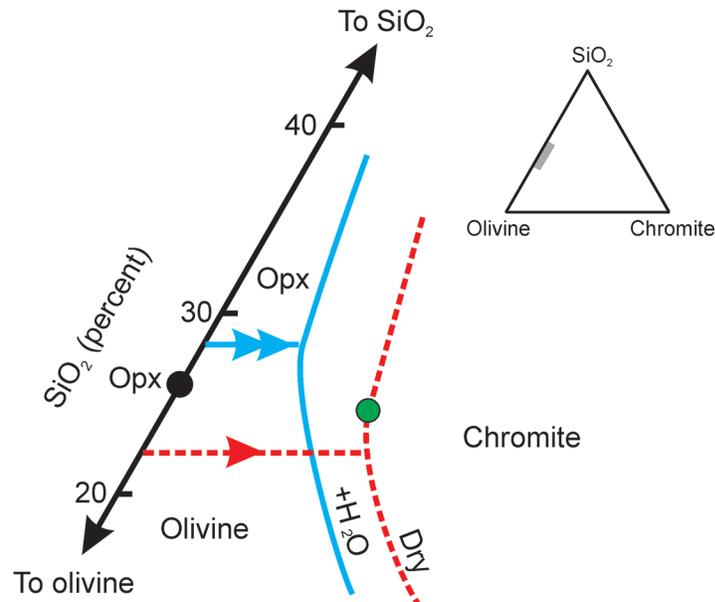


Figure 8. Two endmember models for formation of chromitites. (A) Detail of the chromite–olivine– SiO_2 systems showing a possible normal fractionation path for a liquid starting in the olivine phase field at point “A.” After Irvine (1977), with permission. Also shown are mixing paths that pull an olivine + chromite-saturated liquid (point “B”) into the chromite field. The red dashed “Magma mixing line” is the range of hybrid compositions produced by the mixing of liquid “B” with more evolved, opx-saturated liquid at the end of the green line. (B) Effect of H_2O addition to magma to expand the olivine and chromite phase fields: liquid on opx-chromite cotectic in a relatively “dry” magma (green circle on the red dashed line) becomes saturated in chromite alone with the shift of phase boundaries after the addition of H_2O (blue solid line). Note that expansion of the olivine field changes olivine–opx from a cotectic in the dry system (red lines) to a peritectic in the wet (blue lines). After Boudreau (2016), with permission.

hybrid magma into the chromite-only phase field. After accumulation on the floor, the assemblage is then further enriched in chromite by hydration melting and loss of the partial melt as vapor from the underlying mush migrates upward into the hotter, vapor-undersaturated assemblage.

THE JOHNS–MANVILLE (J-M) REEF, A PGE DEPOSIT

The J-M Reef Pd-Pt deposit is hosted by Olivine-bearing zone I (OB-I), a zone defined by the first reappearance of olivine located about 400 m above the base of the Lower Banded series. (The term “reef” is adopted from South African mining terminology whereby reef implies a mineralized vein and has nothing to do with the sedimentary feature of the same name.) The J-M Reef has among the highest average Pt + Pd grades of all known PGE deposits (~18 ppm with an average Pd/Pt mass ratio of 3; Zientek and others, 2002) with proven and probable reserves of about 20,000,000 oz. It is currently being exploited from two mine sites, the Stillwater mine in the east and the East Boulder mine in the west. It is the only ore zone of the Stillwater Complex that has been successfully mined without government support. In 2016, both mines produced a total 420,200 oz Pd and 125,100 oz Pt for an

average price of \$743/oz. or a total of \$405.1 million with a production cost of \$279.3 million (Stillwater Mining Company 2016 Annual Report). Owing to its economic importance, the J-M Reef and its host rocks have been the subject of a number of studies (e.g., Todd and others, 1982; Bow and others, 1982; Irvine and others, 1983; Barnes and Naldrett, 1985, 1986; Lambert and Simmons, 1988; Page and Moring, 1990; Braun and others, 1994; Boudreau, 1988, 1999; Godel and Barnes, 2008; and many more).

Stratigraphy

The J-M Reef occurs in the Olivine-bearing zone I (OB-I) of the Lower Banded series (fig. 3) and some features of the zone are illustrated in figure 9. It is underlain by Norite zone I (N-I) and Gabbronorite zone I (GN-I) that define the basal units of the Banded series. Norite zone I consists predominantly of a modally and texturally uniform norite with cotectic proportions of plagioclase and orthopyroxene. While Norite zone I is typically massive without significant modal layering, pegmatoids and modal layering are common just above the contact with the Ultramafic series, and a distinctive anorthosite layer occurs in the middle of the unit. The lower part of Gabbronorite zone I continues the cotectic proportions of plagioclase, orthopyroxene,

and clinopyroxene, whereas in the upper part of GN-I modal layering again becomes common, ranging from melano-gabbronorite to anorthosite. Locally, soft-sediment-like deformational features are observed (fig. 9A). Qualitatively, the modal layering appears to be a segregation of pyroxene from plagioclase, with overall proportions approximately cotectic. Discordant pegmatoids also again become common in GN-I (figs. 9B, 9C).

Olivine-bearing zone I is highly variable along strike in terms of thickness and mineralogy and even its stratigraphic location can be variable. Layers need not be laterally continuous, and abrupt lateral pot-

hole-like terminations can occur, most commonly with the olivine-bearing lithologies against gabbronorite but also with anorthosites against gabbronorite. Boudreau (2016) suggested that the section at the Frog Pond area south of Contact Mountain described by McCallum and others (1980) and Todd and others (1982) can be considered the “type section” to which lateral variations can be compared (fig. 10). Here, OB-I is relatively pyroxene-rich in the lower half (the Gabbronorite subzone) and anorthositic in the upper half (the Anorthosite subzone). The rocks in the Gabbronorite subzone are mainly gabbronorite with minor interlayered anorthosite, norite and medium- to coarse-grained

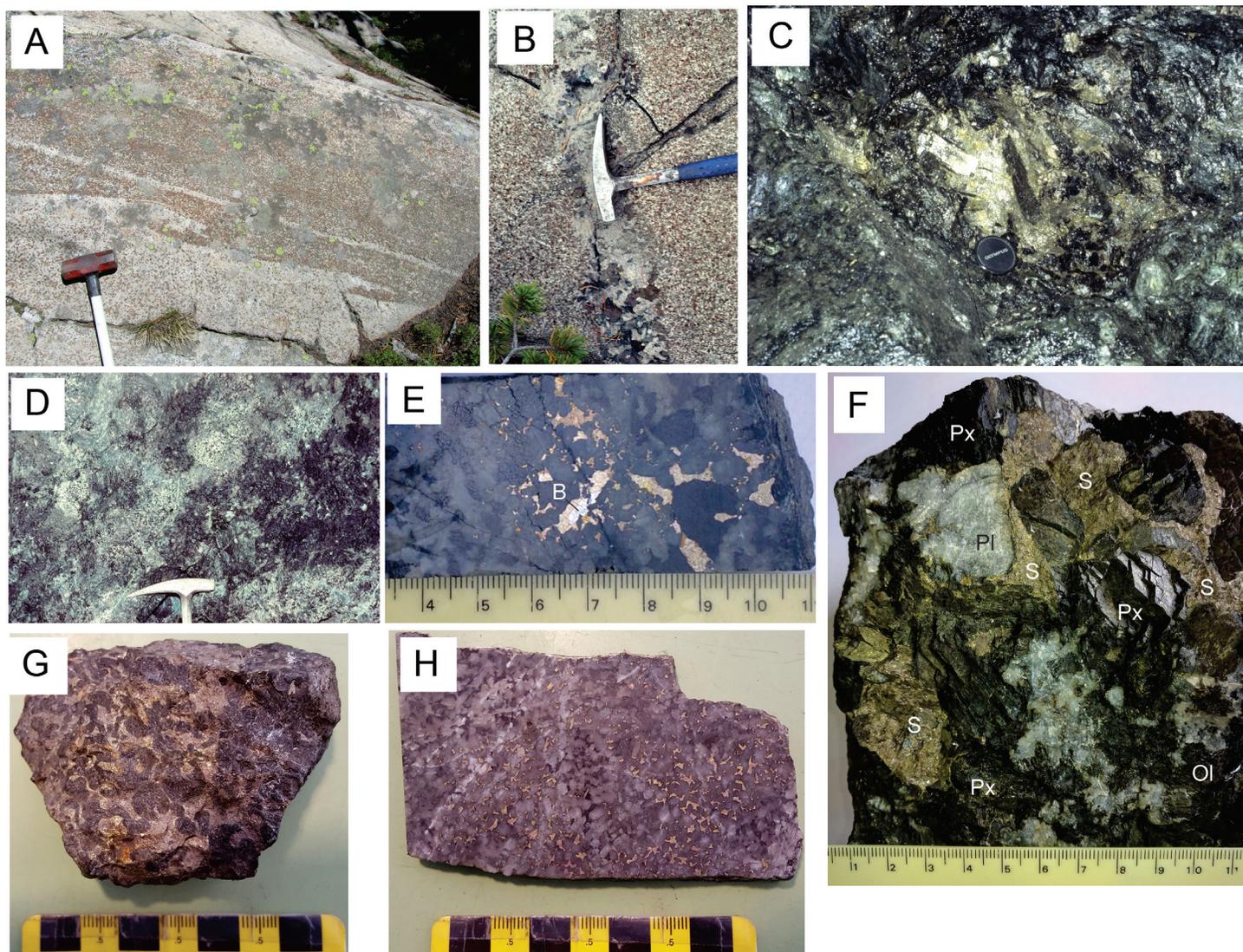
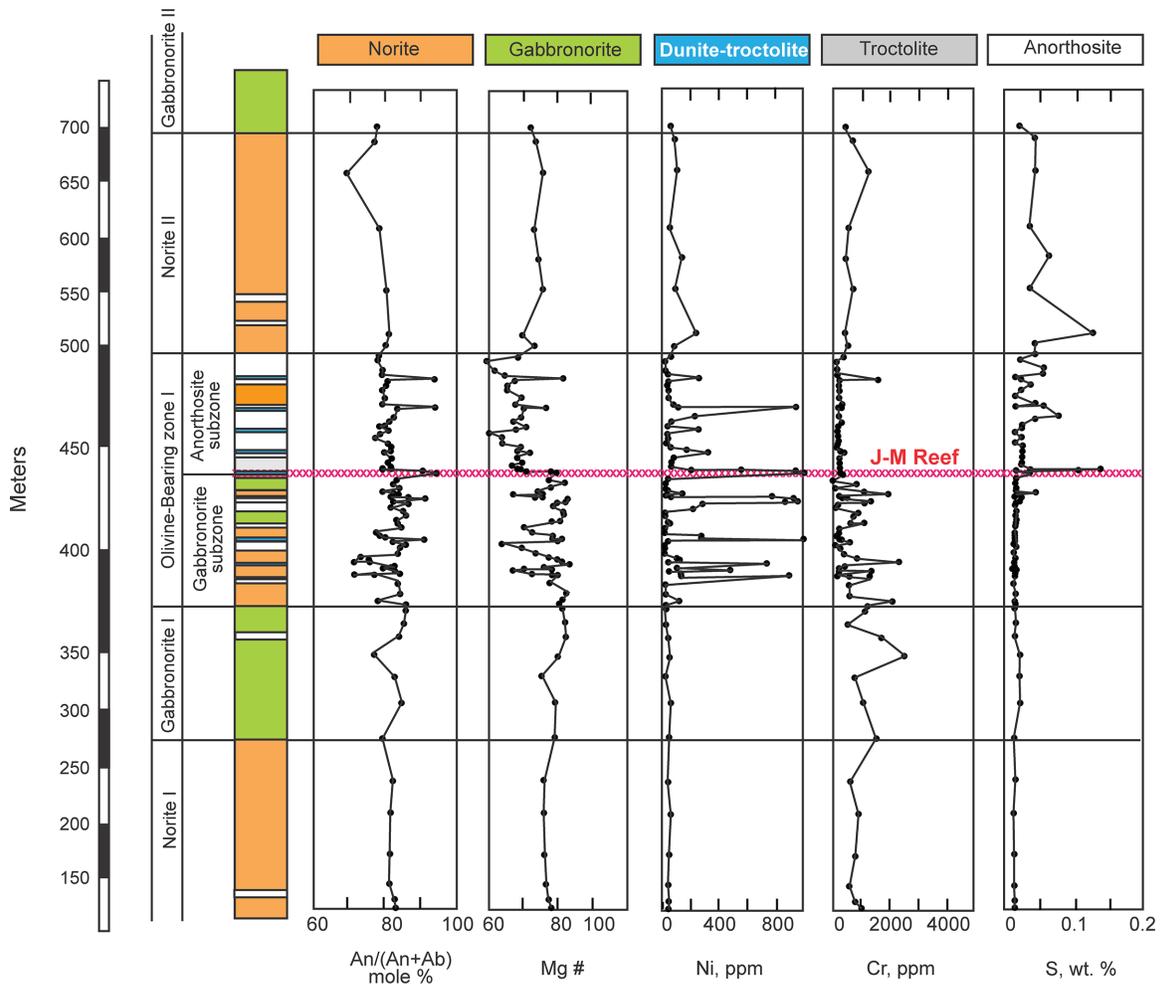


