AN OVERVIEW OF VIS-NIR-SWIR FIELD SPECTROSCOPY AS APPLIED TO PRECIOUS METALS EXPLORATION

By
Phoebe Hauff
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ABSTRACT

Field portable Reflectance Spectroscopy has made a major impact on the exploration industry because it has changed the way geologists think about their rocks and about the different alteration systems and relationships between associated minerals. It has opened new vistas for the field geologist. It allows him now to "think out of the box" and not be tied to models that are not quite his deposit type. Portable spectrometers at the outcrop, in the core shack, in the open pit, give the explorationist instant information about rocks that may all have similar physical appearances. Portable spectrometers provide in-situ data that allows the geologist to immediately integrate several data sets - geophysics, geochemistry and mineralogy - in the field - at the outcrop, where this information can be quickly applied to a better understanding of what is physically observed - instant gratification. This was not possible before their invention. Reflectance spectroscopy is changing the nomenclature of exploration and allowing geologists to look at mineral associations that provide more in depth understanding of not only the rocks themselves, but the processes that have altered them into the precious metal resources so important to the world.

This paper will therefore present an overview of visible through SWIR (Short Wave Infrared) field spectroscopy with a brief summary of theory and applications. Emphasis is placed on alteration systems, especially those associated with low and high sulfidation gold deposits, porphyries, kimberlites, skarns, IOCG (iron oxide copper gold) and unconformity uranium deposits, all of which are major successes for reflectance spectroscopy.

Introduction -

Many innovations have occurred in exploration and mining that have contributed to assisting the industry to find more precious metals resource and to find it faster. One such innovative technology is that of VIS-NIR-SWIR (Visible – Near-Infrared - Short-wave Infrared) reflectance spectroscopy. Its use has caused major changes in the way exploration is done and how explorationists think as now they are provided with instant information about the minerals in their rocks.
The utilization of this range (350 to 2500nm) of the Electromagnetic Spectrum for mineral exploration comes out of remote sensing needs as does the invention and subsequent refinements of the field spectrometers, which have become the prime vehicles for this methodology. Initial field units from the GER Mark series (Geophysical Environmental Resources of Millbrook New York) to the ASD classics and FieldSpec Pros were developed using solar illumination to correspond to the satellite and airborne sensors for which they provided ground truth or verification.

With the addition of an internal light source to eliminate the effects of atmospheric water spectra, users quickly discovered that the portable spectrometers had so many more applications then just ground truthing remotely sensed data and the technology came into its own.

Although there are numerous companies offering instruments in the Visible and short wave infrared for applications ranging from vegetation, water, forestry, wood products, paint, carpets, wheat protein content, pharmaceuticals, forensics, plastics, oil, among others, most are not portable to the field nor general enough for geologic applications. GER was the company with the Mark series of field spectrometers, who has always been more involved in the general remote sensing side of the application rather then specifically targeting the mining industry. Alex Goetz, at that time of Jet Propulsion Laboratory, Ron Lyon of Stanford University, Larry Rowan of the U.S. Geological Survey, Robert Agar of AGARS, Australia, Jon Huntington, CSIRO, Australia, Fred Kruse of AIG, Floyd Sabins of Chevron, Jim Taranik of the University of Nevada-Reno, and the present author, among hundreds of others, have all used this instrument for mostly geologic, but some exploration and mining, applications. At issue has always been the solar illumination and the corresponding atmospheric water interference bands that mask out vital components and absorption features in the mineral spectra. This interference made the application less attractive for alteration characterization. Even 15 years ago, the solar illumination-based technology was the best available.

With the introduction of the internal light source, the PIMA-II field spectrometer by Integrated Spectronics of Sydney, Australia, in the early 1990s, the concept of portability and field characterization of alteration took on a new perspective and many new applications. Over 200 PIMAs, the majority of which are still operational, were pioneering devices for developing new methods of alteration mapping, core logging, drill well targeting, and field reconnaissance. Most of the major mining companies worldwide adopted this new technology and utilized it well for new discoveries and in-depth knowledge of their alteration systems.

