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¹⁹⁰Pt–¹⁸⁶Os and ¹⁸⁷Re–¹⁸⁷Os systematics of the Sudbury Igneous Complex, Ontario

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Abstract—We measured by negative thermal ionization mass spectrometry (NTIMS) Re, Os and ¹⁸⁶Os/¹⁸⁸Os and ¹⁸⁷Os/¹⁸⁸Os in 26 samples of 18 Ni–Cu sulfide ores from the Falconbridge, McCreedy West, and Strathcona mines at Sudbury, Ontario. At McCreedy West and Falconbridge, the isochron Re–Os ages are 1835 ± 70 Ma and 1827 ± 340 Ma, and the initial ¹⁸⁷Os/¹⁸⁸Os ratios 0.514 ± 0.019 and 0.550 ± 0.024, respectively. The ages agree with the canonical value of 1850 ± 1 Ma for the Sudbury Igneous Complex (SIC). For Hangingwall and Deep Zone ores at Strathcona, the age of 1780 ± 7 Ma may reflect resetting by dyke activity. The high initial ¹⁸⁷Os/¹⁸⁸Os of 0.934 ± 0.005 in these ores is distinct from those at McCreedy West and Falconbridge. Strathcona Deep Copper Zone ores have highly radiogenic Os giving a mean model age of 1883 ± 54 Ma that is similar to ages at McCreedy West and Falconbridge, but distinct from other Strathcona sulfides. Initial ¹⁸⁶Os/¹⁸⁸Os in two Strathcona ores with low ¹⁹⁰Pt/¹⁸⁸Os average 0.119 826 ± 0.000 009 (*n* = 3) and 0.119 827 ± 0.000 004 (*n* = 3), respectively, with a grand mean of 0.119 827 ± 0.000 003. This ratio may be slightly lower than the chondritic value at that time. Similar ores at Falconbridge and McCreedy West show more scatter, averaging 0.119 855 ± 0.000 008 (*n* = 6) and 0.119 867 ± 0.000 020 (*n* = 3), respectively. These values are substantially suprachondritic. The Re–Os isotope systematics of Sudbury ores are clearly of crustal origin and may be derived from a binary mixture of Superior Province and Huronian metasedimentary rocks, with Strathcona, Falconbridge, and McCreedy West ores containing, respectively, 55%, 16%, and 12% of Os from Superior sediments. The suprachondritic ¹⁸⁶Os/¹⁸⁸Os at McCreedy West and Falconbridge may be due to admixture of Archean or Paleozoic mafic rocks with ¹⁹⁰Pt/¹⁸⁸Os ≈ 0.1. No trace of the asteroid that produced the Sudbury Structure has been reported. At the Whistle mine S-poor olivine melanorite inclusions with high Ni and Os and low ¹⁸⁷Os/¹⁸⁸Os may contain the signature of a magmatically fractionated asteroidal core contributing 1 to 2.5 % metal. The S-poor melanorite Ni and Os data are equally well explained by admixture of ≈40% mantle peridotite, however. *Copyright © 2002 Elsevier Science Ltd*

1. INTRODUCTION

Meteorite impact is bad for dinosaurs, but good for the nickel industry. The Sudbury Igneous Complex (SIC) in Ontario is presently the world's largest Ni producer and lies within the Sudbury impact structure (Boerner et al., 1994; Deutsch and Grieve, 1994; Dressler et al., 1996). The Sudbury structure is situated at the contact of the Archean basement rocks of the Superior Province and Early Proterozoic Huronian supracrustal rocks of the Southern Province, close to where these are truncated by the Middle Proterozoic Grenville front. The structure consists of three elliptical units that are roughly confocal and trend along a southwest to northeast major axis; the outer Sudbury breccia, the SIC itself, and the Whitewater group that fills the central basin. The Ni–Cu ores are associated with the SIC, which may be interpreted as the inner remnant of the melt sheet of a 280 km diameter impact crater that was formed 1850 ± 1 Ma ago (Krogh et al., 1996). The SIC is now 60 km by 27 km (Fig. 1), with its elliptical shape being truncated at the northeast end by the 37 Ma Wanapitei Lake impact structure (Dence and Popelar, 1972; Winzer et al., 1976). Results of the

LITHOPROBE transect of the Sudbury structure, and more particularly of the SIC, indicate that the elliptical shape is due to significant NW-trending compression of an originally near-circular structure (Milkereit et al., 1994; Boerner et al., 2000). Only the northern rim of the SIC has retained a relatively undeformed arc of the central part of the eroded crater (Deutsch and Grieve, 1994).

The SIC is a layered intrusion with compositions ranging from quartz norite at the base, through gabbro, to a granophyric cap. These rocks are unusually silicic to be derived directly from the mantle, and indeed markedly negative ε_{Nd} values suggest a predominantly (Naldrett et al., 1986) or entirely crustal origin (Faggart et al., 1985). The Ni–Cu (+PGE) ores occur at the base of the complex. The sulfides may be disseminated and interstitial to silicates in a distinctive group of intrusive rocks called the Igneous Sublayer. Alternatively, the sulfides may have been intruded into the highly fractured country rocks beneath the sublayer. Re–Os isotope systematics of ores from four Sudbury mines were studied previously by two relatively low-precision methods; resonance ionization mass spectrometry (RIMS; Walker et al., 1991; 1994) and single collector, inductively coupled plasma source mass spectrometry (ICP–MS; Dickin et al., 1992).

These data were consistent with a formation of the ores at ca.

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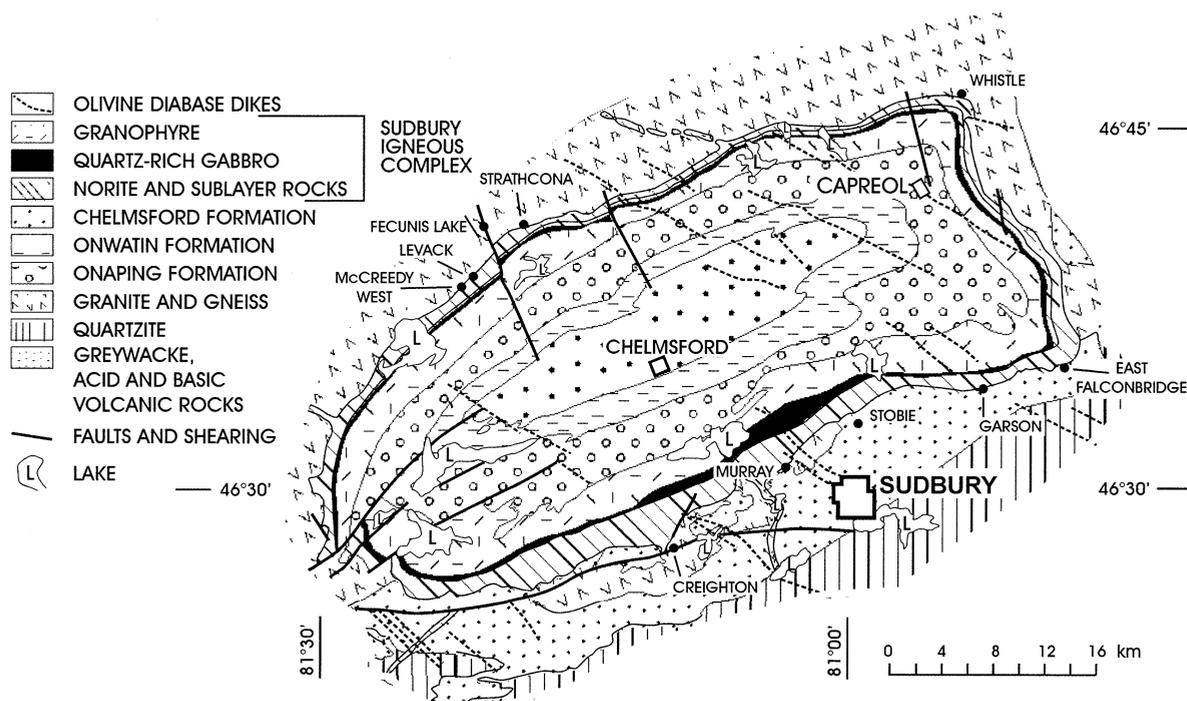
Fig. 1. Map of the Sudbury Igneous Complex (after Walker et al., 1991).

1850 Ma, with very high initial $^{187}Os/^{188}Os$. Initial ratios vary between mines; 0.60 at Creighton (Dickin et al., 1992); and 0.61, 0.64, and 1.05 at Levack West (now McCreedy West), Falconbridge, and Strathcona, respectively (Walker et al., 1991; 1994). These results suggest that the PGE in the ores, like the rocks themselves, have a substantial crustal component. More recent data tend to confirm these conclusions (Dickin et al., 1999).

Recent analytical developments that enable measurements to be made at high precision (e.g., ±0.004%) have greatly expanded the scope of Os isotopic studies. Osmium has two radiogenic isotopes, ^{186}Os and ^{187}Os (Morgan, 1999). As is well known, ^{187}Re transforms by β^- emission to ^{187}Os with a decay constant, λ, of $1.666 \times 10^{-11}A^{-1}$ (Smoliar et al., 1996). Less well known is the α transition of ^{190}Pt to ^{186}Os with $\lambda = 1.542 \times 10^{-12} y^{-1}$ (Walker et al., 1997). Because ^{190}Pt is a minor isotope of Pt (given as 0.0124 atomic %; Walker et al., 1997), long-term and large differences of Pt/Os are required to generate resolvable differences in $^{186}Os/^{188}Os$. Nevertheless, the high precision of multicollector negative thermal ionization mass spectrometry (NTIMS) has resulted in the introduction of the $^{190}Pt-^{186}Os$ isotope system as a geochronometer and particularly as a chemical tracer (Walker et al., 1997; Brandon et al., 1998; 1999; 2000). The coupled $^{187}Re-^{187}Os$ and $^{190}Pt-^{186}Os$ systems throw new light on the geochemistry of Os, other platinum group elements (PGE) and the highly siderophile elements (HSE) in general.

The Sudbury ores have highly radiogenic $^{187}Os/^{188}Os$ and are largely, or even entirely, derived from continental crustal sources. Many ores contain substantial amounts of Os and often have relatively low Pt contents. In addition, the unique positioning of the SIC may allow sampling of more than one upper continental crustal terrain. In this study, therefore, we have made precise measurements of $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$ in unspiked samples from Falconbridge, McCreedy West, and Strathcona mines, as well as determining Re, Os, and Pt abundances in spiked samples of the same ores. We have revisited the $^{187}Re-^{187}Os$ systematics at these mines, and examined the $^{190}Pt-^{186}Os$ geochronology of the Pt-rich, Os-poor Deep Copper Zone at the Strathcona mine. In addition, we have made improved measurements of the isotopic abundance of ^{190}Pt.

2. EXPERIMENT

2.1. Samples

The location of the Strathcona, McCreedy West, and Falconbridge mines are shown in Fig. 1. We reanalyzed the material originally studied by Walker et al (1991; 1994). These whole rock samples are almost entirely sulfide. The sample powders had solidified in many cases and were reground in an agate mortar.

Strathcona and McCreedy West (formerly Levack West) are situated along the North Range of the SIC in very similar regimes, lying, respectively, east and west of the Fecunis Lake fault (Coats and Snadjr, 1984). In the Strathcona deposit (Fig. 2), the Hangingwall ores (samples PMG79-138 and -139) occur as disseminated sulfides in the sublayer norites which dip at 35 to 40° S. The norites are underlain by the Footwall breccia. Main zone ores (not sampled in this work) exist as massive and disseminated sulfide within this highly brecciated gneiss. Deep Zone ores (samples PGM79-126, -128) occur as stringers of massive sulfide in fractures within the gneisses of the footwall (Coats and Snadjr, 1994; Walker et al., 1991; Li et al., 1992). Pyrrhotite is the predominant sulfide mineral in these three ore
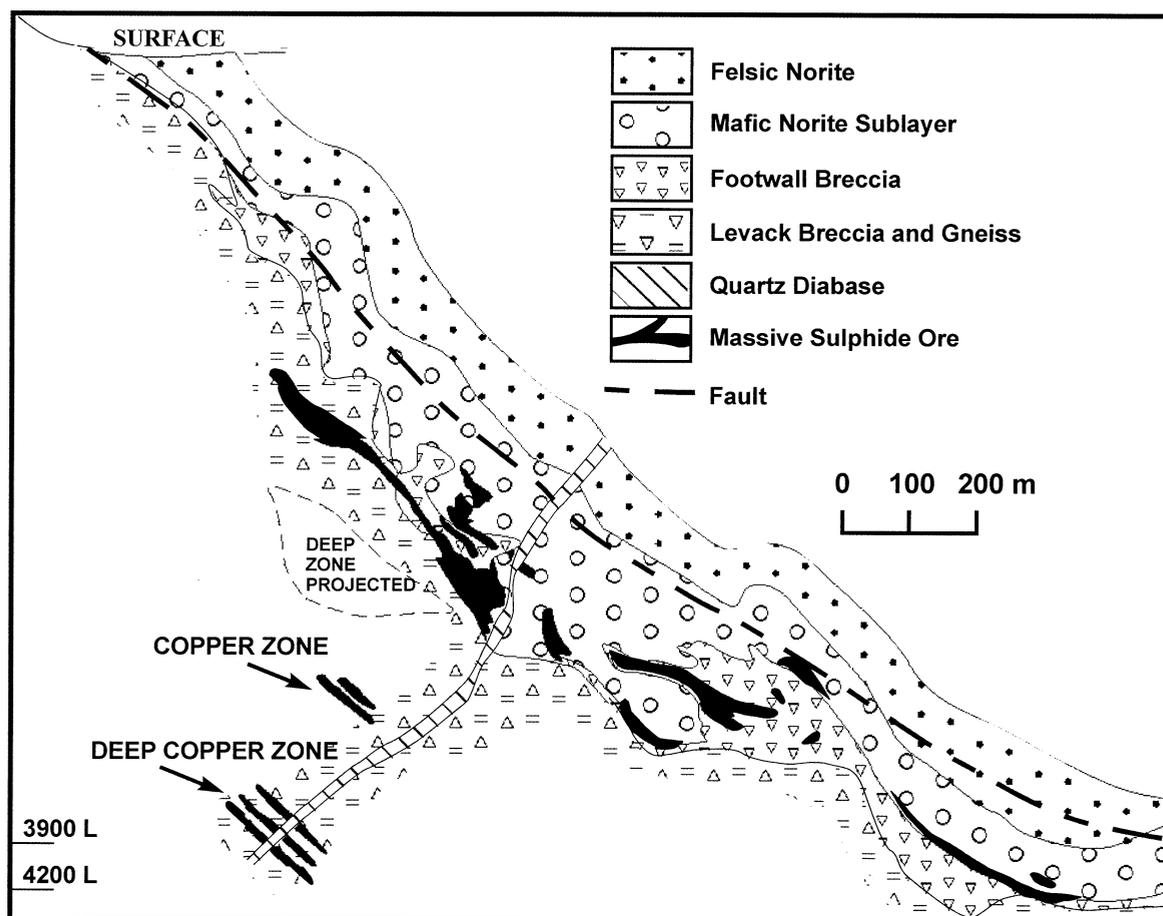
Fig. 2. Schematic vertical cross section of the Strathcona mine. Shown are the general locations of the Hangingwall Zone (PGM79-138 and -139), the Deep Zone as projected on this figure (PGM79-126, -128) and the Deep Copper Zone (DCZ-3,-10, and -19). Numbers at the left are the depth of the levels in feet. Horizontal and vertical scales are the same here, as they are in Fig. 3 (after Walker et al., 1991).