Figure 9. Photographs of some of the features of the J-M Reef from Olivine-bearing zone I (OB-I) and the underlying footwall of Gabbronorite zone I (GN-I). (A) Small soft sediment-like folds in modally layered rocks in GN-I beneath the J-M Reef. (B) Discordant pegmatoid crosscutting gabbronorite in GN-I. (C) Small massive sulfide at the core of a discordant silicate pegmatoid (note lens cap for scale). (D) Texturally and modally mixed olivine-bearing rocks that host J-M Reef sulfide mineralization. (E) Example of sulfide mineralized gabbronorite, composed mainly of pyrrhotite, chalcopyrite, and pentlandite (polished drill core). The silver grain in the center of the core labeled “B” and cut by a small vein is likely braggite, (Pd, Pd, Ni)S. (F) Mineralized very coarse to pegmatoidal olivine gabbronorite. S, sulfide; Px, pyroxene (with prominent cleavage); Pl, plagioclase; Ol, olivine. (G) A well-mineralized troctolite, in which sulfide fills most of the space between the darker, centimeter-sized rounded olivine grains. (H) Typical mineralized anorthosite (polished slab). Scales on images E–H are in centimeters.

(A) Bulk rock and CIPW normative compositional trends



(B) Microprobe mineral compositional trends

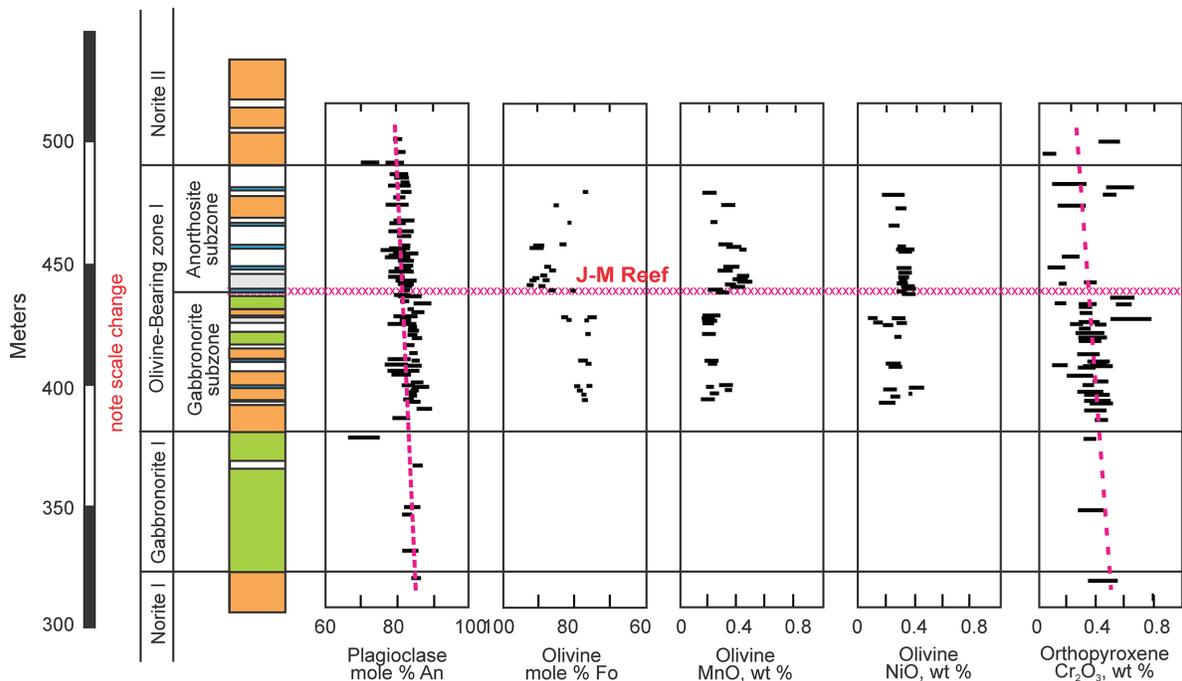


Figure 10. Mineral and bulk rock compositional trends from the upper part of N-I to the base of GN-II illustrating the lack of distinct compositional breaks associated with the J-M Reef. (A) Bulk rock and CIPW normative mineral trends (note scale change through OB-I). (B) Electron microprobe mineral compositional trends. Redrawn after Todd and others (1982), with permission.

troctolite and subordinate dunite, the later typically with pegmatoidal texture. The Anorthosite subzone is composed predominantly of anorthosite, anorthositic troctolite (both with minor oikocrysts of augite and orthopyroxene), and norite. PGE-enriched sulfides of the J-M Reef are most commonly concentrated within the gabbronorite subzone of OB-I, and are particularly associated with a pegmatoidal olivine-bearing unit just below the contact between the Gabbronorite and Anorthosite subzones. PGE-sulfide mineralization can occur in rocks ranging from norite and gabbronorite to troctolite-dunite.

Above OB-I, the rocks return to the mineral sequence observed in N-I and GN-I to form the Norite zone II (N-II) and Gabbronorite zone II (GN-II). However, mineral compositions in N-II and GN-II are more evolved than those in the lower units (McCallum and others, 1980; Todd and others, 1982; Barnes and Naldrett, 1986) and mineral compositional trends from N-I through OB-I and into N-II and GN-II show very little evidence for any kind of distinctly different magma compositions being injected into the Stillwater chamber to form OB-I (fig. 10). For example, plagioclase, which is the only major mineral found in all units, shows a continuous trend to lower An com-

positions with stratigraphic height with little evidence of compositional offsets that would be expected if the reappearance of olivine represented in some sense the injection of a more primitive magma. While olivine shows an offset at the boundary between the Troctolite and Anorthosite subzones of OB-I, the change is to more Fe-rich compositions in leuco-troctolite of the structurally higher Anorthosite subzone and likely represents a compositional shift from variable olivine/trapped liquid abundance. While there is a modest jump to more sulfur-rich rocks above the J-M Reef, it is debatable if this represents the actual point of sulfur-saturation in the magma or if the change represents a loss of sulfur from the rocks below the reef (see below). There is also a sharp change in the Cl content of apatite; the rocks below the Anorthosite subzone of OB-I and especially the olivine-bearing rocks are unusually Cl-rich (fig. 11). Above the reef, the apatite becomes more F-rich, as is typical of other layered intrusions (e.g., Boudreau and McCallum, 1989).

In the upper part of the Lower Banded series, Norite zone II and the lower parts of Gabbronorite zone II are typically massive, commonly laminated, and not modally layered. The rocks again become modally layered in the upper part of GN-II as the thick anorthosites of the Middle Banded series are approached, forming a number of macrorhythmic units (Criscenti, 1984). Olivine-bearing zone II (OB-II) defines the top of the Lower Banded series and contains the distinctive “pillow troctolite” of Hess (1960). The top of the Lower Banded series is defined by the base of the first thick anorthosite unit of the Middle Banded series.