In the late 1990s, ASD Inc. (formerly Analytical Spectral Devices, Inc.) introduced their FieldSpec™ Pro. This instrument did have an internal light source probe on it. It was upgraded in 2005 to the “FR3™”, which had a refined probe. Neither of these instruments were designed for the minerals industry, but
were designed more for general remote sensing with solar illumination as the prime light source. It was the appearance of the ASD TerraSpec™ in 2004 with its flexible and innovative design specifically for minerals and mining that has brought cutting edge, fast, fiber-optic data collection into the field and core shack.

This paper will review the concepts of reflectance spectroscopy, and look at the more practical information that can be obtained from the visible, near-infrared, and short-wave infrared regions of spectra, beyond just simple mineral identification. Routine applications such as core logging, field reconnaissance, pit mapping, and remote sensing ground truth will be illustrated with examples. Selected alteration systems for gold, porphyries, diamonds, unconformity uranium, IOCGs and skarns will be summarized from a spectral perspective and mini-case studies discussed, space permitting.

REFLECTANCE SPECTROSCOPY

Spectroscopy is the study of the interaction between energy as electromagnetic radiation (EMR) and matter. Radiation can be absorbed, emitted or scattered by matter.

Reflectance spectroscopy can be defined as the technique that uses the energy in the Visible (0.4-0.7), Near-Infrared (0.7-1.0) and Short-Wave Infrared (1.0-2.5 µm) wavelength regions of the electromagnetic spectrum (Figure 1) to analyze minerals.

![Generalised reflectance spectra](image)

Figure 1. – Highly generalized cartoon of the EMS showing the various regions that are covered by field portable spectrometers. Source unknown.

The science and techniques of reflectance spectroscopy are based on the spectral properties of materials. Certain atoms and molecules absorb energy as a function of their atomic structures. (Hunt, 1977 and Goetz et al, 1982).
In the visible region, this is caused by electronic transitions such as Crystal Field Effects (atomic energy level transitions), Charge Transfer (inter-element electronic transitions), Conduction Band Transitions (electron transfer over spatially close energy levels) and Color Center Phenomenon (lattice defect induced energy levels). Much of this simply involves release of energy when an electron changes energy levels in an atom.

Absorption features in the SWIR region are a function of the composition of the mineral. They are a manifestation of energy absorption within the crystal lattice from vibrational state transitions. Because these vibrational states correspond to distinct energy levels, the absorption features occur at well-defined wavelength positions. The energy levels that define these wavelengths are a function of the size of the ionic radii of the cations bonded to different molecules. The bonds will vibrate at different wavelengths as a function of the length of the bond. Because the bond lengths between a specific atom and molecule will be consistent, it is possible to predict compositions and compositional changes in minerals being analyzed by the wavelengths and wavelength shifts. (Hunt, 1977)

The transitions between energy levels and compositional differences are manifested by absorption features at defined wavelengths. The common absorption feature positions are listed in Table I.

<table>
<thead>
<tr>
<th>POSITION</th>
<th>MECHANISM</th>
<th>MINERAL GROUP</th>
</tr>
</thead>
<tbody>
<tr>
<td>~1.4 µm</td>
<td>OH and WATER</td>
<td>CLAYS, SULFATES HYDROXIDES, ZEOLITES</td>
</tr>
<tr>
<td>~1.56 µm</td>
<td>NH₄</td>
<td>NH₄ SPECIES</td>
</tr>
<tr>
<td>~1.8 µm</td>
<td>OH</td>
<td>SULFATES</td>
</tr>
<tr>
<td>~1.9 µm</td>
<td>WATER</td>
<td>SMECTITE</td>
</tr>
<tr>
<td>2.02, 2.12 µm</td>
<td>NH₄</td>
<td>NH₄ SPECIES</td>
</tr>
<tr>
<td>~2.2 µm</td>
<td>AL-OH</td>
<td>CLAYS, SULFATES, MICAS</td>
</tr>
<tr>
<td>~2.29</td>
<td>Fe-OH</td>
<td>Fe-CLAYS</td>
</tr>
<tr>
<td>~2.31</td>
<td>Mg-OH</td>
<td>Mg-CLAYS, ORGANICS</td>
</tr>
<tr>
<td>~2.324</td>
<td>Mg-OH</td>
<td>CHLORITES</td>
</tr>
<tr>
<td>~2.35 +/- µm</td>
<td>CO₃⁻²</td>
<td>CARBONATES</td>
</tr>
<tr>
<td>~2.35+</td>
<td>Fe-OH</td>
<td>Fe-CHLORITES</td>
</tr>
</tbody>
</table>