types, with pentlandite and particularly chalcopyrite proportionately increasing from hangingwall to footwall. In the footwall gneisses, two zones very rich in chalcopyrite occur; the Copper Zone ca. 200 m from the basal contact of the SIC, and, at 500 m the Deep Copper Zone (samples DCZ-3, -10, and -19).

McCreedy West is broadly similar geologically to Strathcona. Disseminated sulfides occur in the Igneous Sublayer—corresponding to the Hangingwall Zone at Strathcona—but these have not been mined. The Main ore zone occupies an embayment of the Footwall and consists of massive, breccia and disseminated sulfides in mineralized Footwall breccia (samples F41874, F41878). The Main zone ores grade into massive sulfide veins in footwall rocks that comprise the Footwall ore zone (F41872, F41873, F41875, F41877). At McCreedy West, the counterparts of the Copper and Deep Copper Zones have yet to be found, although Cu-rich stringers occur. The stringers penetrate the gneiss along fractures away from the Footwall ore zone, but were not sampled here.

The Falconbridge mine is situated at the extreme eastern tip of the southeastern side of the SIC in a region where the original structure has been compressed by faulting (Roest and Pilkington, 1994). The ore deposits occupy fault zones that form the southern contact of the SIC and the greenstones of the Stobie Formation. Sulfides form the matrix of a breccia that contains fragments of sheared norite and country rock. Disseminated sulfides also occur in the sublayer norite north of the fault, and pods of massive sulfide are present in the greenstones to the south. The Falconbridge sulfides are notably poorer in Cu than those found elsewhere at Sudbury, having Cu/Ni < 1. Faulting may have removed a major part of the deposits—perhaps more than 50%—including the Cu-rich component of the ore. Samples analyzed in this work were collected from beneath the No. 1 Flat Fault where the Main Fault begins to dip steeply towards the south (Fig. 3).

2.2. Analytical Methods

The isotopic composition of natural Pt was determined particularly to improve the abundance value for ^{190}Pt, which is known presently with poor accuracy and precision. A standard solution (1 ppm in 5% HNO_3) was made from platinum ribbon (from H. Cross Company, 99.999%). Isotopic measurements were made using the *VG Elemental Plasma 54* multiple collector inductively coupled plasma mass spectrometer (MC–ICPMS) at the Department of Terrestrial Magnetism (DTM) of the Carnegie Institution of Washington. This instrument was
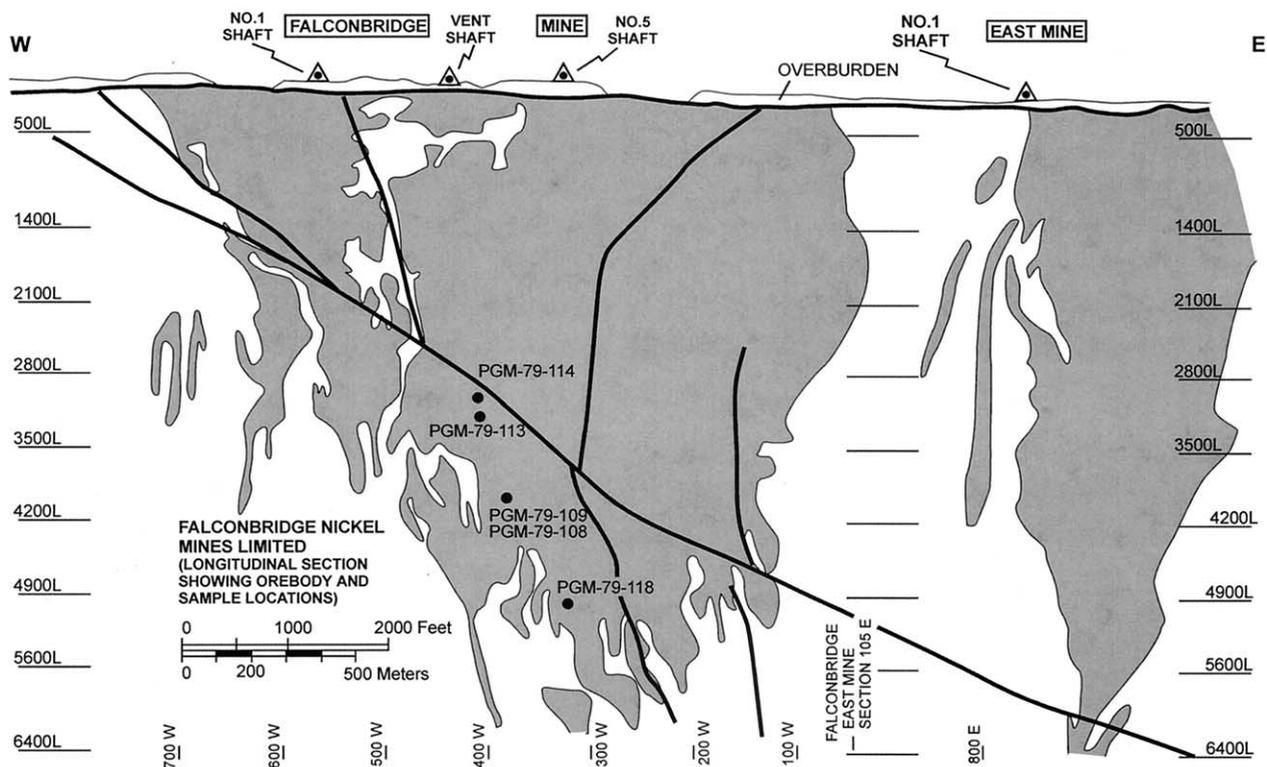
Fig. 3. Schematic vertical cross section of the Falconbridge mine. Approximate sampling locations of the analyzed ores are shown. Numbers at the left and right are the depth of the levels in feet. Numbers at bottom are the grid spacing in feet. East mine is on a different grid from the main Falconbridge mine (after Walker et al., 1991).

configured with nine Faraday collectors for the simultaneous measurement of the Pt isotopes as well as Hg and Os. The *VG Plasma 54* utilizes a Cetac MCN-6000 desolvation nebulizer; measurements were obtained in the static, multicollector Faraday cup mode. In all runs, the intensity of ^{195}Pt was greater than 5×10^{-11} A (equivalent to 5 volts), yielding nearly 2 mV of ^{190}Pt. Baselines were collected at 0.5 amu above and below each mass. Measured ratios of ^{198}Pt/^{195}Pt were corrected for isobaric mass interferences from Hg and Os and for mass fractionation using an exponential law mass fractionation equation normalized to ^{194}Pt/^{195}Pt = 0.9744 (Rosman and Taylor, 1998).

Some initial Pt determinations were made using Carius tube dissolution (Shirey and Walker, 1995). The remaining samples for isotope dilution analysis and isotopic abundance measurements were processed by Na_2O_2–NaOH fusion (Morgan and Walker, 1989; Walker et al., 1997). For isotope dilution analysis, and isotopic analysis with conventional precision, 0.45 g or less of sulfide was fused in 4 g NaOH and 4 g Na_2O_2, together with appropriate amounts of ^{185}Re, ^{190}Os, and ^{198}Pt spikes (Morgan et al., 1995; Walker et al., 1997). The fusion cake was dissolved in H_2SO_4 and Os distilled for 90 min from an H_2O_2–H_2SO_4 mixture at 105 °C. The distillate was trapped in 10 mL of chilled $6M$ NaOH. Osmium was further purified by a second distillation trapping in HBr, and prepared for mass spectrometry using microdistillation (Birck et al., 1997). To the residual liquid after distillation, 5 mL $6M$ HCl was added to complex Pt and the solution was passed through a precondi-

tioned anion exchange column containing 1.75 g dry weight of Bio-Rad AG-1 × 8, 200–400 mesh resin. The column was washed successively with 15 mL each of 5 M H_2SO_4 and 2 M H_2SO_4, 50 mL 1 M HCl and 15 mL 0.8 M HNO$_3$. Re was eluted with 40 mL $4M$ HNO$_3$ and Pt with 30 mL 0.1 M NH$_2$CSNH$_2$ (thiourea)–0.1 M HCl. Rhenium was further purified on a small 0.175 g dry weight column, loading in 0.8 M HNO$_3$ and again eluting with $4M$ HNO$_3$. Immediately after collection, the Pt fraction was evaporated to a small volume (5 to 10 mL) and, in a covered beaker, the thiourea was cautiously destroyed by addition of HNO$_3$. The resulting NH$_4$NO$_3$ was destroyed by digestion with *aqua regia*. For measurements made using the *VG PQ2* at the National Institute of Standards and Technology (NIST), the residual solution was diluted to ca. 5% HNO$_3$ for injection into the instrument. The *VG Elemental Plasma 54* multiple collector inductively coupled plasma mass spectrometer (MC–ICPMS) at the Department of Terrestrial Magnetism (DTM) of the Carnegie Institution of Washington requires more concentrated solutions than the *VG PQ2*, necessitating removal of the small amount of residual H_2SO_4 derived from thiourea. Small clean up anion exchange columns (0.09 g dry weight) were used, following the procedures of Petrie and Morgan (1982; their Fig. 3) appropriately scaled. Procedural blanks averaged: 3 ± 1 pg Os, with ^{187}Os/^{188}Os = 0.493; 48 ± 10 pg Re; and < 1 ng Pt. In most cases except for Re and Os in Deep Copper Zone samples, blank corrections were negligible.

For high precision measurements of ^{186}Os/^{188}Os, we used

unspiked samples. Initially, sample sizes were similar to those used for spiked runs, but in later analyses the fusion procedure was scaled up to use 1.2 – 1.5 g samples. Isotopic ratios for Re and Os were measured by negative thermal ionization mass spectrometry (NTIMS) using the most abundant oxides ReO_4^- and OsO_3^-. The isotopic ratios in spiked samples for Re and some for Os, were measured on a single collector 12 inch, 68° NBS mass spectrometer at the University of Maryland Isotope Geochemistry Laboratory (IGL). Generally, for spiked Os samples we measure $^{187}Os/^{188}Os$ and $^{190}Os/^{192}Os$, and make a fractionation correction setting $^{192}Os/^{188}Os = 3.0827$. In the case of the Deep Copper Zone samples (DCZ-3, -10, and -19), normalization to ^{188}Os introduced very large errors in these low-level, highly radiogenic samples and the fractionation correction introduces more error than no correction at all since ^{188}Os is poorly determined. We therefore ran a second set of Deep Copper Zone samples (new dissolutions), which were heavily spiked (ca. 40 pg ^{190}Os). The mass spectrometer software was rewritten so that the peaks of interest were normalized to the well-determined ^{190}Os peak, and the output ratios were $^{192}Os/^{190}Os$, $^{186}Os/^{190}Os$, and $^{187}Os/^{190}Os$.

Some spiked Os samples and all high precision unspiked Os isotopic measurements were made at the IGL on a *VG Sector 54* mass spectrometer equipped with seven Faraday cups. Spiked samples were run in the static mode, but the dynamic mode was used for the high precision measurement of $^{186}Os/^{188}Os$ ratios on unspiked samples. For these measurements, ideally 30 to 50 ng of Os is needed to obtain the required beam size of ca. 2 volts for mass 240 ($^{192}Os^{16}O_3^-$). Beam voltages of 80 to 120 mV on mass 234 ($^{186}Os^{16}O_3^-$) and 235 ($^{187}Os^{16}O_3^-$) are needed for 100 or more ratios to reach a desired precision of 50 to 80 ppm (2σ) for $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$. These conditions were not always met in some earlier analyses, as with the first three samples of PGM79–113, and particularly PGM79–138, where only $\approx$10 ng was separated for each analysis. In-run fractionation corrections were made for $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$ using $^{189}Os/^{188}Os = 1.219\ 78$.