Primary Structures

This “type section” can be compared with the section as it occurs between the Stillwater River and the West Fork of the Stillwater River. Here, OB-I cuts down into the underlying section, locally reaching the lower parts of N-I (Turner and others, 1985) in what is termed the Dow Meadow Depression by Sibanye-Stillwater geologists (fig. 12). Within this apparent unconformable region, the thickness of OB-I itself decreases, such that at the deepest penetration into the underlying rocks

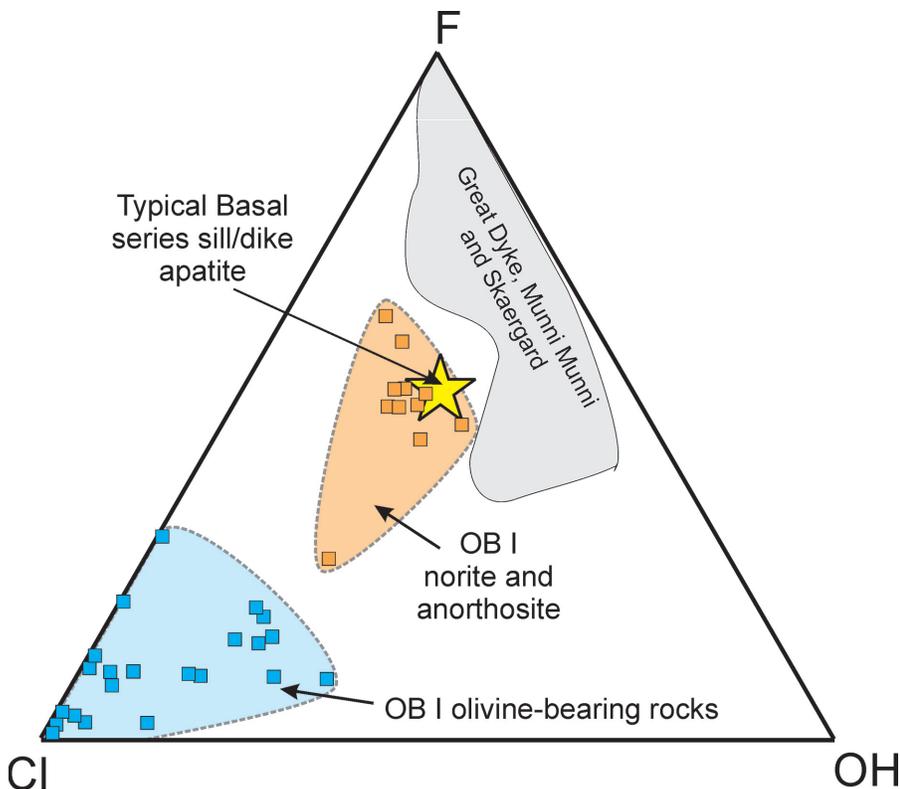


Figure 11. Compositional variation in apatite from OB-I illustrating the unusually Cl-rich nature of apatite from the olivine-bearing rocks as compared with apatite from norite, anorthosite, Basal series sill/dike rocks, as well as apatite from the Great Dyke, Munni Munni, and Skaergaard intrusions. Redrawn after Boudreau and McCallum (1989), with permission.

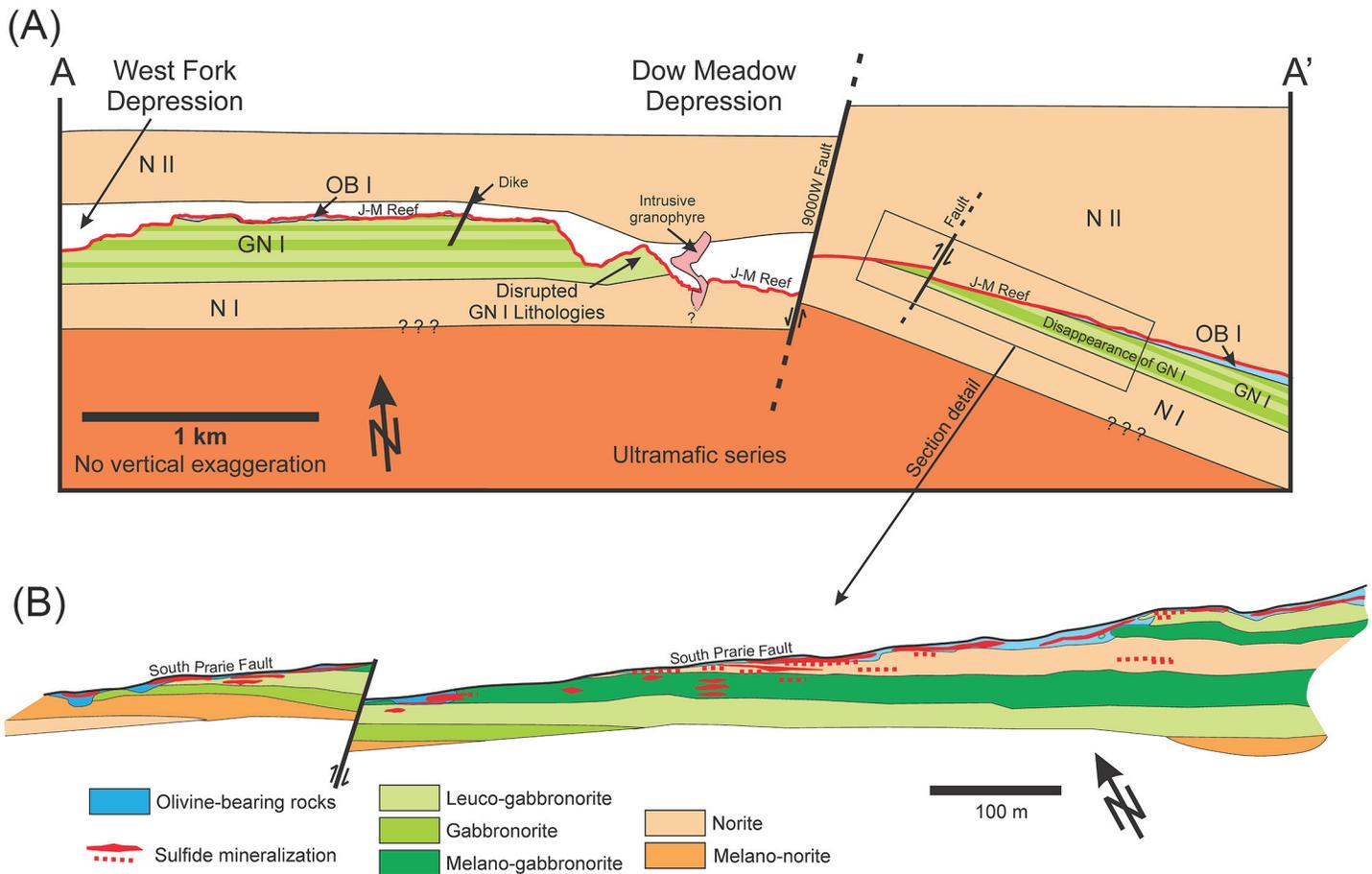


Figure 12. Regional downcutting of the J-M Reef into the underlying footwall rocks at the “Dow Meadow Depression” illustrating stratigraphic variability of the J-M Reef. (A) Regional section constructed along the line of section A–A’ shown. The reef cuts down or replaces original footwall lithologies. Note also the presence of the “West Fork Depression,” partially shown on the left side of the figure. Later faulting on the 9000W fault has resulted in a 15° rotation of the layering just east of the depression. (B) Detail of the reef development into the underlying GN-I and N-I zones showing the presence of olivine-bearing rocks along the unconformity and the irregular distribution of sulfide mineralization that defines the J-M Reef. Part B redrawn after Zientek and others (2002), with permission from the Canadian Institute of Mining, Metallurgy and Petroleum.

there is a very narrow transition from N-I to N-II; i.e., GN-I and OB-I are largely absent. Where present, the olivine-rich layers of the Gabbronorite subzone again can show abrupt lateral changes to melanocratic gabbronorite layers (Turner and others, 1985). Although interpretation is complicated by the layer-parallel faulting on the South Prairie Fault (fig. 12B) in the overlying Anorthosite subzone, this subzone also appears to thin, giving the regional appearance that the Gabbronorite and Anorthosite subzones both form an apparent angular unconformity at their contact. The surface of the unconformity itself is broadly defined by PGE–sulfide mineralization of the J-M Reef, although as shown below, mineralization is considerably less abundant and locally absent in the Dow Meadow Depression. Ongoing underground development is currently defining a second region of downcutting, the West Fork Depression, further to the west near the West Fork of the Stillwater River (fig. 12). The regions of both depressions have otherwise not been exten-

sively studied; it is not known if there is a similar lack of mineral compositional offsets across the unconformity similar to the uniform mineral trends shown in figure 9, for example.

Occurring in the center of the Dow Meadow Depression is a coarse-grained to pegmatoidal felsic intrusion up to several tens of meters in size with a granophyric texture and composed mainly of plagioclase with variable amounts of quartz, biotite, tremolite, and sparse microcline. Its contacts with the Stillwater Complex rock proper are sharp but wavy/irregular, and its stratigraphic extent below the region of mine development and drilling is uncertain. It is locally mineralized, with sporadic total PGM concentrations up to 10 ppm (0.30 oz per ton). Its composition is similar to that reported by Czamanske and others (1991) for a PGE-bearing alaskite/granophyre from the Lower Anorthosite zone of the Middle Banded series at Picket Pin Mountain (Ching, 2017).