The absorption features occur within a reflectance spectrum, with wavelength positions and distinctive profiles that can be used to identify mineral and organic phases. These are shown in Figure 2 below.

Each mineral has a distinctive spectral signature, composed of several absorption features, which is a function of composition, crystallinity,
concentration, water content and environmental considerations. For mineral identification, that signature is compared against characterized references for a validated identification. This can be done manually by using wavelength tables and overlaying references on the unknown or it can be done using computer algorithms operating against a database of references (i.e., a spectral library). For best results these libraries must be set up for specific deposit types, and not allow the algorithm to make a wrong choice.

![Absorption Feature Diagram](image)

Figure 2 – Example absorption spectrum. The units for the horizontal scale are in nanometers or microns while the vertical scale is percent reflectance. The absorption figure shown is a doublet, with two minima or lowest points of the feature. Each minimum has a distinctive wavelength, which is a function of the chemical composition. “FWHM” means Full Width at Half Maximum. It can be used for symmetry determinations that are related to crystallinity.

Although the computer algorithms are useful for data management and organization, care must be taken when using them for phase identification and quantification. There are many unaddressed variables and non-user chosen identifications are not always accurate. Additionally, important information on crystallinity, water, organics, and iron, among other things, is lost.

**VISIBLE RANGE SIGNATURES**

There is an abundance of information to be gained from the visible region of the EMS. This is one of the features that makes the full-range spectrometers so valuable. Figure 3 shows some of the cations that are recognizable by colors and corresponding minerals. Although the visible range is constrained from 300-
720nm, there is additional and complimentary information in the NIR (700 nm to 1000 nm).

![Figure 3. The plot shows some of the more common cations that can be identified in the visible range.](image)

For example, chlorites are more detectable in the visible range because there is less interference from other species compared to in the SWIR range where their signal is usually low unless chlorites are the dominant phase. Iron oxyhydroxides (Figure 4) have many distinctive features that allow them to be discriminated from each other. This allows a discrimination of pH zoning and something not easily done before. The nickel cation also has diagnostic features in the visible range that can be detected in nickel laterite deposits and correlated to the ore zones. Chrome in kimberlite minerals such as chrome diopside and G-10 garnets make spectroscopy invaluable to the identification of indicator minerals. Manganese is also detectable through features in the visible range.

Garnets, pyroxenes, olivine, selected iron and copper oxides, hydroxides, sulfates, and for heavy element carbonates, a NIR major absorption feature.
Figure 4 – The plots in [A] show the classic three most common iron minerals encountered in exploration; jarosite, hematite and goethite. Stronger acidic minerals are shown in [B]. These include, from top to bottom, schwertmannite, jarosite, copiapite, coquimbite and melanterite.

Rare Earth elements (REE) have very distinctive signatures in the visible region as can be seen in Figure 5. Rare Earth minerals include such species as bastnesite (CeFCO$_3$), monzonite (Ce, La,Y,Th) PO$_4$, and xenotime (YPO$_4$). The two most famous REE mines in the world are located at Mountain Pass, California and Bayan Obo, China.

Figure 5 – Plots of europium, neodymium oxide, samarium oxide, praseodymium oxide from the USGS reference library.
SWIR RANGE SIGNATURES

The field spectrometers are probably the best tools invented in the last 10 years, aside from GPS, for field reconnaissance and core logging. They can be taken to the field, to the pit and to the outcrops. They operate from a backpack or from the field vehicle. They can be set up in a field camp, in a core shack, or on an outcrop. They are versatile and flexible and provide the explorationist with "on the spot", in-situ information, right at the outcrop, so there is instant gratification and instant knowledge of the composition of the outcrop or the drill core. Alteration types are instantly identified, mixtures are seen, zoning can be tracked, mineral transition zones characterized. Ten years ago, this was just becoming a possibility. Now it can be done in a tenth of a second.