Interfering ions can be present over the mass range of negative OsO_3^- ions from masses 232 to 240 ($^{184}Os^{16}O_3^-$ to $^{192}Os^{16}O_3^-$) and the most serious interference at mass 234 ($^{186}Os^{16}O_3^-$) would be $^{186}W^{16}O_3^-$. If this ion is present, however, an interference with nearly the same intensity will be manifest at mass 232 ($^{184}Os^{16}O_3^-$) from $^{184}W^{16}O_3^-$. The $^{184}Os/^{188}Os$ ratio in nature is very low—typically 0.0013—and therefore provides a sensitive monitor of tungsten trioxide interference at masses 232 and 234. No W interference was detected in these analyses. On the few occasions when mass 232 was high, other isotopes were checked and the interference was found to be PtO_2^-.

Brandon et al. (1999) has demonstrated the linearity and accuracy of the high precision measurements for $^{186}Os/^{188}Os$ by these techniques. Two series of natural Os standard solutions (designated Johnson Mathey and JWM) in use at IGL have $^{186}Os/^{188}Os = 0.119\ 848.0 \pm 0.000\ 0032$) and 0.119 9755 $\pm$ 0.000 0036, respectively, and $^{187}Os/^{188}Os = 0.113$ 8101 $\pm$ 0.000 0057 and 0.173 9176 $\pm$ 0.000 0069, respectively, where the uncertainty is 2σ of the mean. These solutions were mixed together in variable amounts, and equilibrated in Carius tubes with aqua regia (Shirey and Walker, 1995). The measured $^{186}Os/^{188}Os$ of the mixtures lie on the calculated mixing line between the two standard solutions indicating that differences of $\pm$40 ppm in $^{186}Os/^{188}Os$ can be distinguished with the techniques employed.

For isotope dilution of Pt, we measured $^{195}Pt/^{198}Pt$ ratio either on the single collector *VG PQ2* ICP–MS at NIST, or on the multicollector *VG Plasma 54* ICP–MS at DTM. For the *PQ2* the diluted H_2SO_4 solution resulting from destruction of thiourea in the Pt fraction was diluted to at least 20 mL in ca. 5% HNO_3 for injection into the instrument. The concentration of each solution was adjusted so that a count rate of 300,000 counts per second was not exceeded on the electron multiplier. The *PQ2* ICP–MS utilizes a concentric nebulizer and a spray chamber that was water cooled to 2 °C. Further details of the NIST procedures are given by Beary and Paulson (1995). Later Pt isotope dilution measurements were made using the *VG Plasma 54*. An H_2SO_4-free solution purified by the microcolumn technique (*vide supra*) were made up to 1 mL or less in dilute HNO_3. Using the Faraday cups, a 180 ppb standard solution typically gives a total ion current of 4.3×10^{-11} A for ^{198}Pt.

3. RESULTS

3.1. Isotopic Abundance of ^{190}Pt

The isotopic composition of natural Pt, newly determined, is shown in Table 1. External reproducibility is good, and except for $^{190}Pt/^{195}Pt$ is better than the uncertainty in the IUPAC value used to correct for fractionation ($^{194}Pt/^{195}Pt = 0.9744 \pm 0.0029$). The major aim was to determine precisely the isotopic abundance of ^{190}Pt, and ultimately constrain its decay constant, λ, and half-life, $t_{1/2}$. We find $^{190}Pt/^{195}Pt = 0.000\ 3821 \pm 0.000$ 0010, a value just within uncertainty of the IUPAC compilation value of 0.000 414 $\pm$ 0.000 030. Our $^{192}Pt/^{195}Pt$ and $^{198}Pt/^{195}Pt$ ratios are slightly higher than the 1997 IUPAC values (Rosman and Taylor, 1998), but atomic weights are identical. We find an isotopic abundance for ^{190}Pt of 0.012 92 $\pm$ 0.000 02, which falls between the less precise previous results; 0.0122 $\pm$ 0.0006 (RIMS, Walker et al., 1991), 0.0124 $\pm$ 0.0003 (NTIMS, Walker et al., 1997) and 0.014 $\pm$ 0.001 (IUPAC compilation, Rosman and Taylor, 1998). We will use the revised value in the following discussions, but we have not formally revised the ^{190}Pt decay constant and half-life because these are currently being redetermined in our laboratory. Note that if the new ^{190}Pt isotopic abundance were used to recalculate the half-life from existing data (Walker et al., 1997), previous Pt–Os ages and $^{186}Os/^{188}Os$ intercepts will remain the same because the changes in isotopic abundance and decay constant are mutually canceling. On isochrons, however the X-axis and slope would be different.

3.2. Re–Os Abundances and Isotope Systematics

Data are given in Table 2 for Re and Os abundances and for $^{187}Os/^{188}Os$ in sample suites from McCreedy West, Strathcona, and Falconbridge. High precision analyses for $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$ in unspiked samples with low Pt/Os ratios from these suites are shown in Table 3. Data for Pt–Os systematics are summarized in Table 4. Sulfides from Falconbridge and McCreedy West mines have similar ranges; from 131–286 ppb

Table 1. Isotopic composition of platinum. The standard (1 ppm in 5% HNO_3) was made from platinum ribbon (99.999%) obtained from H. Cross Co. Isotopic measurements were made using an VG P-54 multicollector ICP–MS. Data were obtained in static mode using nine Faraday cups, and corrected for isobaric mass interferences from Hg and Os. Mass fractionation was made using $^{194}Pt/^{195}Pt = 0.9744$ (Rosman and Taylor, 1998) and an exponential law. Uncertainties are given at the 95% confidence level.

	$^{190}Pt/^{195}Pt$	$^{192}Pt/^{195}Pt$	$^{196}Pt/^{195}Pt$	$^{198}Pt/^{195}Pt$
1	0.000 3822	0.023 7196	0.745 9988	0.214 1510
2	0.000 3819	0.023 7200	0.746 0302	0.214 1479
3	0.000 3826	0.023 7172	0.746 0547	0.214 1525
4	0.000 3818	0.023 7194	0.746 0126	0.214 1447
Average	**0.000 3821**	**0.023 7191**	**0.746 024**	**0.214 149**
Uncertainty	0.000 0010	0.000 0021	0.000 038	0.000 0040
% uncertainty	0.25%	0.0088%	0.0051%	0.0019%

Isotope	**190**	**192**	**194**	**195**	**196**	**198**
Percentages	0.012 92	0.801 68	32.933 67	33.798 92	25.214 81	7.240 45
Atomic weight	**195.0796**					

Re and 52–88 ppb Os, and 124–276 ppb Re, and 16–70 ppb, respectively. Abundances at Falconbridge are lower than at the Creighton mine, another South Range mine, where Dickin et al. (1992) found between 137 to 571 ppb Re and 111 to 350 ppb Os. At the North Range Strathcona mine, abundances for both Re and Os are variable between ore zones, decreasing with stratigraphic depth. The ranges are Hangingwall ores, 124–153 ppb Re and 22–25 ppb Os; the Deep Zone, 53–55 ppb Re, and 2.7–2.9 ppb Os; and Deep Copper Zone, 0.6–1.0 ppb Re and

0.022–0.037 ppb, of mostly radiogenic ^{186}Os and ^{187}Os. Deep Zone abundances at Strathcona are lower than those in the stratigraphically equivalent Footwall Zone ores at McCreedy West that contain 124–194 ppb Re and 16–35 ppb Os.

The improved precision of NTIMS measurements over that of previous Sudbury determinations by RIMS (Walker et al., 1991; 1994) and ICP–MS (Dickin et al., 1992) serves to emphasize the nonlinear Re–Os isotope systematics of the ores. Scatter in the data may be due to variation in the initial

Table 2. Re and Os abundances and isotopic data for Sudbury ores. Values in parentheses are the 2σ errors in the last digits of the preceding value.

Sample	Re, ppb	Os, ppb	$^{187}Os/^{188}Os$	$^{187}Re/^{188}Os$	γOs^{1850}	^{187}Re pmol/g	^{187}Os fmol/g
Falconbridge							
PGM79-108	286.1 (10)	88.58 (19)	1.1347 (1)	17.59 (6)	410 (7)		
PGM79-109	274.3 (16)	74.52 (17)	1.1758 (1)	20.14 (12)	377 (12)		
PGM79-113	260.0 (9)	87.60 (19)	1.0561 (1)	16.02 (5)	385 (6)		
PGM79-113	259.1 (9)	82.59 (19)	1.0627 (2)	16.94 (6)	365 (6)		
PGM79-114	135.6 (4)	52.31 (12)	0.9664 (1)	13.84 (5)	366 (5)		
PGM79-118	130.6 (5)	60.64 (13)	0.9090 (1)	11.43 (4)	382 (4)		
McCreedy West							
F41872	124.4 (4)	28.52 (7)	1.2533 (2)	24.08 (8)	337 (7)		
F41872	124.7 (6)	28.88 (7)	1.2521 (3)	23.85 (11)	342 (10)		
F41873	134.1 (5)	15.97 (4)	2.1533 (4)	51.12 (17)	383 (18)		
F41874	275.7 (9)	69.57 (16)	1.1998 (1)	21.75 (7)	354 (7)		
F41874	270.7 (9)	67.36 (16)	1.2018 (3)	22.06 (7)	347 (7)		
F41875	193.9 (9)	34.67 (7)	1.4849 (1)	31.70 (10)	331 (9)		
F41877	150.4 (5)	20.71 (5)	1.8425 (3)	42.80 (15)	339 (14)		
F41878	186.8 (6)	28.61 (7)	1.7002 (4)	37.88 (13)	350 (12)		
Strathcona							
PGM79-126	52.7 (2)	2.677 (7)	5.947 (11)	166.8 (1)	535 (53)		
PGM79-128	54.9 (2)	2.879 (8)	5.721 (10)	158.9 (6)	554 (99)		
PGM79-138	124.9 (4)	21.74 (5)	1.9659 (2)	34.29 (11)	680 (21)		
PGM79-138	124.1 (5)	21.61 (5)	1.9614 (7)	34.27 (14)	677 (21)		
PGM79-139	151.6 (5)	24.68 (6)	2.0468 (2)	36.98 (12)	677 (23)		
PGM79-139	153.3 (5)	25.15 (6)	2.0402 (4)	36.66 (12)	680 (23)		
Strathcona Deep Copper Zone							
DCZ3A	1.02 (2)	0.037 (5)	11.2 (1.1)	329 (45)		34.4 (4)	108 (2)
DCZ3B	0.90 (3)	0.033 (6)	12.3 (1.6)	346 (64)		24.0 (3)	67 (2)
DCZ10A	0.72 (2)	0.023 (5)	11.8 (1.9)	391 (89)		20.4 (3)	64 (2)
DCZ10B	0.63 (2)	0.022 (7)	14.7 (3.1)	426 (130)		30.4 (3)	100 (2)
DCZ19A	0.61 (2)	0.033 (4)	6.5 (0.7)	175 (22)		21.2 (3)	68 (2)
DCZ19B	0.64 (2)	0.029 (5)	9.1 (1.2)	255 (49)		21.7 (3)	70 (2)

Table 3. High precision measurements of Os isotopic ratios in Sudbury ores.

	# Ratios	$^{186}Os/^{188}Os$ Dynamic	$^{187}Os/^{188}Os$ Dynamic	$^{184}Os/^{188}Os$ Static
Falconbridge				
PGM79-113	100	0.119 858 (6)	1.057 88 (2)	0.001 340 (38)
PGM79-113	60	0.119 848 (15)	1.059 63 (5)	0.001 341 (9)
PGM79-113	150	0.119 854 (8)	1.054 57 (3)	0.001 391 (18)
PGM79-113	100	0.119 866 (8)	1.058 99 (2)	0.001 319 (5)
PGM79-113	100	0.119 859 (6)	1.058 85 (2)	0.001 323 (4)
PGM79-113	100	0.119 844 (8)	1.055 36 (2)	0.001 320 (4)
Weighted average ($n = 6$)		0.119 855 (8)		
external population error		1.40E-04		
MSWD = 5.12				
McCreedy West				
F41874	100	0.119 856 (8)	1.203 11 (3)	0.001 315 (6)
F41874	100	0.119 872 (5)	1.201 22 (2)	0.001 316 (3)
F41874	100	0.119 871 (7)	1.200 21 (3)	0.001 313 (5)
Weighted average ($n = 3$)		0.119 867 (20)		
external population error		1.60E-04		
MSWD = 5.93				
Average w/o external error		0.119 869 (4)		
Strathcona				
PGM79-138	100	0.119 825 (12)	1.961 59 (5)	0.001 340 (38)
PGM79-138	30	0.119 825 (21)	1.961 71 (14)	0.001 341 (9)
PGM79-138	60	0.119 830 (20)	1.958 69 (15)	0.001 391 (18)
Weighted average ($n = 3$)		0.119 826 (9)		
MSWD = 0.096				
PGM79-139	100	0.119 828 (6)	2.047 51 (4)	0.001 339 (3)
PGM79-139	100	0.119 824 (7)	2.046 41 (4)	0.001 344 (5)
PGM79-139	100	0.119 830 (8)	2.045 54 (5)	0.001 334 (6)
Weighted average ($n = 3$)		0.119 827 (4)		
MSWD = 0.32				
Total weighted average ($n = 6$)		0.119 827 (3)		
MSWD = 0.67				