The J-M Reef

The PGE enrichments that define the J-M Reef occur primarily both in solid solution within disseminated base-metal sulfides (typically up to 5% volume, and locally in small bodies of massive sulfides in cores of podiform and discordant pegmatoids) and as a variety of discrete trace PGE-rich mineral phases presumably exsolved from the higher temperature sulfide assemblage on cooling (e.g., Barnes and Naldrett, 1985; Godel and Barnes, 2008; Turner and others, 1985; Zientek and others, 2002). PGE tenor is higher in the eastern part of the Stillwater mine than in both the western area of the Stillwater mine and continuing to the Boulder River mine even farther west. Although most typically occurring in olivine-bearing rocks, mineralization can occur in all rock types and textural variations found in OB-I (figs. 9E to 9H).

Although on the scale of the whole complex, the sulfide-bearing rocks that define the J-M Reef are broadly stratabound in that one can trace it across the exposed part of the complex, mineralization is heterogeneously distributed at smaller scales. The J-M Reef is typically described as a zone of mineralization 2 to 5 m thick below the contact of the Gabbronorite–Anorthosite subzones of OB-I, occurring mainly in a coarse-grained to pegmatoidal olivine-rich rock (fig. 9A). However, mineralization is laterally variable to discontinuous, such that effective mining requires drilling at ~15 m centers to avoid excessive mining of barren rock. A long section showing the variation in grade \times thickness of the mineralized zone from both the Stillwater mine as well as from the East Boulder mine is shown in figure 13. This figure illustrates both the local scale variation in mineral grade and larger regional differences. The podiform to lenticular nature of the mineralization is evident, with well-mineralized regions separated by regions of poorly mineralized or barren rock. As noted above, mineralization is largely absent from the Dow Meadow Depression where the reef cuts down into the rocks of the underlying N-I and GN-I zones, although where present it can have very high grades. West of the depression mineralization is somewhat more continuous and with less lateral grade thickness variability than to the east. As detailed below, this regional variability is also seen in variations in the (Pt+Pd)/(Ni+Cu) ratio, with the eastern regions having a higher Pt + Pd tenor.

Ballrooms

“Ballroom” is an informal term used by mine geologists to refer to local thickening of disseminated PGE–sulfide mineralization that extends up to 30 m into lithologies below the usual stratigraphic level of the J-M Reef. It is both the variable thickness of mineralization in general and the ballrooms in particular that are responsible for some of the variations in the grade \times thickness map of figure 13. Ballrooms likely include some of the mineralization referred to by Raedeke and Vian (1986) as Footwall zone, Basal zone, and Main zone mineralization. Sulfide textures are generally similar to those found in the main J-M Reef, although coarse sulfides with irregular radial “veinlets” up to 5 cm long are a common feature of gabbronorite- and norite-hosted ballroom mineralization (Childs and others, 2002). Childs and others (2002) and Harper (2004) have described three general types of ballrooms:

Type 1 Ballroom

The most common ballroom type is a thickening of the mineralized zone extending from roughly the lower contact of the Anorthosite subzone of OB-I into the lower units of OBI. In the areas of the regional unconformity this deeper mineralization occurs in the underlying GN-I and N-I zone rocks. Examples from the Stillwater mine are shown in figure 14. Mineralization crosses existing modal layering, but there otherwise appears to be no major or trace element difference between mineralized ballroom rock and adjacent barren rocks except for the presence of PGE–sulfides (Harper, 2004). In this regard, they are different from “pothole” structures seen in the Bushveld Complex, which are roughly circular areas where parts of the stratigraphy are missing or replaced. In the Stillwater case the mineralization appears to have no relation to the presence of igneous layering. Nor do the ballrooms appear to represent regions where sulfide “settled” into the pore space in underlying rocks while the rocks were still partially liquid, as mineralization can continue up to the nominal contact with the base of the Anorthosite subzone of OB-I. In this sense the ballrooms represent an excess of sulfide mineralization, much more than is normally observed. The distribution of ballroom mineralization in this instance has some similarities with the disseminated sulfide mineralization seen in the Picket Pin sulfide zone at the top of the Middle Banded series, where the better mineralized regions have

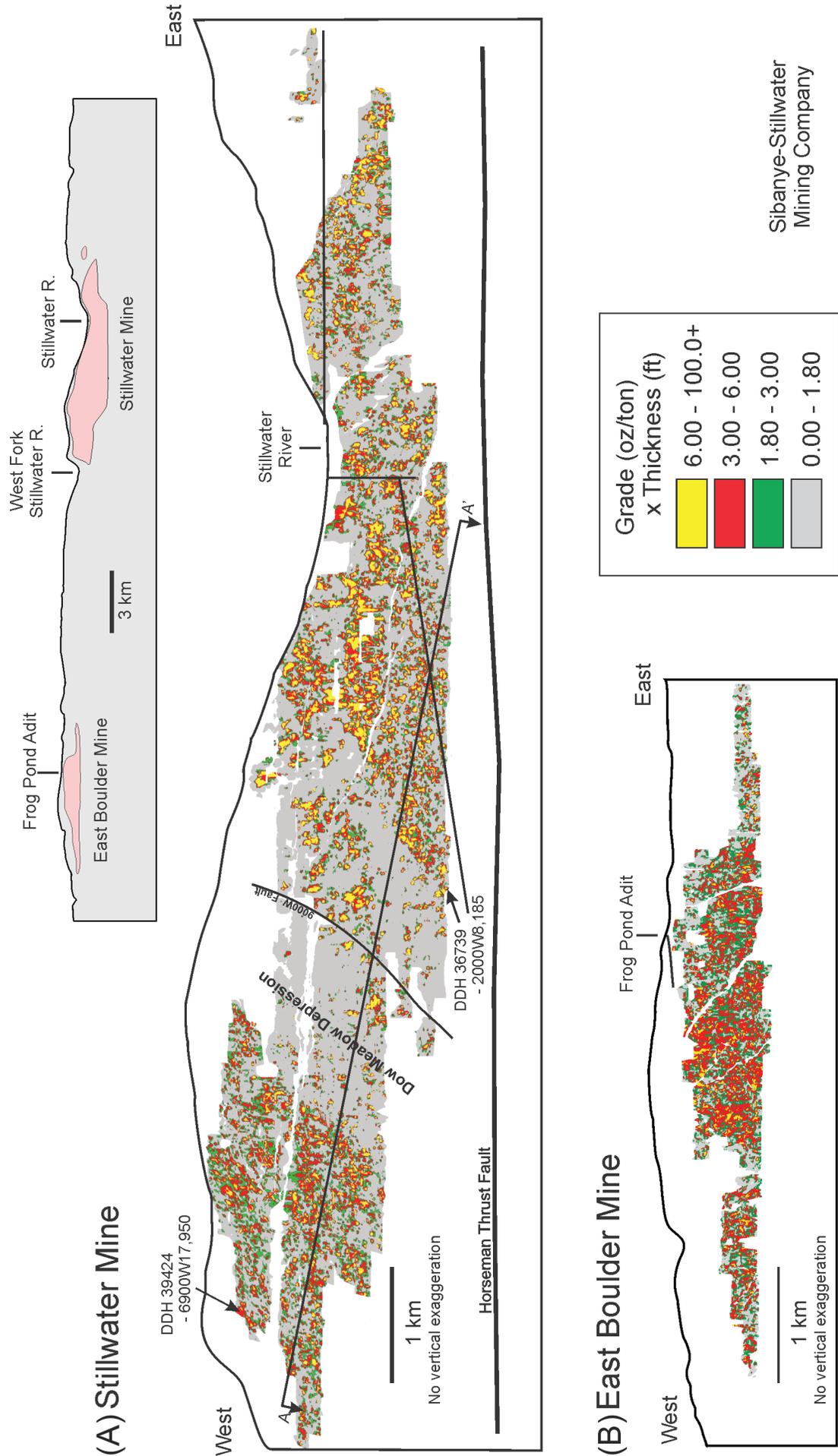


Figure 13. Long section of grade x thickness along the J-M Reef illustrating the non-uniform nature of mineralization along the mineralized surface in (A) the Stillwater mine and (B) the East Boulder mine. Inset at upper left shows location of the currently defined areas (pink). For section a), note especially the lack of mineralization in the region of the Dow Meadow Depression (fig. 12). The section line A-A' of figure 12 and the locations of drillholes from figure 15 are also shown.