What can be seen? Table II, from Thompson and Thompson (1996) summarizes the majority of the common infrared-active minerals that can be detected in the SWIR range.

Table II – Summary of Infrared Active Minerals with SWIR Absorption Features

<table>
<thead>
<tr>
<th>Environment of formation</th>
<th>Standard terminology</th>
<th>SWIR active mineral assemblage (key minerals are in bold)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intrusion-related</td>
<td>Potassic (biotite-rich), K silicate, biotite</td>
<td>Biotite (phlogopite), actinolite, sericite, chlorite, epidote, muscovite, anhydrite</td>
</tr>
<tr>
<td></td>
<td>Sodic, sodic-calcic</td>
<td>Actinolite, clinopyroxene (diopside), chlorite, epidote, scapolite</td>
</tr>
<tr>
<td></td>
<td>Phyllic, sericitic</td>
<td>Sericite (muscovite-illite), chlorite, anhydrite</td>
</tr>
<tr>
<td></td>
<td>Intermediate argillic, sericite-chlorite-clay (SCC), argillic</td>
<td>Sericite (illite-smectite), chlorite, kaolinite (dickite), montmorillonite, calcite, epidote</td>
</tr>
<tr>
<td></td>
<td>Advanced argillic</td>
<td>Pyrophyllite, sericite, diasporite, alunite, topaz, tourmaline, dumortierite, zenyite</td>
</tr>
<tr>
<td></td>
<td>Greisen</td>
<td>Topaz, muscovite, tourmaline</td>
</tr>
<tr>
<td></td>
<td>Skam</td>
<td>Clinopyroxene, wollastonite, actinolite-tremolite, vesuvianite, epidote, serpentinite-talc, calcite, chlorite, illite-smectite, nontronite</td>
</tr>
<tr>
<td></td>
<td>Propylitic</td>
<td>Chlorite, epidote, calcite, actinolite, sericite, clay</td>
</tr>
<tr>
<td>High-sulfidation epithermal</td>
<td>Advanced argillic—acid sulphate</td>
<td>Kaolinite, dickite, alunite, diasporite, pyrophyllite, zenyite</td>
</tr>
<tr>
<td></td>
<td>Argillic, intermediate argillic</td>
<td>Kaolinite, dickite, montmorillonite, illite-smectite</td>
</tr>
<tr>
<td></td>
<td>Propylitic</td>
<td>Calcite, chlorite, epidote, sericite, clay</td>
</tr>
<tr>
<td>Low-sulfidation epithermal</td>
<td>“Adularia” — sericite, sericitic, argillic</td>
<td>Sericite, illite-smectite, kaolinite, chlorite, chlorite, montmorillonite, calcite, dolomite</td>
</tr>
<tr>
<td></td>
<td>Advanced argillic—acid-sulphate (steam-heated)</td>
<td>Kaolinite, alunite, cristobalite (opal, montmorillonite, calcite, dolomite)</td>
</tr>
<tr>
<td></td>
<td>Propylitic, zeolitic</td>
<td>Calcite, epidote, wairakite, chlorite, illite-smectite, montmorillonite</td>
</tr>
<tr>
<td>Mesothermal</td>
<td>Carbonate</td>
<td>Calcite, ankerite, dolomite, muscovite (Cr-V-rich), chlorite</td>
</tr>
<tr>
<td></td>
<td>Chlortic</td>
<td>Chlorite, muscovite, actinolite</td>
</tr>
<tr>
<td></td>
<td>Biotitic</td>
<td>Biotite, chlorite</td>
</tr>
<tr>
<td>Sediment-hosted gold</td>
<td>Argillic</td>
<td>Kaolinite, dickite, illite</td>
</tr>
<tr>
<td>Volcanogenic massive sulfide</td>
<td>Sericitic</td>
<td>Sericite, chlorite, chloritoid</td>
</tr>
<tr>
<td></td>
<td>Chlortic</td>
<td>Chlorite, sericite, biotite</td>
</tr>
<tr>
<td></td>
<td>Carbonate</td>
<td>Dolomite, siderite, ankerite, calcite, sericite, chlorite</td>
</tr>
<tr>
<td>Sediment-hosted massive sulfide</td>
<td>Tourmalinite</td>
<td>Tourmaline, muscovite</td>
</tr>
<tr>
<td></td>
<td>Carbonate</td>
<td>Ankerite, siderite, calcite, muscovite</td>
</tr>
<tr>
<td></td>
<td>Sericitic</td>
<td>Sericite, chlorite</td>
</tr>
<tr>
<td></td>
<td>Albite</td>
<td>Chlorite, muscovite, biotite</td>
</tr>
</tbody>
</table>