Table 4. Pt–Os isotope data. CT= Carius tube; AF=alkali fusion; NIST = run at National Institute of Standards and Technology; DTM = run at Department of Terrestrial Magnetism, Carnegie Institution of Washington. *Italicized* $^{190}Pt/^{188}Os$ have Pt analyzed on a separate aliquot. **Bold** $^{186}Os/^{188}Os$ run on a separate aliquot in the dynamic mode.

| Sample	Pt, ppb CT–NIST	Pt, ppb AF–NIST	Pt, ppb AF–DTM	Pt, ppb Lit.	$^{190}Pt/^{188}Os$	$^{186}Os/^{188}Os$	$^{186}Os/^{188}Os_{init}$	^{190}Pt pmol/g	^{186}Os fmol/g
Falconbridge									
PGM79-113		81		235					
replicate			100		0.001 29 (1)	**0.119 855 (8)**	0.119 851 (8)		
McCreedy West									
F4 18574			442	610	0.007 11 (7)	**0.119 867 (20)**	0.119 848 (20)		
Strathcona									
PGM79-126	354			429	*0.221 (2)*				
PGM79-128	517		515						
replicate		636			0.363 (4)				
PGM79-138			84	55	0.00460 (5)	**0.119 826 (9)**	0.119 813 (9)		
PGM79-139	33			41	*0.00158 (2)*	**0.119 827 (4)**	0.119 823 (4)		
Deep Copper Zone									
DCZ 03	107			744					
replicate		134			8.5 (1.1)	0.189 (30)	0.166 (30)	0.089 (1)	0.27 (38)
replicate			97		7.4 (1.4)	0.177 (40)	0.157 (40)	0.065 (1)	0.12 (39)
DCZ 10	3430			5600					
replicate		2746			297 (67)	1.17 (26)		1.82 (2)	6.2 (3)
replicate			4458		595 (181)	1.61 (49)		2.96 (3)	7.2 (3)
DCZ 19	7281			10,500					
replicate	7208								
replicate		6566			374 (45)	1.43 (17)		4.36 (4)	14.7 (4)
replicate			6595		516 (97)	1.60 (4)		4.38 (4)	12.2 (4)
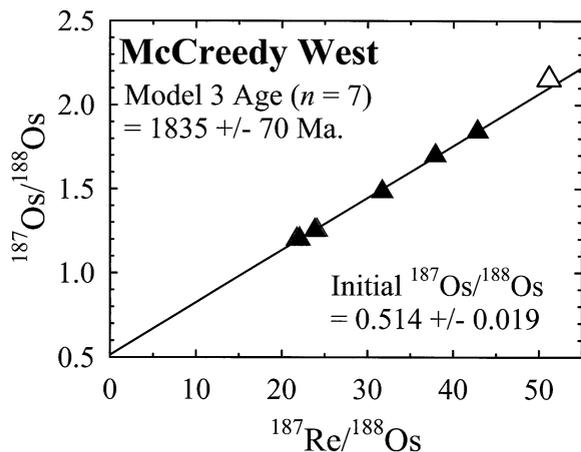
Fig. 4. $^{187}Os/^{188}Os$ vs. $^{187}Re/^{188}Os$ isochron plot for McCreedy West ores. All data ($n=8$) yield a Model 3 solution of 1904 ± 81 Ma with an intercept of 0.483 ± 0.033 (2σ) and MSWD of 79. Excluding the furthest outlier (F41873; open symbol) gives a model 3 age of 1835 ± 70 Ma, an intercept of 0.514 ± 0.019 (2σ) and MSWD of 45.

Fig. 5. $^{187}Os/^{188}Os$ vs. $^{187}Re/^{188}Os$ isochron for Falconbridge ores. All data ($n = 6$) yield a Model 3 age of 1911 ± 500 Ma, with an intercept of 0.53 ± 0.042 and MSWD = 488. Exclusion of the most distant outlier (PGM79-108; open symbol) gives a Model 3 age of 1827 ± 340 Ma, with an intercept of 0.550 ± 0.024 and MSWD = 228.

$^{187}Os/^{188}Os$ or later disturbance (Dickin et al., 1992). Isochron regressions were fitted using Isoplot/Ex 2.00 (Ludwig, 1999). McCreedy West data yields a Model 3 isochron ($n = 8$) of 1904 ± 81 Ma with an intercept of 0.483 ± 0.033 and MSWD of 79. Exclusion of the furthest outlier (F41873 from Footwall Zone) gives a model 3 age ($n = 7$) of 1835 ± 70 Ma, with an intercept of 0.514 ± 0.019 and an improved MSWD of 45 (Fig. 4). The age agrees within uncertainty with the accepted age of the SIC of 1850 ± 1 Ma (Krogh et al., 1984; 1996). The weighted average of γ_{Os}^{1850} for all McCreedy West data is 346 ± 10 ($n = 8$, MSWD = 74) or, without F41873, 344 ± 8 ($n = 7$, MSWD = 42).

At Falconbridge, the small range in $^{187}Re/^{188}Os$ and scatter about the regression line limit the usefulness of the geochronological data. Using all data yields a Model 3 age of 1911 ± 500 Ma, with an intercept of 0.53 ± 0.04 and MSWD = 488. Exclusion of the most distant outlier (PGM79-108) gives a slightly better Model 3 age of 1827 ± 340 Ma, an intercept of 0.550 ± 0.024 and MSWD = 228 (Fig. 5). The weighted average γ_{Os}^{1850} of all Falconbridge data is 380 ± 17 ($n = 6$, MSWD = 473) and, excluding PGM79-108, 375 ± 12 ($n = 5$, MSWD = 177).

Data from McCreedy West and Falconbridge include three sets of duplicate analyses that agree internally, but do not lie exactly along the ca. 1850 Ma isochron. In each case the trend suggests a much younger age; PGM79-113, 421 ± 36 Ma; F41872, 303 ± 200 Ma; and F41874, 399 ± 140 Ma. Each pair could give this age randomly, but the weighted average of 416 ± 34 Ma (95% confidence) may relate to a real event. Two diabase dykes from near Killarney southwest of Sudbury give K–Ar whole rock ages of 431 ± 80 and 416 ± 76 Ma. Consequently, the "noise" in these isochrons seems to reflect minor open system behavior, and not variations in initial Os isotopic composition.

The geochronology of the Strathcona mine differs from that at nearby McCreedy West and more distant Falconbridge, exhibiting a Model 1 age of 1780 ± 7 Ma, with a initial $^{187}Os/^{188}Os$ = 0.9336 ± 0.0047 ($n = 6$) (Fig. 6). Unlike

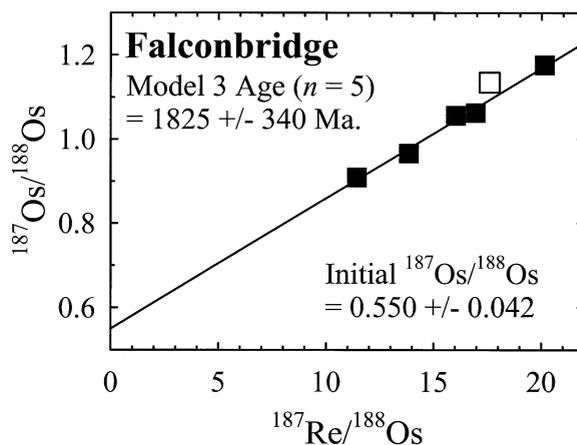replicates at McCreedy West and Falconbridge, those at Strathcona reproduce tightly, and all lie within uncertainty of the 1780 Ma regression as indicated by MSWD = 1.6. The γ_{Os}^{1780} values give a precise weighted mean of 713 ± 2 ($n = 6$, MSWD = 1.3). The Strathcona isochron seems a virtual two-point join between the closely grouped Hangingwall samples and the two Deep Zone samples. Within each group, however, the internal systematics is within uncertainty of 1780 Ma age, though neither categorically excludes an 1850 Ma age. The four Hangingwall samples give a Model 1 age of 1845 ± 86 Ma (MSWD = 1.7) whereas two Deep Zone samples suggest a younger Model 1 age of 1686 ± 210 Ma. [Recent unpublished Carius tube analyses of Deep Zone ore PGM79-126 give a similar age of 1735 ± 230 Ma ($n = 4$), MSWD = 1.6 (Anders Schersten, private communication).]

The 1780 Ma isochron for all of our Strathcona data plots within uncertainty of the two Deep Zone data, but an 1850 line
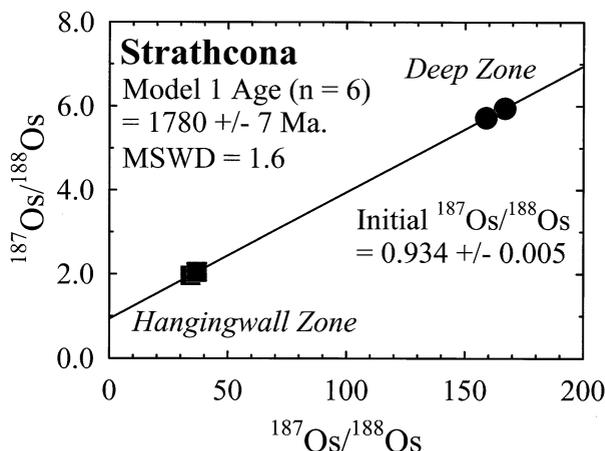

Fig. 6. $^{187}Os/^{188}Os$ vs. $^{187}Re/^{188}Os$ isochron plot for Strathcona Hangingwall Zone and Deep Zone ores. All data ($n = 6$) give a well constrained Model 1 age of 1780.3 ± 6.7 Ma, with a precisely defined initial $^{187}Os/^{188}Os$ = 0.9336 ± 0.0047 (MSWD = 1.6).

would only barely do so. Often, isochrons with apparently discrepant ages are mixing lines between end members with different initial ratios. Such correlations particularly are suspect when the elemental abundances in the samples show systematic variations, as is the case here (Chauvel et al., 1985). Although we cannot categorically exclude mixing, at Strathcona the consistent trends in abundance are those expected for a crystallizing sulfide liquid (Naldrett et al., 1982; Li et al., 1992). The Hangingwall sulfides are derived almost entirely from cumulate monosulfide solid solution (mss), whereas the Deep Zone ores represent a moderately fractionated liquid that has crystallized *in situ* with little loss of more fractionated melt (Naldrett et al., 1999). Since both types of ore are derived from the same parental liquid, they represent unmixing rather than mixing and the differences in abundance largely reflect the mss/liquid partition coefficients. We conclude that the precise regression of the Strathcona isotopic data is unlikely to be a mixing line but represents resetting of the Os isotope systematics by the evident dyke activity at Strathcona (Fig. 2) possibly related to post-Penokean lower crustal melting (Holm et al., 1998).

The Deep Copper Zone (DCZ) ores represent the most fractionated Strathcona sulfides, and many are characterized by high Cu, Pt, and Pd, and low Ni, Re, Os, and Ir (Walker et al., 1991; Naldrett et al., 1999). In highly fractionated sulfide liquids ($\approx$90% crystallized) Re and Os abundances are very low, but as Cu reaches ca. 32%, the fractionation trend abruptly reverses. A Cu-rich intermediate solid solution (iss) is then the cumulate phase, in which Cu is compatible and such elements as Re and Os incompatible. The Deep Copper Zone ores lie along this very late liquid fractionation trend (Naldrett et al., 1999). In these ores, Re and particularly Os are strongly depleted, so the DCZ samples are highly enriched in radiogenic ^{187}Os. The conventional measurement of Os ratios relative to ^{188}Os in highly radiogenic, low Os samples is imprecise because of the small ^{188}Os peak, which may be largely blank (Morgan, 1999; Stein et al., 2000)

For the three DCZ samples, we ran two sets of mass spectrometric analyses. These were different dissolutions, because, for our ca. 0.4 g sample size, all the Os was consumed in a single analysis. For the first set (A), we used conventional software that measures ratios relative to ^{188}Os, with a modification to include ^{186}Os/^{188}Os. The second set (B) was heavily spiked with 40 pg of ^{190}Os and ratios were measured relative to this dominant peak. Thus the absolute abundance of ^{187}Os may be measured directly from ^{187}Os and ^{190}Os, which are the most prominent peaks in the mass spectrum. For comparison, results were calculated in two ways for both A and B sets; either as ratios as is conventional, and set B data were recalculated to ^{188}Os ratios from ^{192}Os, the most abundant nonradiogenic isotope (Table 2). The data for set A taken alone are scattered and do not give a reasonable isochron. Set B yields a Model 1 age of 1931 $\pm$ 210 Ma (MSWD = 0.92), with a poorly determined initial ^{187}Os/^{188}Os of 0.9 $\pm$ 1.2. This age is clearly within error of both the canonical age of 1850 Ma for the SIC and also the apparent age of 1780 Ma of the other Strathcona ores. Two Set A data (DCZ 03 and DCZ 19) fall close to the set B isochron, and give a slightly worse Model 1 age of 1894 $\pm$ 230 Ma (MSWD = 2.3) with an identical initial ^{187}Os/^{188}Os of 0.9 $\pm$ 1.1. These five data points for the DCZ lie systematically
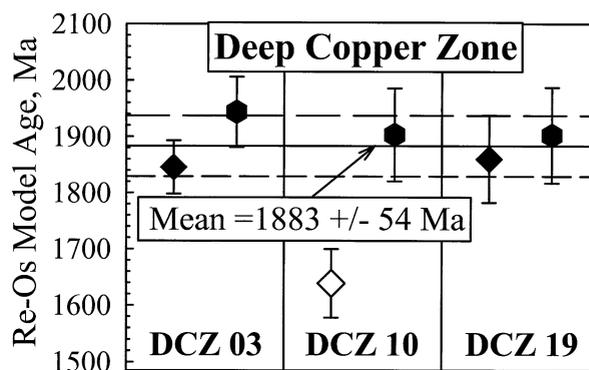

Fig. 7. Model ^{187}Re–^{187}Os age for Strathcona Deep Copper Zone ores based on the measured Strathcona initial ^{187}Os/^{188}Os of 0.9336 $\pm$ 0.0047. Symbols: diamonds, ^{188}Os normalization; hexagons ^{190}Os normalization. Five of the six samples fall within a narrow age range and yield a mean model age of 1883 $\pm$ 54 Ma (95% confidence) with MSWD of 1.8. The remaining DCZ-10 sample (open diamond) has a model age of 1640 $\pm$ 60 Ma and is clearly an outlier.

above those for other ores from the Hangingwall and Deep Zone at Strathcona. (No figure is plotted because the DCZ samples have large and highly correlated uncertainties that make a plot of ^{187}Os/^{188}Os vs ^{187}Re/^{188}Os visually incomprehensible.)