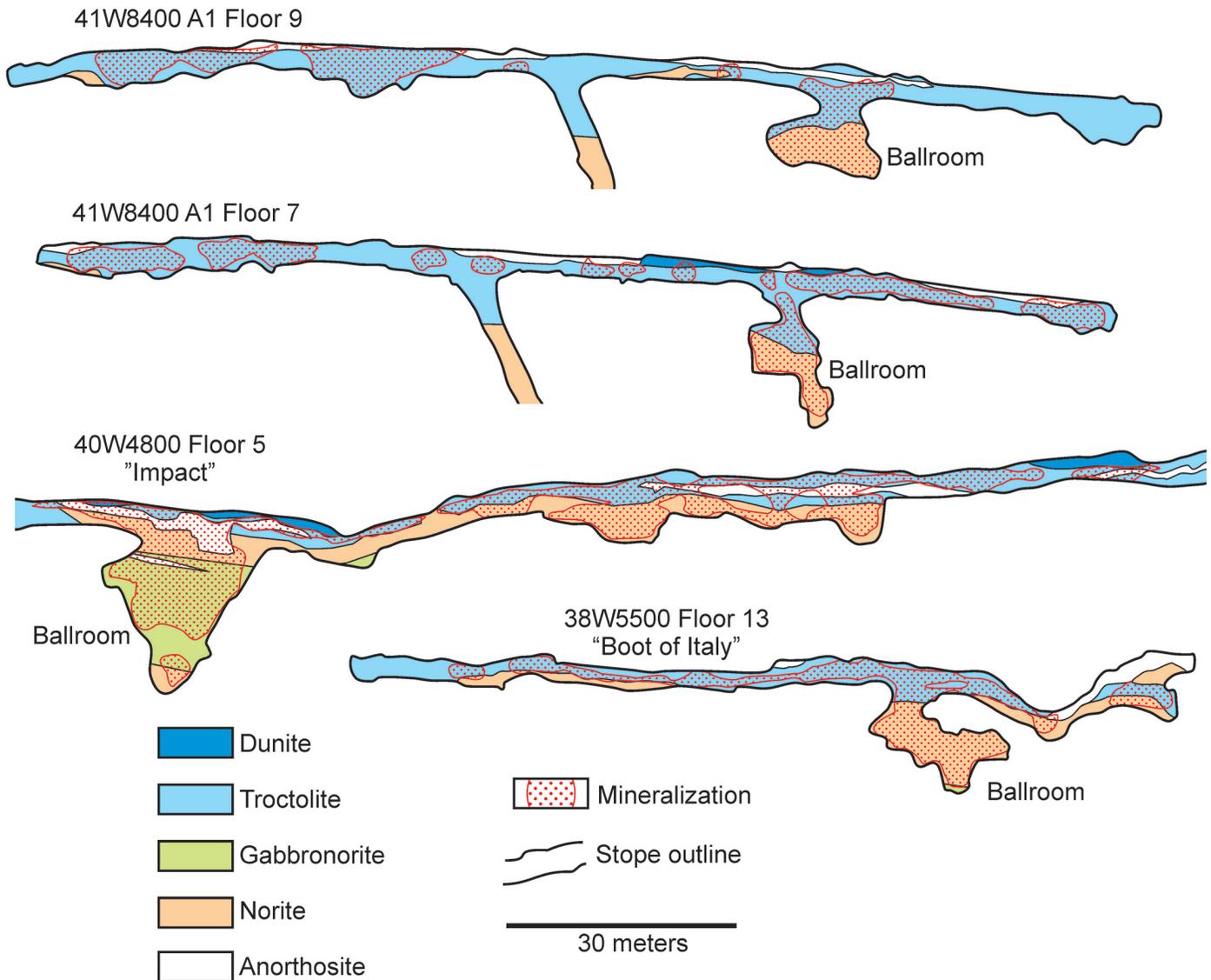


Figure 14. Examples of ballroom mineralization as defined by underground mining illustrating the lack of lithologic control on the distribution of PGE-sulfide mineralization. Each figure shows the plan view of a stope on the steeply dipping to near-vertical mineralized zone; thus the plan view shows the stratigraphic section with stratigraphic “up” to the top. The top two plan views are from the same 41W8400 A1 stope but at different heights (floor levels, in the case ~6 m vertical separation), with the ballroom extending into the foot-wall norite and gabbronorite as noted. The 40W4800 Floor 5 view shows both a typical Type 1 ballroom and a small Type 3 isolated ballroom beneath it. Note also the discontinuous nature of mineralization along the stratabound portion of the J-M Reef.

pod and trails of sulfide mineralization leading up to them from below (Boudreau and McCallum, 1985; see below).

Type 2 Ballroom

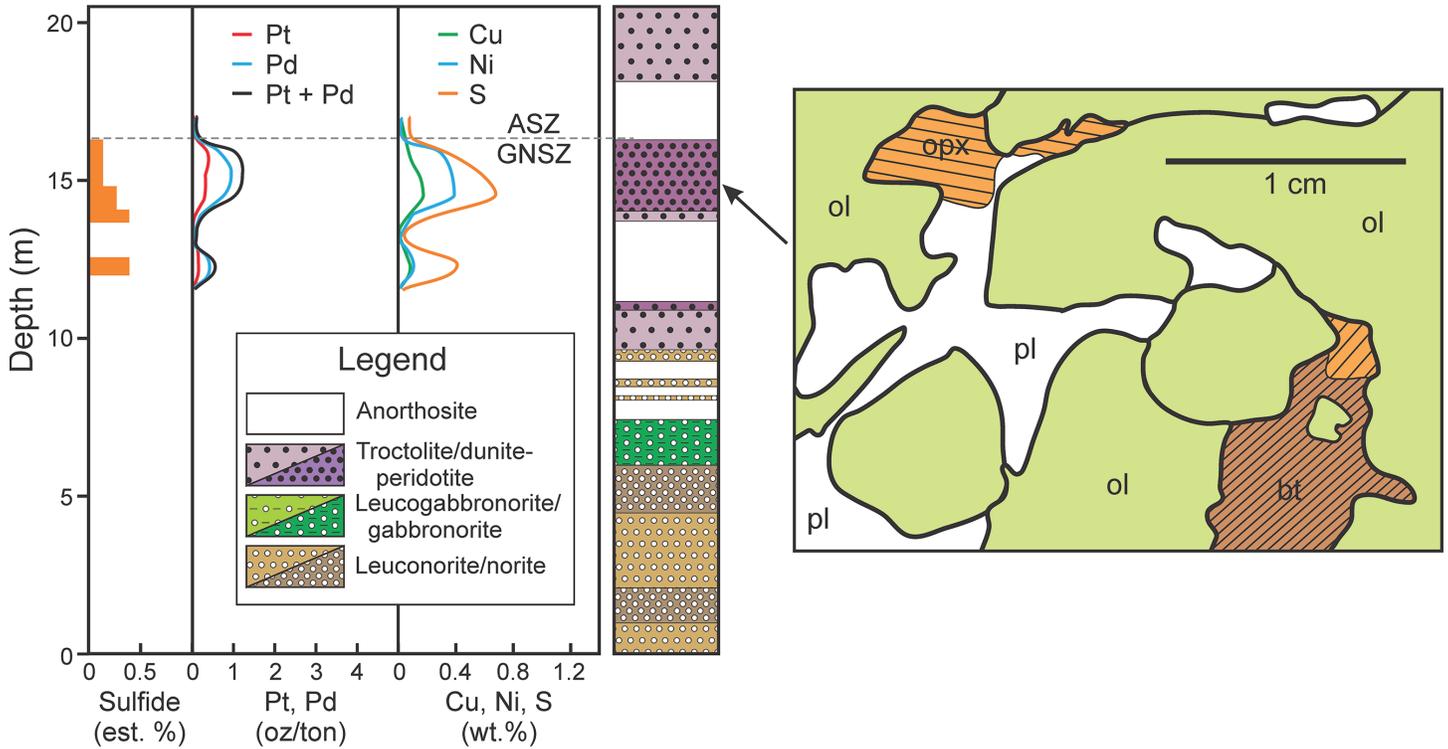
The second type of ballroom involves a thickening of mineralization within a thickened olivine-bearing package of rock. Unlike the type 1 ballrooms, this second type is broadly analogous to the potholes of the Bushveld Merensky Reef, although the latter mineralization is mostly confined to the contact regions and does not occupy the whole volume as the Stillwater ballroom mineralization does. Again, the type 2 ballrooms represent overall higher amounts of mineralization than found in “normal” reef.

Type 3 Ballroom

The least common type occurs as apparently isolated pockets of mineralization in footwall lithologies approximately 10 m below the main level of the J-M Reef (fig. 14). Mineralization can occur with isolated, podiform bodies of olivine-bearing rock that can be either stratabound or discordant to local layering, similar to replacement troctolite that occurs in OB-II and throughout the Middle Banded series (Boudreau, 2016). The disconnected, podiform nature of the disseminated mineralization is also similar to that seen in the low-grade Picket Pin Pt/Pd deposit that occurs at the top of the Middle Banded series described below.

Figure 15 shows mineralization in two drillholes,

(A) DDH 39424 - 6900W17,950



(B) DDH 36739 - 2000W8,185

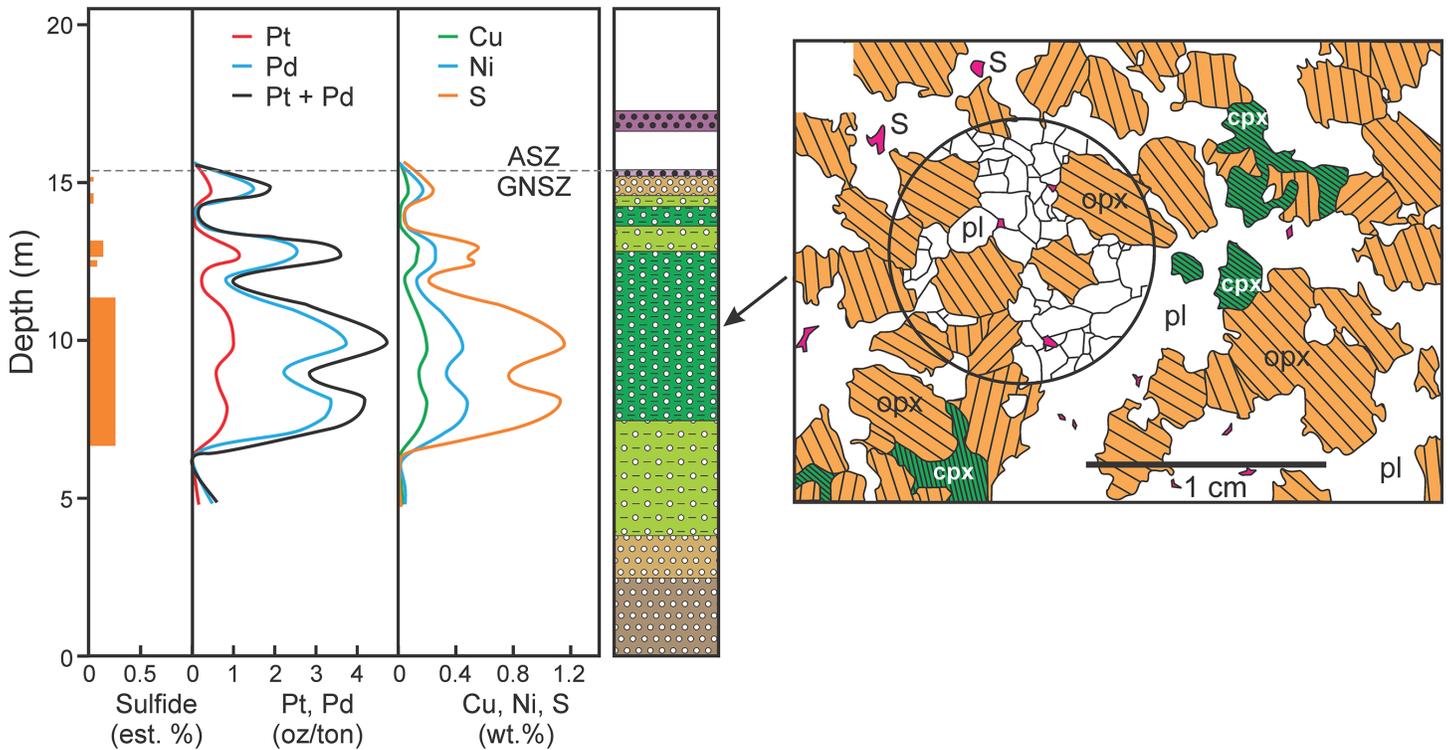


Figure 15. Example drillholes through (A) “normal” and (B) “ballroom” reef illustrating the variable thickness, host lithology, and tenor of mineralization. Thin section sketches show typical textures of the (A) mineralized olivine-rich rocks and (B) the mineralized gabbronorite, respectively. For (B), the circle on the thin section sketch shows more mineral/textural detail for the plagioclase. Drillhole locations as shown in figure 12. Abbreviations: ASZ/GNSZ, contact of the gabbronorite subzone of OB-I with the overlying Anorthosite subzone; ol, olivine; pl, plagioclase; opx, orthopyroxene; cpx, clinopyroxene; bt, biotite; S, sulfide.