Minerals are grouped by assemblages of alteration minerals, and keyed to commonly used terminology; Complete assemblages are in Thompson and Thompson (1996)
The SWIR method does not detect minerals per se, but the vibrational energy of the chemical bonds between atom and molecule in the octahedral layer of a compound. This gives more flexibility in detecting amorphous materials. Therefore, it is possible to detect Si-OH bonds and thus look at amorphous silicas, such as opaline phases.

SWIR is also very sensitive to water, which will be illustrated in a later section. This is particularly valuable when tracking fluid migration through a deposit.

The common molecules detected are H$_2$O, OH, CO$_3$, and NH$_4$. Selected sulfates are seen when an OH bond is present, as are phosphates (Figure 6). Table II lists most of the common infrared-active minerals. The groups most likely encountered, with selected mineral examples, include clays (illite, illite/smectite, kaolinite, smectites, dickite) and all pylosilicates, carbonates (calcite, dolomite, ankerite, siderite, malachite, azurite, cerrusite), micas (muscovite, biotite, phlogopite, paragonite), chlorites (Mg, Fe, Al), Sulfates (alunite, jarosite, gypsum, Fe-sulfates), tourmaline (schorl, elbaite, dravite, rubellite), amphiboles (tremolite, actinolite, hornblende), pyroxenes (diopside, augite, hypersthene), garnets (pyrope, almandine, grossularite), pyroxinoids (rhodonite, sodalite) selected silicates (talc, topaz, scapolite, pyrophyllite, zonyite, epidote, olivine, serpentes, dumortierite, idocrase) evaporates (sulfates, niter, anhydrite, gypsum, borax minerals), hydroxides (diaspore, brucite), zeolites (stiblite, heulandite, clinoptilolite), and silica (opal, quartz, sinter)

Limits of detection will vary with the matrix composition and the active versus inactive components. It varies between mineral mixtures. It is very difficult to do accurate quantitative mixtures without first building calibration files of percentage mixtures.
SPECTRAL INFORMATION

Field Spectrometers, whether they are based on an integrating sphere or fiber optic cable, allow the collection and presentation of spectral data. Although the primary objective of this operation is the identification of the minerals found in different alteration systems, there is other information that can be obtained from the data and which is of great significance in the interpretation of the spectral information. These include chemical substitution, crystallinity, effects of water, paragenesis and temperature.

CHEMISTRY

The ability to discern compositional differences in mineral groups that have solid solution substitution is very useful. Some examples are the carbonates, amphiboles, chlorites, illite/micas, tourmalines, smectite clays, alunites, all of which exhibit wavelength shifts with cation substitution (Figure 7A, 7B, and 7C).