Apparently, the Deep Copper Zone escaped the resetting represented by the 1780 Ma isochron of the Hangingwall and Deep Zone ores. The Deep Copper Zone and other Strathcona ores appear to have a similar initial ^{187}Os/^{188}Os and that would change little in the 70 Ma between 1780 and 1850 Ma. If we use the measured Strathcona initial ^{187}Os/^{188}Os of 0.9336 $\pm$ 0.0047 from Fig. 6, we may calculate model ages for the DCZ samples (Fig. 7). Five of the six samples fall within an age range between 1846 $\pm$ 47 and 1944 $\pm$ 63 and yield a mean model age of 1883 $\pm$ 54 Ma (95% confidence, MSWD = 1.8). The remaining DCZ 10 sample has a model age of 1639 $\pm$ 61 Ma and is clearly an outlier (Fig. 7). The mean model age of five Deep Copper Zone samples agrees well with the canonical SIC age of 1850 Ma and clearly differs from the age of 1780 $\pm$ 7 Ma of other Strathcona samples.

3.3. Pt–Os Isotope Systematics

We made precise Os isotope measurements in the dynamic mode of unspiked samples from Falconbridge, McCreedy West, and Strathcona known to have low Pt/Os (Walker et al., 1991). At least three, and in one case six (Falconbridge PGM79-113), replicate measurements were made on separate dissolutions for each sample (Table 3). At Strathcona, where the ^{187}Re/^{188}Os–^{187}Os/^{188}Os correlation is high (MSWD = 1.6), the six ^{186}Os/^{188}Os analyses of two different samples agree extremely closely (MSWD = 0.30). At McCreedy West and Falconbridge, however, where the Re and Os isotopic data show considerable scatter, we find that replicates also show variations in ^{186}Os/^{188}Os that are large compared to measurement uncertainty (MSWD = 5.9 and 5.1, respectively). In addition, Strathcona samples are nonradiogenic, but those from McCreedy West and Falconbridge show significant ^{186}Os en-
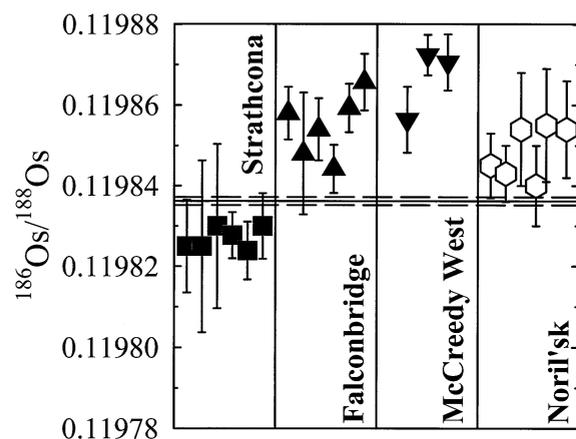
Fig. 8. ^{186}Os/^{188}Os variation in Sudbury ores from Strathcona (PGM79-138 and -139, solid squares) Falconbridge (PGM79-113, solid upright triangles) and McCreedy West (F418574, solid inverted triangles). For comparison is shown data for a plume-derived Noril'sk–Talnakh, Russia, ore (KZ 1713, 931.3, open hexagons). The present day mantle composition is represented by the mean (solid horizontal line) and 2σ errors (dashed lines) for seven abyssal peridotites (Brandon et al., 2000).

Fig. 9. ^{186}Os/^{188}Os vs. ^{190}Pt/^{188}Os isochron for Strathcona Deep Zone ores. In the trio of DCZ samples measured using ratios to ^{188}Os, ^{186}Os/^{188}Os vs. ^{190}Pt/^{188}Os are well correlated (MSWD = 0.040) and yield an isochron with a Model 1 age = 2295 ± 68 Ma and initial ^{186}Os/^{188}Os = 0.161 ± 0.027. The high age most likely is not a real value and may be an artifact of the on-line fractionation and spike correction when little ^{188}Os is present. The intercept is probably a good estimate, however.

richment. These differences cannot be attributed to *in situ* growth in 1850 Ma (Table 3).

Strathcona ore PGM79-138 was analyzed for ^{186}Os/^{188}Os using three aliquots weighing between 0.4 and 0.5 g, each containing ca. 20 ng total Os, an amount less than the optimum (25 to 50 ng) for highest precision. The replicates nevertheless agree well, averaging 0.119 8260 ± 0.000 0089, an uncertainty of ±74 ppm with MSWD = 0.10. For the second Strathcona sample, PGM79-139, three larger aliquots (ca. 50 ng total Os) gave a more precise average of 0.119 8271 ± 0.000 0039, (±32 ppm), MSWD = 0.64. Pooling all six data gives 0.119 8269 ± 0.000 0035 (30 ppm), MSWD = 0.30 (Fig. 8).

A single Falconbridge ore (PGM79-113) was analyzed six times, in two triplicate batches. In the first set, samples each contained ca. 35 ng total Os and replicates agree within uncertainty with average ^{186}Os/^{188}Os = 0.119 8555 ± 0.000 0046 (±38 ppm), MSWD = 0.87. Samples in the second set contained ca. 140 ng Os, but not all was used since loadings of >50 ng confer no advantage in signal voltage, and in fact may make matters worse. Although these samples individually are within uncertainty of the first set, the second trio has scatter greater than the measurement error and give a weighted average of 0.119 858 ± 0.000 026 (±0.022%) with MSWD = 11.3 indicating a large external error. Pooled data for all six Falconbridge samples average 0.119 8568 ± 0.000 0082 (±68 ppm), MSWD = 4.9 (Fig. 8). A similar result was obtained analyzing an ore from McCreedy West (F41874). Triplicate analyses of samples each containing ca. 100 ng Os yield an average ^{186}Os/^{188}Os = 0.119 869 ± 0.000 019 (±160 ppm) with an MSWD of 5.8 indicating moderate external error. As can be seen from Table 3 and Fig. 8, two replicates agree very well —0.119 872 ± 0.000 005 and 0.119 871 ± 0.000 007, respectively— but the third analysis is significantly lower at 0.119 856 ± 0.000 008. The variability at Falconbridge and McCreedy West is unlikely to be analytical. There is no interference apparent in the ^{184}Os/^{188}Os ratios in these samples, which were run during

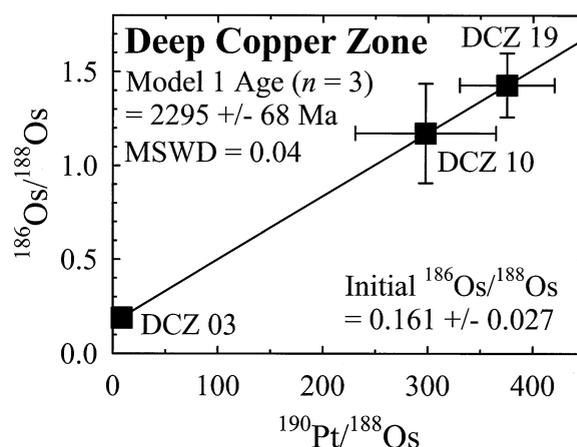the same time period and under the same conditions as the consistent data for Strathcona. The mass spectrometry has been demonstrated to be highly reproducible. For example, 10 replicates of a solution of osmiridium give a weighted mean of 0.119 8342 ± 0.000 0014 (±12 ppm), MSWD = 1.05 (sample W5, Walker et al., 1991, mean recalculated using Isoplot/Ex version 2.00, Ludwig, 1999). Similarly, Brandon et al. (2000) report for seven abyssal peridotites a weighted mean of 0.119 8362 ± 0.000 0010 (±8 ppm) with MSWD = 0.55 (Fig. 8). In the Deep Copper Zone, high ^{190}Pt/^{188}Os may occur (up to ≈600), and it was in one of these ores (DCZ 19) that substantial enrichment of the ^{186}Os/^{188}Os ratio in nature was first reported (Walker et al., 1991). Because of the very low Os content of these ores, the isotopic composition was measured on the 12 in. NBS machine using an electron multiplier. Accordingly, the precision of ^{186}Os/^{188}Os measurements was orders of magnitude worse than in the determinations discussed above. In the trio of DCZ samples measured using ratios to ^{188}Os, ^{186}Os/^{188}Os vs. ^{190}Pt/^{188}Os are well correlated (MSWD = 0.040) and yield an isochron with a Model 1 age = 2295 ± 68 Ma and initial ^{186}Os/^{188}Os = 0.161 ± 0.027 (Fig. 9). The high age most likely is not a real value and may be an artifact of the on-line fractionation and spike correction with so little ^{188}Os present. Nevertheless, the intercept is probably a good estimate. The calculated initial ^{186}Os/^{188}Os values at 1850 Ma for two relatively nonradiogenic samples of DCZ 03 are 0.166 ± 0.030 and 0.157 ± 0.040, and have a weighted mean of 0.163 ± 0.024 that is in excellent agreement with the isochron value. At present it is not clear whether the apparently radiogenic initial ^{186}Os/^{188}Os in the Deep Copper Zone is real or has a bias exacerbated by the very low signal and highly radiogenic ^{187}Os/^{188}Os composition. Taken at face value, the Deep Copper Zone initial of 0.163 ± 0.024 is significantly higher than those of other Strathcona samples (0.119 82), or indeed those from McCreedy West and Falconbridge (0.119 85). The implication is that the small amount of Os in the Deep Copper Zone
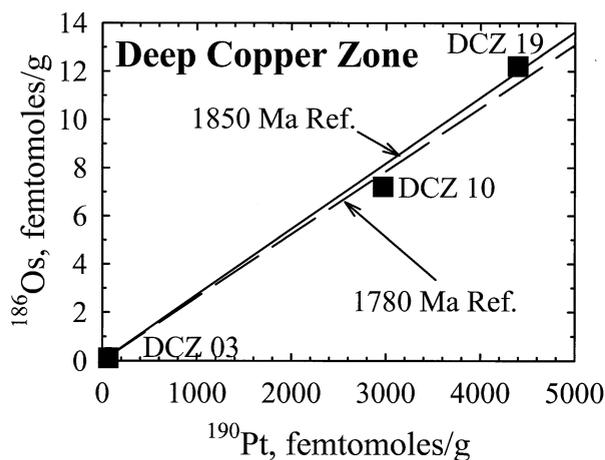
Fig. 10. Plot of ^{186}Os vs. ^{190}Pt for the second trio of DCZ samples that were measured directly against ^{190}Os spike. These do not give an isochron, but lie almost within error of an 1850 Ma or a 1780 Ma reference line. The DCZ-19 sample has ^{186}Os/^{188}Os = 1.60 ± 0.04, which yields a model ^{190}Pt–^{186}Os age of 1886 ± 62 Ma.

ores may have evolved for a significant time in a regime with high ^{190}Pt/^{188}Os. If we take, for example, the age of the enclosing Levack Gneiss complex of 2711 ± 7 Ma (Krogh et al., 1984), the initial ^{186}Os/^{188}Os at 1850 Ma in the Deep Copper Zone would imply development for the intervening 860 Ma from ^{190}Pt/^{188}Os = 34 ± 19. This value is about 0.1 of the average for the Deep Copper Zone. The estimated initial ^{186}Os/^{188}Os at 1850 Ma of 0.163 ± 0.024 may be used to derive an absolute amount of radiogenic ^{186}Os from the DCZ measurements. Data from the first trio of DCZ samples (measured using ratios to ^{188}Os) again gives a good linear fit in a ^{186}Os vs. ^{190}Pt plot and of course yields a similar old age; Model 1 age = 2292 ± 89 Ma, with an intercept of −10 ± 330 attomoles ^{186}Os and MSWD = 0.045. The second set (measured directly against ^{190}Os spike) does not give a good isochron, but the data lie almost within uncertainty of an 1850 Ma reference line (Fig. 10). The ore sample DCZ 19 in this set has the most radiogenic value yet reported of ^{186}Os/^{188}Os; 1.60 ± 0.04. This value yields a model ^{190}Pt–^{186}Os age of 1884 ± 62 Ma in agreement with the ^{187}Re–^{187}Os isochron age for the Deep Copper Zone.