DDH 39424 - 6900W17,950 through a normal reef mineralized zone and DDH 36739 - 2000W8,185 through a mineralized ballroom. For the normal reef, mineralization is typical in that it occurs at and below the contact between the gabbronorite subzone and the overlying anorthositic subzone, where it mainly occurs within a pegmatoidal olivine-rich lithology. In contrast, there is very little olivine in the ballroom drillcore and mineralization instead is hosted by medium-grained gabbronorites typical of Gabbronorite zone I. In both cases, PGE grades can be correlated with observable sulfides, but the general PGE tenor of mineralization is higher in the ballroom.

Regional Variations in Grade

In addition to the variable thickness of the mineralization ranging from normal to ballroom mineralization, there is a marked west to east change in the tenor of the sulfide mineralization (fig. 16). To the west of the Dow Meadow Depression and continuing into the East Boulder mine area, mineralization has a lower total Pt + Pd concentration relative to both the amount of visible sulfide and the bulk rock Cu + Ni concen-

tration. As one goes east, however, the Pt and Pd tenor increases relative to Cu + Ni. This appears to be related to the more localized nature of mineralization to the east. Whatever is responsible for the variable thickness mineralization, including ballroom mineralization, also appears to produce a higher tenor mineralization.

Although complicated by faulting, the regional variation in mineralization in the J-M Reef are broadly associated with regional changes in the thickness of the Ultramafic zone. In the western Stillwater, where mineralization along the J-M Reef is much more continuous that it is to the east, the underlying Ultramafic series at Chrome Mountain is only 840 m thick. In contrast, in the eastern portion where J-M Reef mineralization occurs in more localized to high “grade x thickness” regions, the underlying Ultramafic series is 2,065 m at Mountain View (Raedeke and McCallum, 1983). It is also noted that the mineralization-poor Dow depression broadly occurs between these two regions.

Raedeke and McCallum (1983) suggested that the Mountain View section formed in a subsiding region whereas the Chrome Mountain section formed on a

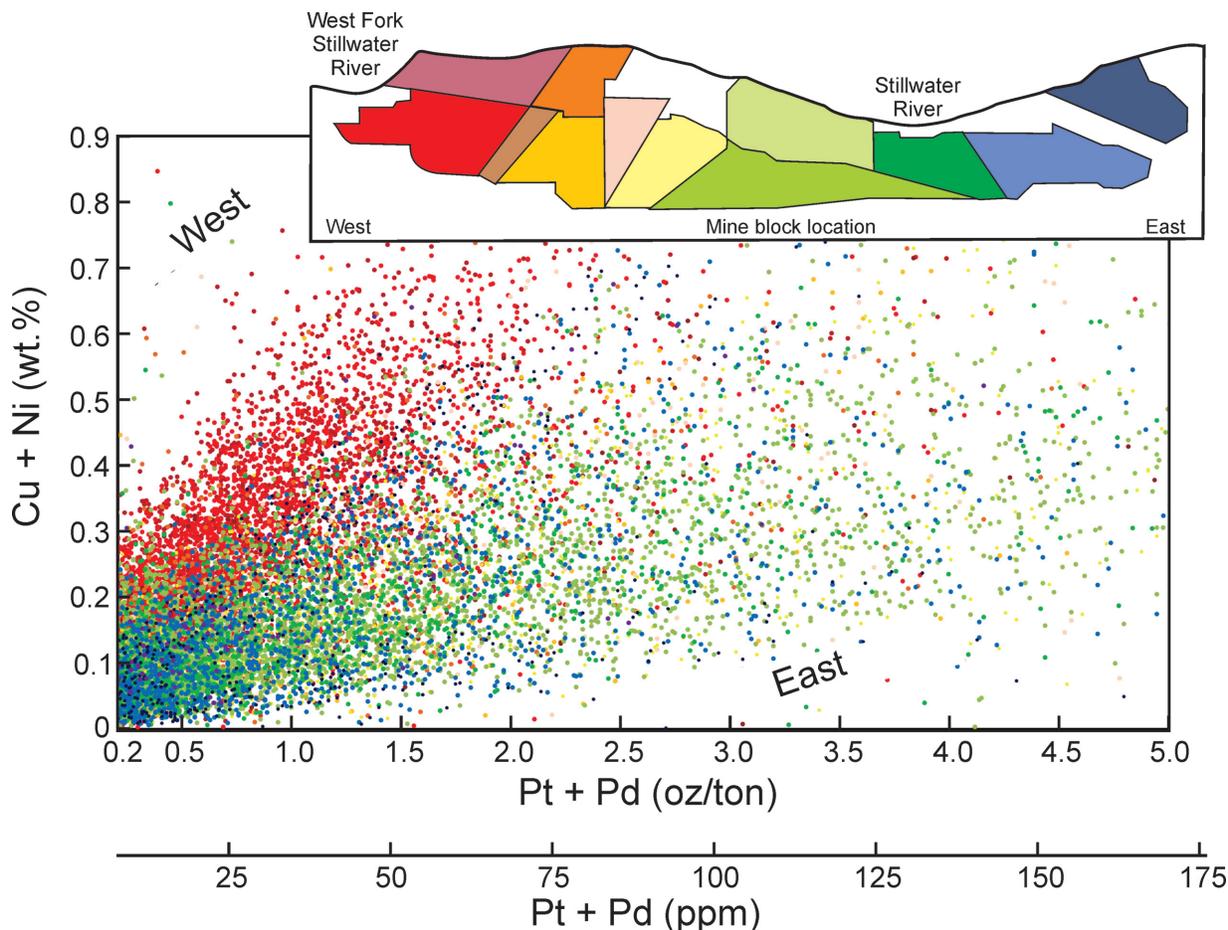


Figure 16. Plot of bulk rock Cu + Ni against Pt + Pd illustrating the general increase in PGE tenor from west to east in the Stillwater mine. Data is color coded to the mine blocks as shown in the inset.

stable, higher area of the chamber floor. Extensive variations in thickness also occur in the Lower Zone rocks of the Eastern Bushveld complex, on a regional scale similar to that seen in the Stillwater. The floor rock deformation associated with the Bushveld Complex present in the field as domes and periclinal folds, and were first mapped in the 1960s and 1970s. Most interpretations view the regions of upwelling as diapirs (Button, 1978; Uken and Watkeys 1997; Gerya and others, 2003, 2004; Clarke and others., 2005) driven by the heating of the floor rocks and density difference between the mafic complex and the underlying sedimentary layers.

Pegmatoids

Pegmatoids are locally common in the upper part of the Bronzite zone, along the contact of the Lower Banded series contact with Ultramafic series, and in the upper parts of GN-I and in the Gabbronorite subzone of OB-I. The reappearance of pegmatoids in GN-I coincides with locally significant layer deformation (fig. 9B). The pegmatoids are composed of the same minerals (plagioclase, orthopyroxene, and clinopyroxene \pm olivine) as the host rocks and in approximately the same proportions, but with grains that may be in excess of several tens of centimeters in length and the pyroxene variably replaced by horn-

blende. The whole-rock compositions of the pegmatoids are enriched in lithophile incompatible major and trace elements relative to the host rocks (Braun and others, 1994). Locally, the pegmatoids can have cores of massive sulfide up to about 0.5 m in diameter. The example shown in figure 9C occurs adjacent to a pothole-like feature in the Gabbronorite subzone of OB-I. The Pt + Pd vs Cu concentrations of two other examples of PGE-enriched pegmatoids beneath the J-M Reef are shown in figure 17. The Bronzite zone samples consist of a pegmatoidal orthopyroxenite with disseminated sulfide minerals and local concentrations of graphite occurring just below the contact with the Banded series (Volborth and Housley, 1984). The Gabbronorite zone I samples are described as an orthopyroxene-rich pegmatoid in GN-I below the J-M Reef and a dike-like body of orthopyroxenite that cuts the Lower Banded series rocks (McIlveen, 1996).