Figure 7 – The spectral plots in [A] are for alunites of different compositions from Ca to Na to K to NH₄. Note that the diagnostic feature ranges from 1495nm for Ca to 1460nm for NH₄. In Figure 7B, the diagnostic wavelength shift occurs in the 2330 nm range for Mg-chlorites shifting to 2350+ for Fe-chlorites. The carbonate group is shown in [C]. These mineral show a significant shift with different cation compositions ranging from (top to bottom) aragonite (Ca), Dolomite (Mg, Ca), calcite (Ca), rhodocrosite (Mg), stronianite (Sr), magnesite (Mg), cerrusite (Pb), siderite (Fe), ankerite (fe, Mg), malachite (Cu), and azurite (Cu).
IRON OXYHYDROXIDES AND SULFATES

Iron oxides, hydroxides, oxyhydroxides and sulfates are very important not only to precious metals exploration, but also in the determination of oxidation states and acid drainage creation potential. They can be used in the field to provide approximate pH's in an environment, especially along a stream, as shown in the cartoon below. The three most common iron minerals are plotted in Figure 9.

Figure 8 - This cartoon shows pH changes along a stream from a sulfide metals source through neutralization of acid waters by dilution and volume effects.

Figure 9 - The three most common iron minerals encountered in gold and copper deposits are jarosite, goethite and hematite. The plot shows spectral profiles and wavelengths for these minerals in the visible range, where most of their diagnostic features occur. The emission features are more consistent and reproducible than the absorption features. These features all have a range. These minerals are usually mixtures of each other and therefore a wavelength range is more applicable to defining them.
**IRON OXIDES OXYHYDROXIDES**

![Graph](image_url)

**Figure 10** - For most of these iron species, the definitive features are found in the visible region. The more common iron oxides and hydroxides are plotted in Figure 10A and include lepidocrocite, ferrhydrite, maghemite, goethite, and hematite.

The iron sulfates are very useful for helping determine pH of their environment and potential acid drainage factors. They are plotted in Figures 10B for SWIR range, and include schwertmannite, jarosite, copiapite, coquimbite, and melanterite. Except for jarosite, they show mostly water features in the SWIR. A full range spectrometer is therefore essential for their definitive identification.

**ACID IRON SULFATES**

![Graph](image_url)
Another perspective of this shift is the aluminum substitution in illites and Al-micas.

Figure 11- Aluminum content of illites can be estimated from the 2.2 µm absorption feature, which shifts relative to the percent aluminum present. There appears to be a deposit-specific correlation in that when illite/sericite/muscovite alteration is present, there are higher amounts of aluminum apparently associated with the ore zones. This also has been documented by Post and Noble (1993) and their data is plotted below against spectral wavelength values collected from their published samples.

Figure 12- Feature positions for a cross section of muscovites and illites from SPECMIN™ range from 2198nm to 2212 nm, with the majority falling within 2200-2204 nm. The illites plotted in Figure 9 are from different environments and from top to bottom are [A] Hog Ranch, Nevada, epithermal gold deposit; [B] Chuquicamata, Chile, porphyry copper; [C] Leadville, Colorado, gold vein system; [D] Cananea, Mexico, porphyry copper deposit; [E] Round Mountain, Nevada, disseminated gold deposit; [F, G, H] sedimentary illites from Illinois shales.

Degree of crystallinity can be determined from change in the profile, using a calibration data set for comparison (Figure 13).

Note how the kaolinite doublet in the 2200 nm region softens and broadens from a sharp, well-defined minimum to a shoulder as ordering decreases.

This change in profile can be mathematically defined and utilized as an index. It is a function of random and sporadic filling of metal sites in the octahedral layer of the mineral. Empty and random sites decrease order and, therefore, the sharpness of the profile.
PARAGENESIS AND WATER

Profiles also give information related to paragenesis and, indirectly, temperature (Figure 15). Broad, open absorption features coupled with large water features such as those found in smectites, zeolites, and opaline silica minerals indicate ambient to low temperatures from low energy environments. As the profile sharpens and becomes well defined, such as in muscovites, pyrophyllite, epidote, and dickite, the temperature of formation increases and the crystal lattice becomes ordered in a high-energy environment.

Figure 16 - plots different water sites from channel water in beryl to molecular water in gypsum to zeolite channel water to interlayer water in smectite to surface water.
Type of water present also can be tracked relative to profile and wavelength (Figure 16). As seen in the figure, different types of water (channel, structural, interlayer, surface, silica, encapsulated), have different profiles which provide information about how the water was formed and where it resides.