4. DISCUSSION

4.1. Origin of Sudbury Ores and the Development of the Os Isotopic Signature

There is general agreement that the Sudbury structure is of asteroidal impact origin (Deutsch and Grieve, 1994; Grieve, 1994; Golightly, 1994) and that geophysical evidence no longer mandates a large mafic body beneath the SIC (McGrath and Broome, 1994) even though a mafic contribution to the melt rock may be required (Naldrett et al., 1999). Instead of mixing between crustal and mantle-derived magmas, the differences we find here in initial isotopic Os signature between Strathcona on the one hand, and McCreedy West and Falconbridge on the other, may be due to compositional variations in the crustal target rocks. [Dickin et al. (1992; 1999) reached rather similar conclusions based on ^{187}Os/^{188}Os only.] A more speculative

interpretation suggesting a possible contribution from the impacting body will be discussed in a later section.

We may summarize briefly, as follows, the extensive previous work on Sr, Nd, and Pb isotope systems (Rao et al., 1984; Deutsch, 1994; Faggart et al., 1985; Dickin et al., 1996; 1999; Dickin and Crocket, 1998). These isotope systems indicate aspects of the geochemistry of the Sudbury ores quite different from those implied by the Os isotopic results. The Nd isotopic data indicate a relatively homogeneous crustal origin for the SIC, and reveal no significant differences between individual mines, or systematically between North and South Ranges (Faggart et al., 1985; Dickin and Crocket, 1998; Dickin et al., 1996; 1999). Model Nd ages relative to depleted mantle of both North and South Range material of ca. 2700 to 2800 Ma point to ultimate derivation chiefly from Superior Province rocks. This result does not exclude the contention that much of the target for the Sudbury impact was from the Huronian sedimentary apron, since these sediments were ultimately derived from Superior Province parental material. The Sr and Pb data are more sensitive than those of Nd to variations in target material, and breccias and the ore-bearing sublayer show distinctive North and South Range Sr and Pb isotopic signatures. The Sr isotopic composition shows a marked difference between Sudbury breccias and sublayer of the North and South Ranges, respectively. Conversely, the norites, gabbros, and granophyres of the main mass have intermediate values for initial ^{87}Sr/^{86}Sr and are identical at North and South Ranges, indicating that the melt rocks were well homogenized for this isotope system (Rao et al., 1984; Deutsch, 1994; Dickin et al., 1999). The Pb isotopic data reported by Dickin et al. (1996; 1999) also show systematic differences between mines at the North and South Ranges. Initial ^{207}Pb/^{204}Pb ratios show separate groupings of all North Range breccias and ores (including Strathcona) on the one hand, and, at slightly more radiogenic values, all South Range breccias and ores on the other. The silicate melt rocks are clearly progressively homogenized. The South Range norite and, to a lesser degree, the corresponding gabbro retain some of the radiogenic character of the South Range breccia. The South Range granophyre, however, and the North Range rocks appear homogeneous in their Pb isotope signature (Dickin et al., 1999). In neither isotope system does Strathcona differ from other North Range mines. The melt rock data for Sr and Pb indicate that in the enormously chaotic regime of a giant impact, mixing of even totally molten material is well advanced but not complete.

In the Sudbury ores, Os isotopes appear to fall into two distinct groups. At McCreedy West (North Range) and at Falconbridge (South Range), initial ^{187}Os/^{188}Os ratios are 0.512 ± 0.014 and 0.550 ± 0.016, respectively, when all samples are included. Earlier measurements of ores from the Creighton mine in the South Range have initial ratios averaging 0.599 ± 0.004 (Dickin et al., 1992; for only the "undisturbed" samples, recalculated using ^{187}Re λ = 1.666 × 10^{-11} A^{-1} from Smoliar et al., 1996). The initial ^{187}Os/^{188}Os signature in each of these three mines is similar, as pointed out previously by Walker et al., (1991; 1994) and more recently by Dickin et al. (1999) but may not be identical (Dickin et al., 1992). The new data suggest that there is a small, but significant, difference between McCreedy West mine in the North Range and each of the two South Range mines, Falconbridge and Creighton. Iso-

topic Os in the Strathcona mine is very different from the other three mines, but the interpretation is complicated by the Strathcona isochron that yields an apparent age of 1780 ± 7 Ma with an initial ^{187}Os/^{188}Os of 0.934 ± 0.005. Dickin et al. (1992) consider that the Sudbury impact melt was sufficiently homogenized for all ores to have the same ^{187}Os/^{188}Os at 1850 Ma, although the later Pb data suggests that homogenization was not complete (Dickin et al., 1996; 1999). Dickin et al. (1996) "attribute the radiogenic initial ratios of the main sublayer mineralization at Strathcona to postmagmatic disturbance", and claim that the wide range of initial ratios at Strathcona overlaps those of other mines. This is clearly not the case with the new Strathcona data. If we calculate initial ratios at 1850 Ma we obtain two groups; for the Hangingwall ores, 0.891 ± 0.002 and for the Deep Zone, 0.739 ± 0.015. Both values are significantly higher than those of the other three mines, and since the Deep Copper Zone appears even more radiogenic, post-magmatic disturbance is unlikely to produce such radiogenic initial ratios. The parsimonious explanation of the Strathcona data is that the system was completely reset at 1780 Ma. To produce in the intervening 70 Ma the observed initial ^{187}Os/^{188}Os ratio of 0.933 from a value of 0.512 similar to that at nearby McCreedy West, would require an *average* ^{187}Re/^{188}Os of 351. Although such high values are seen at Strathcona in the Deep Copper Zone, these are accompanied by extremely low Re and Os contents and are an improbable source for the amounts of radiogenic ^{187}Os required.

We may use a more compelling argument based on ^{190}Pt/^{186}Os systematics. Hangingwall ores at Strathcona have very low ^{190}Pt/^{188}Os and ^{186}Os/^{188}Os ratios. Their present day ^{186}Os/^{188}Os ratio of 0.119 8269 ± 0.000 0035 is lower than that of the present day mantle and is about the value for a chondritic mantle at 1850 Ma. McCreedy West and Falconbridge have significantly higher present day values and, at 1850 Ma, had ^{186}Os/^{188}Os values of 0.119 848 and 0.119 851, respectively. Therefore, the ^{186}Os isotopic signature at Strathcona could not be produced in any plausible way from one resembling that at McCreedy West or at Falconbridge. It may be possible to derive the isotopic signatures of these mines from the components present at Strathcona, however.

In Pb isotope systematics, ores and breccias plot between the data for Superior Province and Huronian rocks. For the Re–Os system, data for these country rocks is limited to two rocks analyzed by Walker et al. (1991). Sample S-90-1 is unaltered tonalitic Levack gneiss from the late Archean country rock adjacent to the North Range of the SIC. Values recalculated for an improved NTIMS Re spike recalibration (Walker et al., 1994) and renormalized to ^{188}Os rather than ^{186}Os as in the original RIMS work, are as follows; 0.118 ± 0.004 ppb Re, 0.0073 ± 0.0002 ppb Os, ^{187}Re/^{188}Os = 78 ± 3, and ^{187}Os/^{188}Os = 3.75 ± 0.08. At 1850 Ma, this rock would have an initial ^{187}Os/^{188}Os = 1.32 ± 0.13 and a model age of ca. 2750 Ma for derivation from a mantle-like reservoir (Shirey and Walker, 1998). Sample SM-75-110 is a Huronian age (ca. 2400 Ma) glacially derived metasedimentary rock, probably of Superior Province provenance. The rock was collected from the Gowganda Formation approximately 200 km northeast of the SIC and contains 0.448 ± 0.009 ppb Re, 0.108 ± 0.001 ppb Os, ^{187}Re/^{188}Os = 20.1 ± 0.5, and ^{187}Os/^{188}Os = 1.044 ± 0.013. The corresponding initial ^{187}Os/^{188}Os at 1850 Ma is
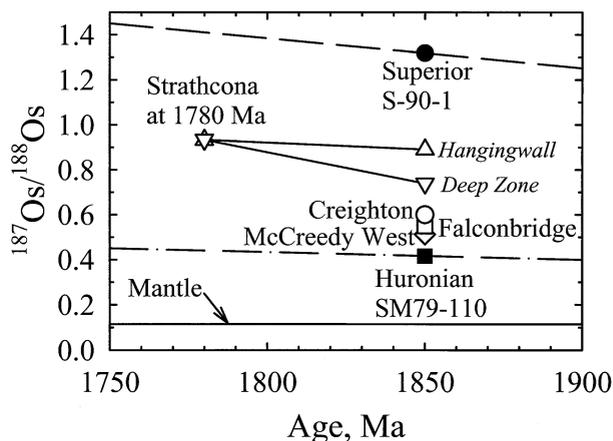

Fig. 11. ^{187}Os/^{188}Os isotopic composition of Sudbury ores at 1850 Ma plot between two possible end members; an unaltered tonalitic Levack gneiss (sample S-90-1) from late Archean country rock adjacent to the North Range and a Huronian metasedimentary rock (sample SM-75-110) of probable Superior Province provenance. The lines through the Archean and Proterozoic end members are hypothetical growth curves for derivation from the mantle at ca. 2750 Ma.

0.416 ± 0.019 and the model age is 2740 Ma. As was the case with the Pb isotopes, the country rock data suggest that the initial ^{187}Os/^{188}Os ratios of all the ores fall between those of the country rocks (Fig. 11). We may construct two mixing models to assess the contribution of each source to the Os isotopic composition of the ores. The first is based on the contribution of the rocks representing the end members, and this solution factors in the Os content of each. The second solution is based on the fraction of Os of each end member composition required to satisfy the observed isotopic compositions.

In constructing the mixing model based on the observed Os abundances and compositions of S-90-1 and SM79-110, we calculate the variation of the ^{187}Os/^{188}Os at 1850 Ma as a function of the mass fraction of the former, the Superior Province component (Fig. 12). In the model, the radiogenic composition only partly compensates for the low absolute abundance of Os in the tonalitic gneiss. Even the least radiogenic ores (McCreedy West, Falconbridge, and Creighton) require between 66 and 80 % of the Superior Province rock similar to S-90-1 in the impact melt to produce the observed isotopic ratios. For the more radiogenic Strathcona ores, still larger fractions are required. If the apparent 1780 Ma Strathcona isochron is interpreted as a mixing line between ores that are in fact 1850 Ma old, then the initial ^{187}Os/^{188}Os at 1850 Ma is 0.739 ± 0.015 for the Deep Zone and 0.891 ± 0.002 for the Hangingwall Zone. These values correspond to a Superior Province rock fraction of 90% and 95%, respectively. If, on the other hand, the Strathcona systematics were reset at 1780 Ma, the original Os composition at 1850 Ma was probably close to that of the Hangingwall ores since these contain most of the Os. If this were the case, an admixture of Superior Province material of close to 95% would satisfy the Os isotopic evidence.

The mixing model between source materials from the Superior Province and the Huronian Supergroup is consistent with the Nd, Pb, and Re–Os isotope systematics, and in that respect may answer the question of the immediate provenance of the
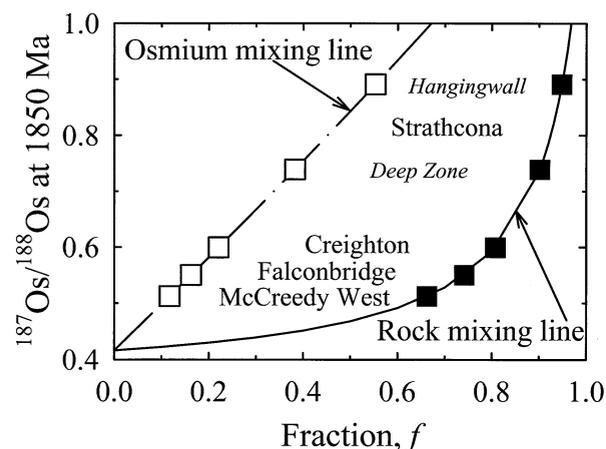
Fig. 12. Mixing model for $^{187}Os/^{188}Os$ in Sudbury ores assuming an Archean end member (Superior S-90-1) with 0.0072 ppb Os and ($^{187}Os/^{188}Os)_{1850\ Ma}$ = 1.319, and a Proterozoic end member (Huronian SM-75-110) with 0.108 ppb Os and ($^{187}Os/^{188}Os)_{1850\ Ma}$ = 0.416.