Hanley and others (2005, 2008) describe a variety of fluid inclusions in the pegmatoids below the J-M Reef that they interpret to represent magmatic-hydrothermal to lower temperature metamorphic fluids. Those interpreted to be high-temperature primary inclusions range from hydrosaline to carbonic fluids and can have up to 16 precipitated solid phases. They note that about 10% of the brine inclusions have significant concentrations of Pt, ranging from 0.4 to 4 ppm. Cu/Pt

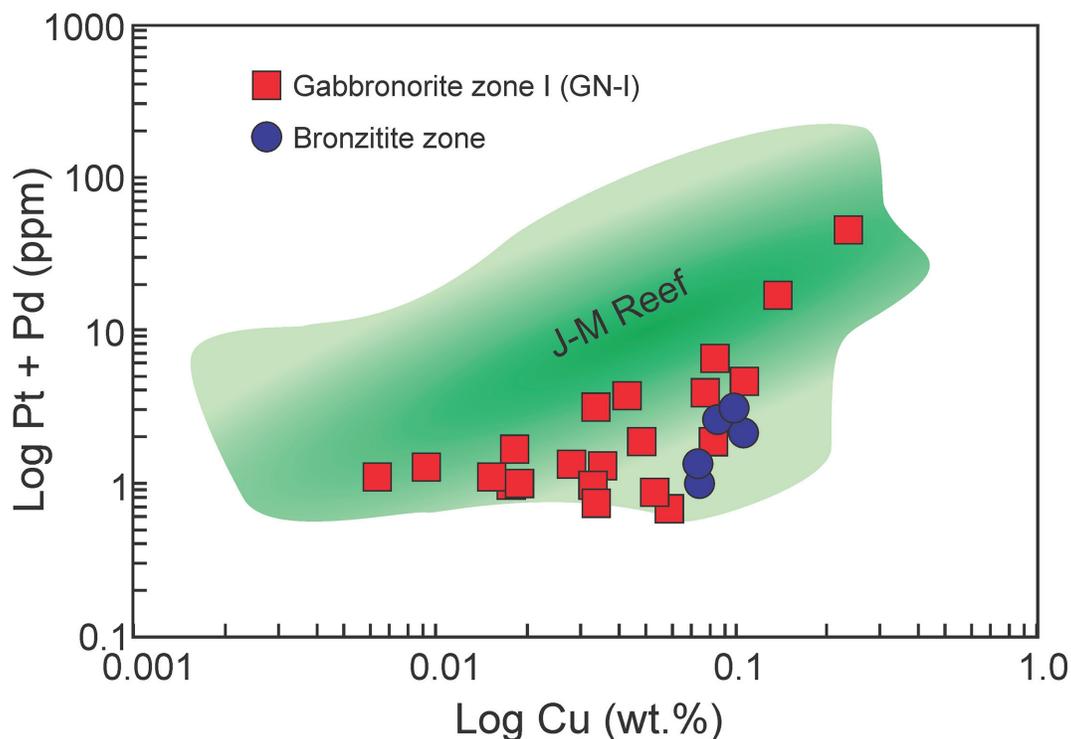


Figure 17. A comparison of mineralized pegmatoids from the upper Bronzite zone and Gabbronorite zone I beneath the J-M Reef with the compositional field of the J-M Reef (Shaded field, darker represents higher data point density). Note the generally similar Pt + Pd concentrations but higher base metal concentrations relative to the J-M Reef. Redrawn after Zientek and others (2002), with permission from the Canadian Institute of Mining, Metallurgy and Petroleum.

ratios in the brine range from 27 to 342, which compares with the Cu/Pt ratio of ~ 400 for the J-M Reef.

MODELS FOR THE FORMATION OF THE J-M REEF

Models for the petrogenesis for OB-I and the J-M Reef range from strictly magmatic models involving mixing of magmas and sulfide saturation in the hybrid liquid (e.g., Irvine and others, 1983; Campbell and others, 1983; Campbell and Barnes, 1984; Barnes and Naldrett, 1986; Wooden and others, 1991; Lambert and Simmons, 1988; Lambert and others, 1994) to those that involve volatile fluids percolating up from below (Boudreau and others, 1986; Boudreau, 1988, 1999, 2019; Boudreau and McCallum, 1989, 1992). These endmember models have been informally called the “downer” and “upper” models (Barnes and Campbell, 1988), and both have had substantial support from studies of the Stillwater Complex. Although details among the various orthomagmatic downer models vary, these studies have attributed crystallization changes to two distinct parent magma types. Below OB-I, the rocks are interpreted to have largely crystallized from a “U” magma of “ultramafic” parentage and record the apparent crystallization order olivine ± chromite → orthopyroxene → orthopyroxene + plagioclase → orthopyroxene + plagioclase + augite. This progression is interrupted in OB-I where the rocks now include troctolite, anorthosite, and dunite. This change is proposed to have resulted from a mixing event that blended the evolved “U” liquid with an anorthositic “A” liquid that had plagioclase as the first mineral to crystallize. In these models, the observed stratigraphic modal changes document variations in primary crystallization resulting from these magma mixing events, and volatiles are assumed to have played only a secondary role in modifying original textures (Barnes and Campbell, 1988). Immiscible sulfide liquid separating from this hybrid magma and settling down through a thick column of magma acts as an effective collector for the PGE owing to very high sulfide liquid/silicate liquid partition coefficients on the order to 10,000 or higher (e.g., Campbell and Naldrett, 1979; Campbell and others, 1983).

In contrast to these orthomagmatic models, the “upper” hypothesis interprets the mineral changes in OB-I as the result of incongruent melting of pre-existing, partly molten rock by the fluxing effect of the upward percolation of mineralizing hydrothermal

fluids (Boudreau, 1988, 1999, 2016). These “hydro-magmatic” models are analogous to the hydration melting models proposed for arc magmatism, where it is envisioned that volatiles coming off a subducting slab induce partial melting as they move into the overlying hot mantle wedge. In the case of the Stillwater Complex, the fluid came from the crystallization of the interstitial liquid in the mush below the J-M Reef. This fluid moved upward into hotter rocks that had not yet reached vapor saturation. The hydromagmatic model proposes that mineralization was introduced by the upward migrating Cl-rich vapor that redissolved into vapor-undersaturated liquids near the top of the crystal pile. Furthermore, this hydration event led to the incongruent melting of what were originally pyroxene-rich lithologies to produce olivine and minor chromite as discussed above in regards to chromitites of the Ultramafic series. The partial melt that resulted from the hydration and incongruent melting of the gabbro-norite was nominally enriched in felsic components as the mafic components were left in the olivine-rich residue. The anorthosite and troctolite of the Anorthosite subzone of OB-I represent mixing of this felsic partial melt generated in the crystal pile with the drier magma above the floor of the chamber.

Boudreau (2016) summarized the evidence for the fluid introduction of the mineralization, which also led to the formation of olivine ± chromite by incongruent flux melting: (1) The pegmatoidal nature of the mineralized rocks, abundance of PGE-bearing pegmatoids occurring well below the J-M Reef into the Ultramafic series, and relative abundance of hydrous minerals associated with mineralization all point to high volatile element abundances. In contrast, pegmatoids are not found in the Anorthosite subzone of OB-I and in the immediate overlying units of N-II. (2) There is an abrupt change in halogen concentration in apatite at the level of the J-M Reef, from Cl-rich below the Reef to Cl-poor above. (3) The metal ratios in fluid inclusions in the pegmatoids below the J-M Reef are broadly similar to the metal ratios seen in the J-M Reef (Hanley and others, 2008). (4) The common presence of hydrous melt inclusions (now crystallized to polymineralic assemblages) in olivine and chromite of the J-M Reef is consistent with the growth of these minerals from a hydrated liquid. (5) Mineral compositional trends from GN-I up into N-II, particularly for plagioclase, do not show offsets expected from injection of more primitive magma. (6) The rounded and reversely zoned plagioclase found in the olivine-bearing

ing rocks is consistent with the partial remelting of plagioclase. (7) The pothole-like structures are analogous to seafloor pockmarks that can form where rapid fluid escape from the underlying sediments blows out sediment from the seafloor (e.g., Boudreau, 1992).

The origin of the apparent erosional features such as the Dow Meadow Depression could be related to the above mechanism. Boudreau (1995) interpreted the unconformity, the thinning of both the olivine-bearing and overlying anorthosite units, and the poor mineralization to be the result of slumping and local loss of a reactive volatile-rich mush. He suggested that soft-sediment-like folds seen in the footwall rocks was evidence of this (e.g., Page and Moring, 1990; fig. 9C). However, olivine-bearing lithologies of clearly replacement nature are seen throughout the Middle Banded series (e.g., Meurer and others, 1977). As noted above, in the type section in the Frog Pond area, there is no break in the trend of plagioclase going from the rocks below to those above the J-M Reef. An analysis of mineral compositions going across the apparent unconformity should answer the question of whether it is a true unconformity or if the olivine-bearing lithologies are a late replacement that is simply happening at different levels within the crystal pile. A true unconformity in which material was physically lost should show a marked offset of plagioclase compositions to that seen in N-II in the Frog Pond area. If a replacement, then there might not be any significant offset going across the disconformity, similar to the lack of offsets seen in the Frog Pond area.

A number of explanations have been proposed for the origin of the discontinuous distribution of sulfides, the regional variations in the degree of concentration of the sulfide of mineralization shown in figures 12 and 13, the ballroom mineralization, and the variation in PGE/base metals as shown in figure 16. For example, Raedeke and Vian (1986) suggested that laterally discontinuous sulfide precipitations were the result of sulfide precipitation controlled by small-scale convection currents in the magma near the floor of the chamber. It is not clear, however, why this would affect only the sulfides while the major minerals continued to form horizontal layering.

The observation that the disseminated mineralization crosses the modal layering suggests that it formed after the igneous layering. In this regard, Chung and Mungall (2009) suggested that the sulfide can settle through interconnected pore space into the underlying

mush. However, the ballrooms are characterized by being regions of overall thicker sections of mineralization and are not just the same amount of sulfide in typical reef displaced to a lower level. Nor does it account for the discontinuous nature of the mineralization. A settling mechanism would require the sulfide move laterally towards the ballroom mineralization as well as downward

The hydrothermal model would suggest that the crosscutting ballrooms represent fossil channelways for the upwelling mineralizing fluids, similar to the disseminated sulfide-bearing pipes seen in the Picket Pin PGE zone described below. The more widely spaced and more concentrated mineralization seen in the eastern part of the complex than the west implies that these mineralizing channels similarly were more widely spaced in the east. Raedeke and McCallum (1984) note that the Ultramafic series at Mountain View, overlooking the Stillwater River in the eastern part of the complex, is about 1.2 km thicker than the section at Chrome Mountain in the west. Degassing of the thicker eastern section allowed more vertical distance for fluids to become concentrated into fewer channels, much as thick pyroclastic ash flow sheets will form well-developed fumaroles toward their upper surfaces. The net result is more discontinuous, but more concentrated, mineralization in the east (fig. 18).