II. DATA COLLECTION AND ANALYSIS  Applications Examples
Outcrop mapping  -  Pamel

*Pamel:* An alteration study of the Pamel prospect in the Western Cordillera of Peru was conducted for Candente Resource Corp. The results from approximately 12 samples were integrated with geologic information and outcrop patterns. Based on 3 days work, combined with the previous geologic data, an alteration map was created for the property. The results clearly showed distinct alteration zones and helped delineate zones of interest (Figure 17). The alteration varies, from silicification to alunite-dickite, to alunite-kaolinite, to kaolinite dominant, and outward to sericite, illite, and chlorite. Small amounts of diaspore, topaz, and tourmaline were also noted locally (Thompson et al., 1999).

![Alteration Map - Candente Resources Corp.](image)

*Figure 17. - alteration map created by A. Thompson from PIMA outcrop samples.*
Core Logging

Spectral core logging is one of the major applications of reflectance spectroscopy. Figure 18 presents a spectral core log.

![Figure 18 - An Excel chart that shows mineralogy against alteration type against gold assay values. Plotting the minerals in this way gives an excellent comparison of the presence of gold related to mineralogy. The alteration classifications are indigenous to the high sulfidation system and represent mineral associations.](image)

Pit Bench Mapping - Blast Holes

Considerable information can be gained through the analysis of blast holes in an open pit mine. Although blast holes are assayed, in-situ mineralogical analysis can provide alteration types, which if spectrally defined, will give the metallurgist information needed for efficient mill recovery or heap leach pad blending.

![Figure 19 - This plot is a compilation of different alteration types from a copper mine, defined using spectroscopy. These include [A] muscovite, [B] weathered muscovite shown by large water feature, [C] illite with very minor kaolinite overprint, [D] illite - kaolinite mixture, [E] kaolinite – illite mixture, [F] kaolinite. By matching metallurgical data to the clay type, it was possible to determine that too much kaolinite would cause recovery losses.](image)
ALTERATION SYSTEMS FROM THE SPECTRAL PERSPECTIVE

The more common deposits that can be defined by reflectance spectroscopy include epithermal gold, low and high sulfidation systems; porphyries, kimberlites, IOCG (Iron Oxide-Copper-Gold), skarns, and uranium (both sandstone hosted sediment and unconformity types). Although there are numerous other deposit and alteration types that can be analyzed successfully with this method, space constraints allow only an overview of the above list.

This section therefore presents a very brief definition of the deposit type, reference to a graphic model, list the better-known deposit examples, and discuss the alteration from the spectral perspective. The information presented here as background to the deposit type and alteration systems has been compiled from references and from websites.

EPITHERMAL GOLD - LOW SULFIDATION

Definition of deposit type

Epithermal gold deposits form in hydrothermal systems related to volcanic activity. These systems, while active, discharge to the surface as hot springs or fumaroles. Epithermal gold deposits occur largely in volcano-plutonic arcs (island arcs as well as continental arcs) associated with subduction zones, with ages similar to those of volcanism. The deposits form at shallow depth, <1 km, and are hosted mainly by volcanic rocks.

Quartz veins, stockworks, and breccias carrying gold, silver, electrum, argentite, and pyrite with lesser and variable amounts of sphalerite, chalcopyrite, galena, rare tetrahedrite, and sulphosalt minerals form in high-level (epizonal) to near-surface environments. The ore commonly exhibits open-space filling textures and is associated with volcanic-related hydrothermal to geothermal systems.

There are two end-member styles of epithermal gold deposits, high sulfidation (HS) and low sulfidation (LS). The two deposit styles form from fluids of distinctly different chemical composition in contrasting volcanic environments. The fluid responsible for formation of LS ore veins is similar to waters tapped by drilling beneath hot springs in geothermal systems, waters that are reduced and having neutral-pH.