Fig. 13. Growth of $^{186}Os/^{188}Os$ in Proterozoic mafic dykes. Growth line is calculated assuming Pt/Os $\approx$ Pt/Ir = 100, corresponding to $^{190}Pt/^{188}Os$ = 0.09. The Strathcona ore $^{186}Os/^{188}Os$ (open square) is the present day value. Points for Falconbridge and McCreedy West (upright and inverted triangles, respectively) are $^{186}Os/^{188}Os$ values at 1850 Ma.

melt rock from which the SIC and the sulfides originated. There are difficulties with this model in terms of the $^{190}Pt–^{186}Os$ systematics, however. As adumbrated earlier, $^{186}Os/^{188}Os$ in the Strathcona Hangingwall samples is subchondritic at 1780 Ma, whereas there is significant ^{186}Os enrichment in McCreedy West and Falconbridge. If $^{186}Os/^{188}Os$ enrichment in the latter pair of mines is a reflection of the Huronian contribution, then there should be a measurable enrichment in the Strathcona samples. In the *rock* mixing model, the Strathcona Hangingwall samples contain only ca. 5% Huronian contribution, but, as shown in Fig. 12, this corresponds to ca. 45% of the Os. Thus in a simple two component rock mixing model we would see at Strathcona slightly less than half of the ^{186}Os enrichment seen at McCreedy West and Falconbridge. Since this is not apparent, either we have chosen the wrong end members, or there is a minor third component not present at Strathcona that gives the enriched ^{186}Os at McCreedy West and Falconbridge. The end members, Levack tonalitic gneiss and Huronian metasedimentary rock, were most likely the major target rocks, but a reconstruction of the target area suggests the presence of mafic rocks that could contribute Os with a near- mantle isotopic composition. Magmatic rocks in the southern part of the target area may have appreciable PGE abundances because some boast substantial Ni contents; the Nipissing gabbros, for example, contain $\approx$130 ppm Ni (Golightly, 1994). If these mafic rocks were derived from the upper mantle, upon eruption they would carry the chondritic $^{186}Os/^{188}Os$ ratio that predominates in mantle materials: osmiridium, ophiolitic chromites, and abyssal peridotites (Walker et al., 1997; Brandon et al., 1998; 1999; 2000). Nevertheless, the basalts may have sufficiently high Pt/Os ratios to generate measurable $^{186}Os/^{188}Os$ enrichment in the period between the age of the formation of the Superior Province and the impact of the Sudbury bolide. Data for both Os and Pt in basic rocks of this age are scarce, but Proterozoic dyke swarms in the Superior Province have mean Ir and Pt contents of 0.09 ppb Ir and 9 ppb Pt (Brügmann et al., 1990, cited by Brügmann et al., 1993). Assuming that Os and Ir abundances are similar in these dykes, then the Pt/Ir ratio of $\approx$100 translates to $^{190}Pt/^{188}Os$ = 0.095. The growth of

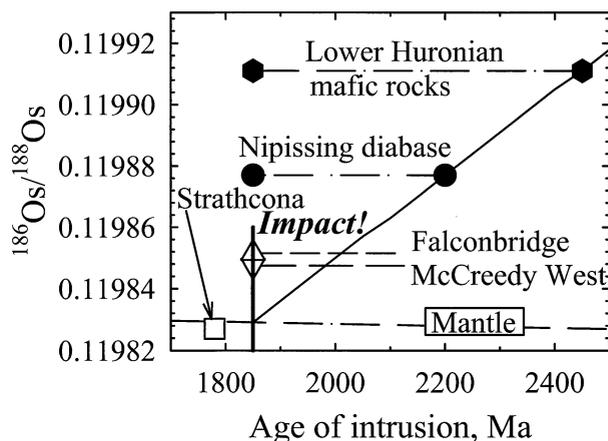$^{186}Os/^{188}Os$ from rocks of this composition can be appreciable in the time period between the formation of the Levack gneiss at 2711 Ma and the Sudbury impact (Fig. 13). The amount of mafic rock required to produce $^{186}Os/^{188}Os$ similar to that in Falconbridge or McCreedy West ores depends on the age of the mafic component, but about 30% (Lower Huronian mafics) to 50% (Nipissing diabase) seems reasonable. Mixing models suggest that this much mafic rock can be accommodated by the major element chemistry of the SIC, even though there is considerable latitude in the choice of actual components (Walker et al., 1991; 1994; Golightly, 1994; Grieve, 1994). Clearly, the value of the isotopic mixing models could be greatly improved by precise measurements of $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$ in the country rocks. Recent technical developments for the low-blank analysis of very large samples now make this feasible (Brandon et al., 1998; 1999; 2000).

Admixture of Proterozoic mafic rocks may explain enhanced $^{186}Os/^{188}Os$, but these rocks also could markedly affect $^{187}Os/^{188}Os$. There are no reliable Re–Os isotopic data for the suite of target mafic rocks in the Sudbury area. Compilation of basaltic and komatiitic rock data indicates, however, that $^{187}Re/^{188}Os$ could range between unity and several thousands, with basalts assuming the higher values (Shirey and Walker, 1998). Taking a geometric mean $^{187}Re/^{188}Os$ of ca. 300 for basalts, in the 350 Ma and 600 Ma before the Sudbury impact $^{187}Os/^{188}Os$ would become 1.9 and 3.0 for Huronian and Nipissing mafics, respectively. The range of $^{187}Os/^{188}Os$ introduced by a mafic-ultramafic component is constrained by data for ultramafic inclusions from the sublayer of the SIC sampled at the Whistle mine (Lightfoot et al., 1997; Cohen et al., 2000). A suite of 14 melanorite inclusions includes three rock types: S-rich melanorite, S-rich olivine melanorite, and S-poor olivine melanorite. The S-rich melanorites (n = 5) have 0.69–21.4 % S, 660–1790 ppm Ni, 6–12 ppb Re, and 1 to 1.5 ppb Os. These rocks have model ages between 2430 and 2580 Ma averaging 2500 $\pm$ 60 Ma and therefore largely derive from Huronian magmatic rocks, with a slight admixture of older material. Initial $^{187}Os/$

[188]Os at 1850 Ma range between 0.4 and 0.72 with an average of 0.63 ± 0.12 that is within the range in the ores. S-rich olivine melanorites ($n = 4$) contain 0.4–1.9 % S, 780–1820 ppm Ni, 4–18 ppb Re and 0.4–4 ppb Os. These rocks have more variable model ages, ranging from 2640 to 3090 Ma with a mean of 2840 ± 190 Ma that is within error of the Superior province basement age. Initial $^{187}Os/^{188}Os$ ratios at 1850 Ma are more variable than in olivine-poor melanorites. Three of the values vary between 0.51 and 0.79, and the fourth value is 1.74. The mean of the four values is similar to the Strathcona initial at ca. 0.9. These results suggest that a contribution of material similar in composition to the S-rich melanorites and olivine melanorites sufficient to enhance $^{186}Os/^{188}Os$ will not outrageously perturb the $^{187}Os/^{188}Os$ systematics of the resulting ores.

The S-poor olivine melanorites are less radiogenic than the other inclusions, with initial $^{187}Os/^{188}Os$ ratios at 1850 Ma between 0.11 and 0.24. These samples give an age of 1970 ± 250 Ma that is within uncertainty of the time of the Sudbury impact. The lowest initial ratios are found in two of the most S-poor samples. Yet these inclusions have amongst the highest Ni and Os contents of this group of samples, suggesting a source of siderophiles independent of the almost ubiquitous sulfides.

4.2. Where is the Meteorite?

It is generally agreed now that the SIC was generated by a meteorite impact, and yet little evidence has been found of the signature of the impacting body. Highly siderophile elements (primarily PGE and particularly Ir) are a sensitive indicator of meteoritic influx (Peucker-Ehrenbrink and Ravizza, 2000) and impact (Evans et al., 1993). Siderophile element analysis has been outstandingly successful in identification of the worldwide chondritic signature of impact at the Cretaceous–Tertiary boundary (Ganapathy, 1980; Kastner et al., 1984; Evans et al., 1993), but this achievement has distracted attention from puzzling results at impact craters recognized by other criteria. Melt rocks from smaller craters often carry a signature of the impactor as, for example, at the 8.5 km Wanapitei Lake crater (Wolf et al., 1980; Grieve and Ber, 1994). In craters larger than ca. 30 km diam., however, melt rocks often show little or no PGE enrichment as at the 70 km Manicouagan, Quebec crater (Wolf et al., 1980). Nevertheless, the size distinction is not always clearcut since small craters such as the 1.8 km diameter Lonar, India, crater may be found with no meteoritic signature (Morgan, 1978), whereas the ≈70 km Morokweng, South Africa, crater has impact melts containing large amounts of siderophiles (Koeberl et al., 1997; Reimold and Koeberl, 1999). Clearly, the ability to detect a meteoritic component in melt rock within a crater is a complex function of meteorite type, sampling strategy, country rock composition, and cratering mechanics.

High levels of PGE in the SIC make meteoritic material more difficult to detect than at most craters, but Os isotope systematics provides a sensitive indicator (Koeberl and Shirey, 1997). Most meteorites that contain significant amounts of siderophile elements have $^{187}Os/^{188}Os$ ratios well below those in Sudbury ores and breccias. The most promising candidate for a meteoritic signature at Sudbury may be found among the
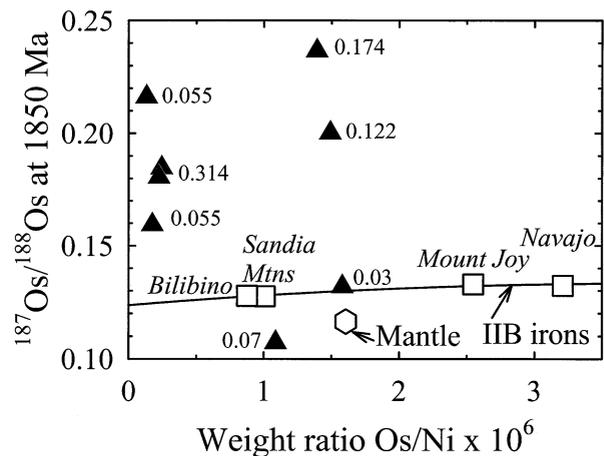

Fig. 14. Comparison of $^{187}Os/^{188}Os$ vs. Os/Ni in S-poor olivine melanorites from the Whistle mine (Cohen et al., 2000; solid upright triangles), with those in IIB iron meteorites (Morgan et al., 1995; open squares) and the Earth's upper mantle (Meisel et al., 2001; Morgan, 1986; open hexagon). Sulfur contents, in percent, of the melanorites are shown.

melanorite inclusions from the Whistle mine (Cohen et al., 2000). The S-rich inclusions appear to have Os isotopic signatures derived from old crustal sources and give Re–Os ages of 2460 ± 530 Ma and 2670 ± 320 Ma for melanorites and olivine melanorites, respectively. The most S-poor samples also contain significant Ni and Os, however, and in fact are higher in these elements than many inclusions with moderate S content. Furthermore, the S-poor melanorites have low initial $^{187}Os/^{188}Os$ in the meteorite range and a Re–Os isochron age of 1970 ± 210 Ma that clearly suggests a temporal relation to the impact. Thus, contamination in these samples by sulfur-rich material with a crustal signature is small and is insignificant for the most S-poor inclusions. Cohen et al. (2000) demonstrated that the variation of $(^{187}Os/^{188}Os)_{1850\ Ma}$ vs. S content can be adequately emulated by mixing between sulfide with a radiogenic Os crustal signature and a "silicate" component with unradiogenic Os. These authors consider the silicate component to be formed at the time of impact as either mantle-derived magma, or as differentiates from the melting of high-Mg basement rock. A third possibility may be that the Os isotope signature is meteoritic and derived from the impactor.

In S-poor melanorites, Os/Ni weight ratios range between 0.13×10^{-6} to 1.6×10^{-6} and are significantly lower than the CI chondrite value of 44×10^{-6} (Anders and Grevesse, 1989). This range, however, resembles that of such fractionated magmatic irons as the IIB meteorites (Morgan et al., 1995). On a plot of $(^{187}Os/^{188}Os)_{1850\ Ma}$ vs. Os/Ni, S-poor ultramafic inclusions fall close to the iron meteorite trend (Fig. 14). Inclusion 92PCL-18 has the lowest S content (0.03%) of the ultramafics analyzed by Cohen et al. (2000), but contains as much Ni and Os as many S-rich melanorites; 930 ppm and 1.5 ppb, respectively. With Os/Ni ratio = 1.6×10^{-6} and $(^{187}Os/^{188}Os)_{1850\ Ma} = 0.132$, this melanorite lie closely on the IIB trend, plotting between the iron meteorites Mount Joy and Sandia Mountains (Morgan et al., 1995). The evidence of this plot is not conclusive, since terrestrial mantle peridotites also plot quite close to the intersection of the sulfur-poor inclusion and the iron meteorite trends. The melanorite

92PCL-18 has (^{187}Re/^{188}Os) = 0.93, however, thus providing additional evidence for a IIB iron connection since this value is different from the chondritic to subchondritic ratios in peridotites but only slightly higher than the ratio in IIB irons; Mount Joy has (^{187}Re/^{188}Os) = 0.80, for example (Morgan et al., 1995). Cohen et al. (2000) did not measure ^{186}Os/^{188}Os in the Whistle inclusions, but this would readily have distinguished between ($_{186}$Os/^{188}Os)$_{1850\ Ma}$ of 0.119 832 for the mantle and the corresponding IIB iron values of 0.120 to 0.121 (Walker et al., 1997). If there were a small IIB iron component, it might provide a significant enrichment of ^{186}Os/^{188}Os in the melt rock.

The IIB iron meteorites are the result of significant fractional crystallization of an asteroidal core. An impactor derived from such a core would retain the IIB signature only if the parent body were severely fragmented to remove the relatively unfractionated IIA irons. There are two major opinions how asteroidal cores crystallized; as a whole solidifying from the center outward (Kracher and Wasson, 1982; Pernicka and Wasson, 1987), or as kilometer-sized dendrites extending inward from the core-mantle boundary of the asteroid (Narayan and Goldstein, 1982; Haack and Scott, 1993; Morgan et al., 1995). Only the former model is likely to segregate enough mass of IIB iron for crater-forming purposes. Even for a large asteroidal core it would be difficult to derive a purely metallic object of average IIB composition large enough to produce the Sudbury crater. A composite bolide containing part of the outer core and lower mantle might satisfy present observations, and provide at least some Mg enrichment.

Impact of a bolide with significant metal content might be expected to introduce large amounts of Os, but in IIB iron the content of this element is actually low compared to the 800 to 66,000 ppb amounts found in IIA irons. For example, such meteorites as Mount Joy and Sandia Mountains with Os/Ni and ^{187}Os/^{188}Os bracketing those of the most S-poor olivine melanorite inclusion have only 145–59 ppb Os (Morgan et al., 1995). Thus the amount of Os and Ni in inclusion 92PCL-18 would correspond to an addition of 1–2.5 % of IIB metal phase. But in solid metal–liquid metal fractionation, Pt abundances vary only slightly, and these same meteorites contain 15,000 ppb Pt and 10,000 ppb Pt, respectively (Pernicka and Wasson, 1987; Walker et al., 1997). Thus, although the projectile may introduce small amounts of Os that can only be distinguished isotopically, there is the potential to introduce significant amounts of Pt.