Fluid inclusion evidence from pegmatoids suggests that these fluid channelways can persist to subsolidus conditions (Hanley and others, 2008), allowing later fluids to percolate through the mineralized ballroom. These can have only a modest effect on the PGE but can cause some loss of the base metals and sulfur, resulting in the distinctive PGE/base metal difference from east to west. This later, but still high, temperature alteration would be similar to that proposed by Kinloch (1982), who noted that potholes in the Merensky Reef of the Bushveld Complex are characterized by an abundance of platinum-group metal alloys and a paucity of PGE-sulfides as compared with "normal" Merensky Reef. Kinloch suggested that the potholes define zones of channeled fluid loss from the footwall rocks through which later fluids preferentially caused S loss from the originally sulfide-dominated assemblage. A similar situation could have occurred in the J-M Reef, differing only in that in the Stillwater case, alteration was more moderate, leading to a partial loss for the more soluble S and base metals.

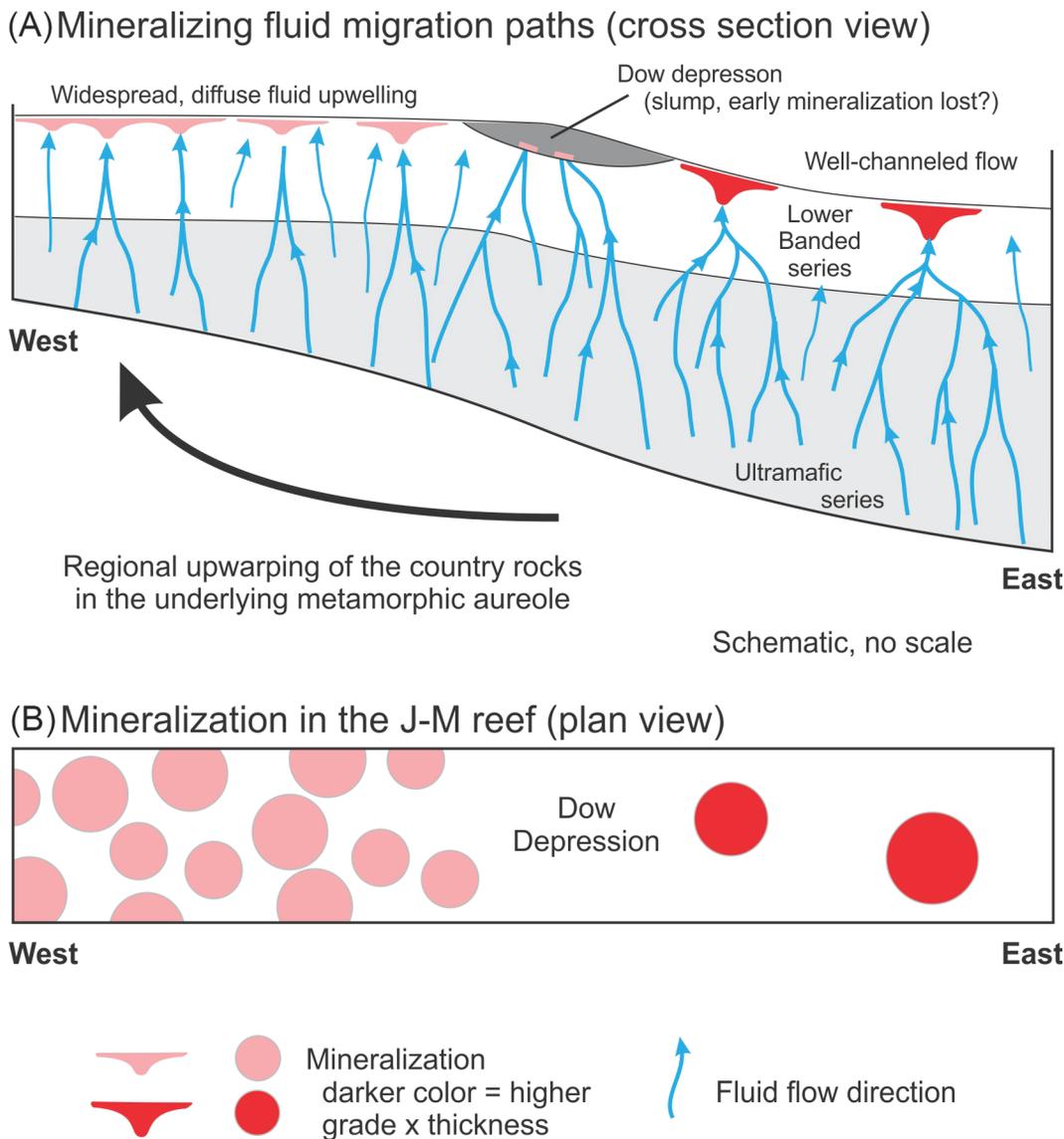


Figure 18. Cartoon showing the development of the J-M Reef by a hydromagmatic model. (A) Heating of the floor rocks and regional upwelling leads to a thicker Stillwater section in the east half of the Stillwater complex relative to that in the west. Mineralizing fluids migrating upward from the crystal mush below the J-M Reef become increasingly localized to fewer channels in the thicker eastern section. The result is fewer but higher “grade x thickness” regions in the east than in the west, as shown schematically in (B). In addition, the regional upwelling produces a localized steepening of the section, leading to slumping and loss of earlier-formed mineralization in the region of the Dow depression.

THE PICKET PIN PGE–SULFIDE ZONE OF THE UPPER ANORTHOBSITE ZONE

Sulfides modestly enriched in the platinum-group elements occur in the coarse-grained anorthosite immediately below the medium-grained anorthosite member of the Upper Anorthosite zone and have been described by Boudreau and McCallum (1985, 1986), Corkery (2002, 2003), and Boudreau (2016). The sulfide zone is stratabound in the sense that it is traceable at the same stratigraphic interval over 22 km of exposed strike length. Like the J-M Reef, however, sulfide mineralization is not continuous on the local scale, being erratically developed as podiform and lenticular accumulations of 1–5% disseminated sulfides

(mainly chalcopyrite, pyrrhotite, and minor pentlandite; fig. 19). Discordant trails of spotty disseminated sulfides and larger transgressive sulfide-rich “pipes” occur to depths of about 150 m into the Upper Anorthosite zone. These discordant zones of mineralization typically underlie the better-mineralized sections near the top of the anorthosite, mimicking the ballroom structures seen in the J-M Reef noted above.

Sulfide mineralization is about an order of magnitude lower grade than the J-M Reef, the rocks rarely containing more than 1–2 ppm combined Pt + Pd. Mineralization is most commonly associated with the pyroxene-poor, incompatible element-rich parts of the Upper Anorthosite zone, where it is associated with



Figure 19. Photo of the typical weathered podiform occurrence of the disseminated PGE-sulfide mineralization of the Picket Pin Pt-Pd deposit that occurs in the upper part of the Upper Anorthosite zone at the top of the Middle Banded series. Note hammer (at arrow) for scale.

accessory quartz and apatite. Locally, sulfide forms the core of small (10–20 cm) patches of pegmatoidal anorthosite; the sulfide is badly weathered, but what remains is predominantly pyritic. Some sulfide occurs in the medium-grained anorthosite member at the top of the Upper Anorthosite zone, but these sulfides are PGE-barren. Sulfide is also rare in the overlying barren troctolite of OB-V.

The transgressive zones of sulfide leading up to the main sulfide concentrations, the association of the sulfides with the late crystallizing minerals quartz and apatite, and the incompatible element-rich nature of the sulfide pipes and pods led Boudreau and McCullum (1986) to the interpretation that the ore component was introduced by the percolation of mineralizing solutions during solidification of the Upper Anorthosite. They interpreted the discordant pipes and sulfide trails as fossil channelways, or the plutonic equivalent of fumarolic zones, through which both late evolved liquid and the mineralizing fluids migrated upward. Corkery (2002, 2003) noted that patches and stringers of clinozoisite–epidote–quartz lead up-section to the PGE-enriched sulfide intervals. Assay values of these clinozoisite patches away from obvious sulfide min-

eralization are up to 600 ppb combined PGE, whereas the surrounding rocks are barren. The alteration mineralogy is zoned about the sulfide: epidote–magnetite–albite \pm quartz occurs away from the sulfides and clinozoisite–quartz–albite occurs in contact with the sulfides. Corkery suggested that a hypersaline hydrothermal fluid at an approximate temperature of 500–550°C transported the PGE to enrich a preexisting sulfide; he did not otherwise explain the podiform distribution of sulfide in the Upper Anorthosite zone. Boudreau (2016) suggested that mineralization began as a high-temperature event (a mineralizing fluid in equilibrium with the igneous silicate assemblage) but that fluid continued to move through the anorthosite as temperature fell over time to produce the later, but still relatively high, temperature alteration. The observed PGE enrichments below the sulfide pods were attributed to a chromatographic effect as sulfur moved upward and left some of the PGE behind.

CONCLUSIONS

The result of numerous studies past and ongoing, as well as the extensive mine development, are still yielding clues as to the nature and origin of the

impressive mineral reserves of the Stillwater Complex. Vigorous debate over the petrogenesis of these deposits is still very much ongoing. However, unlike intrusions such as the Bushveld Complex of South Africa and the Skaergaard Intrusion of east Greenland, much of the work has concentrated on the ore zones themselves, whereas much of the rest of the complex has received only scant attention. It is clear that much detailed geochemical and petrographic detective work remains to be done.

ACKNOWLEDGMENTS

This work represents the efforts of numerous Sibanye–Stillwater geologists, with a special thanks to Diane Wolfram. It was improved by the helpful reviews of Michael Zientek of the U.S. Geological Survey and Edmond Mathez of the American Museum of Natural History. This work is dedicated to the memory of I.S. McCallum, who has put in a lifetime of work on the Stillwater Complex.

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