4.3. ^{186}Os/^{188}Os in Ancient Continental Crust

The Sudbury ores and the unique positioning of the SIC allow us to sample two upper continental crustal terrains; Superior Province, representing the mid levels of the Archean crust, and the Huronian Proterozoic upper crust. If the mixing model discussed earlier is correct, Os in the Hangingwall ores at the Strathcona mine was sampled and extracted by the Sudbury impact from an approximately 1:1 mixture of Os from Archean Superior Province sedimentary rocks and Os from Huronian sediments. The present day mean ^{186}Os/^{188}Os of these ores of 0.119 8269 ± 0.000 0035 is nonradiogenic relative to the mantle value of 0.119 8362 ± 0.000 0010 in abyssal peridotites (Brandon et al., 2000). The mantle growth curve for ^{186}Os/^{188}Os and the upper mantle ^{190}Pt/^{188}Os are not well known, but a mantle ^{190}Pt/^{188}Os of 0.002 369 may be estimated from the abyssal peridotite value and the solar system initial value of 0.119 820 (Walker et al., 1997). By assuming ^{190}Pt/^{188}Os of sediments to be zero, we may estimate a minimum age of separation from the mantle to be ≈2600 Ma. If we take the canonical Superior Province age of 2711 Ma, the observed present day ^{186}Os/^{188}Os 0.119 8269 in Strathcona implies ^{190}Pt/^{188}Os in the ore precursor of only 0.0001. This value is lower than even the modest ^{190}Pt/^{188}Os ratios of 0.004 60 and 0.001 58 in ores PGM79-138 and PGM79-139, respectively. Apparently, the Pt values in these ores are the result of very recent enrichment, sample cross contamination or analytical error. In any event, it is clear that Archean or Proterozoic metasedimentary crust as represented by Superior Province and Huronian Supergroup rocks are unlikely sources for Pt in Sudbury ores.

If the extremely low Pt/Os ratios inferred above are true, then this result begs the question of the provenance of Pt in the Sudbury ores. There appear to be two possible sources. The impacting body as a source of Pt was discussed in the preceding section. A more likely source of Pt in the impact melt is the Archean and Proterozoic mafic rocks. The Os isotopic compositions of Falconbridge and McCreedy West ore may be regarded as a composite of the Proterozoic metasedimentary crust represented by Strathcona, and of Archean and Proterozoic mafic rocks, sampled at 1850 Ma. The Os isotopic evolution of such a crustal composite has important geochemical implications additional to those related directly to the Sudbury ores.

The coupled ^{187}Re–^{187}Os and ^{190}Pt–^{186}Os systems in some plume basalts suggest that the source of these magmas may be at the core–mantle boundary and that it has been slightly contaminated by siderophile-rich outer core metal (Walker et al., 1995; 1997; Brandon et al., 1998; 1999). Crystallization of the inner core may significantly enhance ^{187}Re/^{188}Os and ^{190}Pt/^{188}Os in the outer core. Consequently, the observed ratios in plume basalts may be accommodated within the range of experimentally determined solid metal –liquid metal partition coefficients at high pressure and temperature (Walker, 2000) and by models of core growth (Stevenson et al., 1983; Stacey and Loper, 1984). Alternatively, crustal rocks often have very high Re/Os and some may have high Pt/Os. Long-term decay of ^{190}Pt and ^{187}Re in these rocks then could result in enhanced ^{186}Os/^{188}Os and ^{187}Os/^{188}Os, respectively. Incorporation of ancient recycled crust in plumes has been proffered as an explanation of the radiogenic ^{187}Os/^{188}Os ratios in Hawaiian lavas and other ocean island basalts (Hauri and Hart, 1993, Hauri et al., 1996; Lassiter and Hauri, 1998). Analysis of loess and terrigenous sediments shows that old continental crust is highly radiogenic in ^{187}Os/^{188}Os, ca. 1.2–1.3 (Esser and Turekian, 1993), but there are no corresponding estimates published for ^{186}Os/^{188}Os. If we are to investigate the effect of mixing such crustal material into the source of plume basalts, we need to estimate the Os isotope systematics close to the present day. An important point in the inference of a contribution of outer core material has been the correlation between ^{186}Os/^{188}Os and ^{187}Os/^{188}Os in Hawaiian basalts (Brandon et al., 1998; 1999). If this correlation were due to crustal contamination, then the contaminant must lie on an extension of the plume basalt correlation line. We can estimate the likely values of these ratios in the mafic component from the ^{186}Os/^{188}Os

and $^{187}Os/^{188}Os$ ratios in the Falconbridge and McCreedy West ores, the Huronian metasediment and the mantle. We can relate $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$ in the mafic component from the following equation:

$$\left(\frac{^{187}Os}{^{188}Os}\right)_B - \left(\frac{^{187}Os}{^{188}Os}\right)_M = \left[\left(\frac{^{186}Os}{^{188}Os}\right)_B - \left(\frac{^{186}Os}{^{188}Os}\right)_M\right]$$

$$\times \left[\frac{\left(\frac{^{187}Os}{^{188}Os}\right)_M - \left(\frac{^{187}Os}{^{188}Os}\right)_A}{\left(\frac{^{186}Os}{^{188}Os}\right)_M - \left(\frac{^{186}Os}{^{188}Os}\right)_A}\right], \quad (1)$$

where component A is the sediment, component B is the mafic rock(s), and M represents the mixture of sediment and mafic rocks. All ratios are calculated at 1850 Ma. We do not know the mafic component ratios in the equation above, but can estimate other values as follows.

The ratios in the mixture are simply the means of the ratios in the Falconbridge and McCreedy West ores, giving $(^{186}Os/^{188}Os)_M = 0.119\,8495$ and $(^{187}Os/^{188}Os)_M = 0.5312$. We put sedimentary $(^{186}Os/^{188}Os)_A = 0.119\,8269$ from the Strathcona measurement, and $(^{187}Os/^{188}Os)_A = 0.416$ from metasediment SM75-110 which appears to provide most of the Os in the McCreedy West and Falconbridge ores.

If we choose a time for separation of the mafic component from the mantle, T_0, and a likely range of Pt/Os ratios, we may evaluate $(^{186}Os/^{188}Os)_B$ in the mafic component from the equation

$$\left(\frac{^{186}Os}{^{188}Os}\right)_B^{1850} = \left(\frac{^{186}Os}{^{188}Os}\right)_{Mantle}^{T_0} + \left(\frac{^{190}Pt}{^{188}Os}\right)_B^{Now} \times (e^{\lambda T_0} - e^{\lambda\,1850\times10^6}).$$

$$(2)$$

By substituting the values of $(^{186}Os/^{188}Os)_B$ in Eqn. (1), we can then calculate the corresponding $(^{187}Os/^{188}Os)_B$ and in turn the appropriate present day $(^{187}Re/^{188}Os)_B$ from the equation

$$\left(\frac{^{187}Re}{^{188}Os}\right)_B^{now} = \frac{\left(\frac{^{187}Os}{^{188}Os}\right)_B^{1850} - \left(\frac{^{187}Os}{^{188}Os}\right)_{Mantle}^{T_0}}{(e^{\lambda T_0} - e^{\lambda\cdot1850\times10^6})}. \quad (3)$$

From these data we calculate the present day $^{186}Os/^{188}Os$ and $^{187}Os/^{188}Os$ for a range of $^{190}Pt/^{188}Os$, from 0.002—approximately mantle—to 0.1, the value estimated from the Pt/Ir in Proterozoic dyke swarms from the Superior Province (Brügmann et al., 1990). For illustrative purposes we have used two values for the age of separation of the mafic component from the mantle, T_0; 2700 Ma (Superior) and 2450 Ma (Huronian). These values are compared with results for Hawaiian basalts and mantle peridotites (Fig. 15). Although the mafic component from Sudbury is likely to be significantly enhanced in $^{186}Os/^{188}Os$, the isotopic signature is even more markedly enriched in $^{187}Os/^{188}Os$. On a plot of $^{186}Os/^{188}Os$ vs. $^{187}Os/^{188}Os$, the resulting slope is shallower than that of plume-derived material. The trajectories of isotopic composition cross the mantle value for $^{186}Os/^{188}Os$ at high values of $^{187}Os/^{188}Os$; 1.10 and 1.39 for T_0 ages of 2711 Ma (Superior) and 2450 Ma (Huronian), respectively. Thus, the crustal mafic rocks as represented at Sudbury could not be end members for a mixing
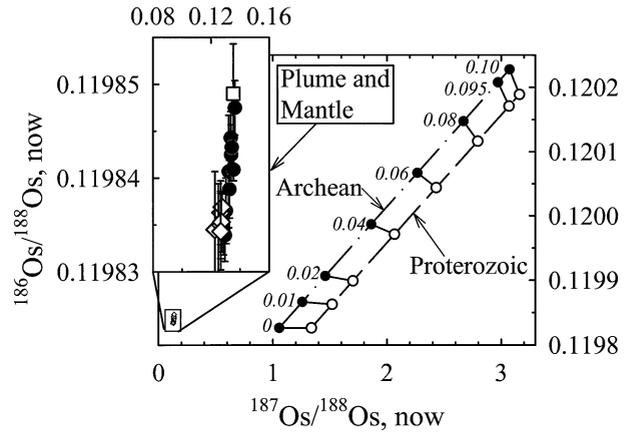

Fig. 15. Comparison of $^{186}Os/^{188}Os$ vs. $^{187}Os/^{188}Os$ in possible Archean and Proterozoic mafic component with those in plume basalts and Noril'sk ore. Because of the relatively shallow slope of the crustal mafic components these cannot be end members in any plausible crustal contamination model. Numbers and the symbols on the lines in the main diagram denote the assumed $^{190}Pt/^{188}Os$ used to calculate values. In the inset diagram, which gives a greatly magnified view of the mantle and plume data, the open diamonds are abyssal peridotites (Brandon et al., 2000), the filled circles are Hawai'ian plume basalts (Brandon et al., 1999), and the open square is Noril'sk ore KZ17133-931.3 (Walker et al., 1997).

line that produces the array seen in plume-derived material. The correlated Os isotopic data in plume material are best attributed to an admixture of outer core material in the source region of the plumes near the core–mantle boundary.

5. SUMMARY

Ores from McCreedy West and Falconbridge give imprecise isochron ages in agreement with the canonical age of the Sudbury Igneous Complex of 1850 Ma, and yield initial $^{187}Os/^{188}Os$ ratios of 0.514 and 0.550, respectively. Initial $^{186}Os/^{188}Os$ ratios for the same mines have radiogenic values averaging 0.119\,8475 and 0.119\,8515, respectively. Hangingwall and Deep Zone ores from the Strathcona mine give an isochron age of 1780 ± 7 Ma and a precisely defined initial $^{187}Os/^{188}Os = 0.934$. The regression seems not to be a mixing line, and Os isotopics may have been reset by post-Penokean dyke activity. Two Hangingwall ores with very low Pt have nonradiogenic *present day* $^{186}Os/^{188}Os$ averaging 0.119\,8269. This value is almost identical to the mantle value at 2700 Ma, the age of the Levack gneiss. The high initial $^{187}Os/^{188}Os$ and low $^{186}Os/^{188}Os$ at Strathcona cannot be derived simply from material with an Os isotopic signature similar to that at McCreedy West and Falconbridge.

Deep Copper Zone ores at Strathcona have low Re and Os, high Pt and highly radiogenic Os isotopic signatures. These ores give a Re–Os age of 1894 ± 230 Ma with an initial $^{187}Os/^{188}Os$ of 0.9 ± 1.1. Utilizing the Strathcona initial of 0.934, the Deep Copper Zone samples yield a mean model age of 1883 ± 54 Ma that agrees with the 1850 Ma age of the SIC and is distinct from the age of 1780 Ma of other Strathcona ores. The DCZ ores have an extremely radiogenic initial $^{186}Os/^{188}Os$ of 0.163. One ore, DCZ-19, has the most radiogenic

^{186}Os/^{188}Os yet reported, 1.60 ± 0.04, and yields a model Pt–Os age of 1884 ± 62 Ma.

A binary mixture of Superior Province and Huronian metasedimentary rocks can reproduce the ^{187}Re–^{187}Os isotope systematics of Sudbury ores, but ^{186}Os/^{188}Os enrichment at McCreedy West and Falconbridge must be due to a third component, most likely Archean or Paleozoic mafic rocks sampled by the Sudbury impact. These mafic rocks may have substantially enriched ^{186}Os/^{188}Os that is accompanied by correspondingly high ^{187}Os/^{188}Os. The trajectories of isotopic composition indicate that the correlated ^{186}Os/^{188}Os–^{187}Os/^{188}Os array in plume-derived material is not the result of contamination by such crustal mafic rocks. The Os isotopic data in plume material are best explained by an admixture of outer core material to the source region of the plumes near the core-mantle boundary.

The Sudbury Structure is the product of asteroid impact, but no distinct trace of the impactor has been reported. S-poor olivine melanorite inclusions from the Whistle mine at Sudbury have high Ni and Os with low ^{187}Os/^{188}Os (Cohen et al., 2000) that may be the signature of a bolide containing part of a magmatically fractionated asteroidal core. The present evidence is equally well satisfied by a contribution of mantle peridotite, however.

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