Many hydrothermal minerals are stable over limited temperature and/or pH ranges. Thus, mapping the distribution of alteration minerals in areas of epithermal prospects allows the thermal and geochemical zonation to be reconstructed, leading to a model of the hydrology of the extinct hydrothermal system. Alteration minerals are also crucial to distinguish the style of deposit, LS or HS.
Boiling of liquid in the LS geothermal environment leads to precipitation of gold in veins accompanied by a variety of features such as adularia and bladed calcite cementing colloform and brecciated quartz. Silica sinters may be the surface expression of such veins and may be accompanied by nearby zones of surficial steam-heated acid alteration.

These deposits form in both subaerial, predominantly felsic, volcanic fields in extensional and strike-slip structural regimes and island arc or continental andesitic stratovolcanoes above active subduction zones. Near-surface hydrothermal systems, ranging from hot springs at surface to deeper, structurally and permeability focused fluid flow zones are the sites of mineralization. The ore fluids are relatively dilute and cool solutions that are mixtures of magmatic and meteoric fluids. Mineral deposition takes place as the solutions undergo cooling and degassing by fluid mixing, boiling, and decompression.


Model for Low Sulfidation Epithermal Vein Systems

Figure 20 - Figure shows a model for a low-sulfidation epithermal gold system. Note distribution of alteration zoning and know deposits. (Corbett and Leach, 1998)
Major Global Deposits

EL PEÑON, Chile  MARTA, Peru
ESQUEL, Argentina ROUND MOUNTAIN, Nevada
CERRO VANGUARDIA, Argentina COMSTOCK, Virginia City, Nevada
HISHIKARI, Japan SLEEPER, Nevada
GOSAWONG, Indonesia MIDAS, Nevada
KUPOL, Russia WAIHI, New Zealand
ROSIA MONTANA, Romania GOLDEN CROSS, New Zealand
LIHIR, PNG CERRO BAYO, Chile
TRES CRUCES, Peru KORI KOLLO, Bolivia

COMMON ALTERATION MINERALS in LOW SULFDATION SYSTEMS

ILLITE  KAOLINITE  CHLORITES
ILLITE/SMECTITE BUDDINGTONITE EPIDOTE
Montmorillonite ADULARIA* ZEOLITES
QUARTZ CALCITE HEMATITE

*Adularia is not infrared active.

Alteration types associated with Low Sulfidation Systems

Silicification is extensive in ores as multiple generations of quartz and chalcedony and commonly is accompanied by adularia and calcite. Pervasive silicification in vein envelopes is flanked by sericite-illite-kaolinite assemblages. Intermediate argillic alteration [kaolinite-illite-montmorillonite (smectite)] forms adjacent to some veins; advanced argillic alteration (kaolinite-alunite) may form along the tops of mineralized zones. Propylitic alteration dominates at depth and peripherally. Panteleyev, A.(1996). Buddingtonite can also be present.

Figure 21 - Plot shows illite, mixed layer illite/smectite, montmorillonite, quartz, calcite, dolomite, kaolinite and pyrite.

Figure 22 - Silicic alteration is pervasive replacement of the rock by silica minerals. Minerals include opal, chalcedony, quartz, hematite, and pyrite.
This assemblage also contains adularia. The chemical bonds for feldspars are not detectable in the SWIR range. This varies from alteration around veins, fractures and permeable zones to selective replacement of plagioclase in alteration envelopes.

**Figure 23 - QSA alteration plot shows quartz, illite, muscovite with quartz, and pyrite.**

**Argillic**

This type of alteration occurs as wall rock alteration around veins and replacement zones in permeable lithologies. This may show the change in aluminum content and temperature change away from the vein in a progression from illite → illite/smectite → smectite. Carbonates are also important in these systems and may reflect condensation of CO₂ from deeper boiling zones.

In these systems is also seen amethyst, quartz pseudomorphs after calcite, barite, fluorite, Ca-Mg-Mn-Fe carbonate minerals such as rhodochrosite. Panteleyev (1996):
Steam Heated Argillic
This alteration type is found above the water table and forms an extensive area. It is related to the condensation and oxidation of gases (H$_2$S). It is found in geothermal areas with mud pots, fumaroles and native sulfur.

Weathered outcrops are often characterized by resistant quartz ± alunite 'ledges' and extensive flanking bleached, clay-altered zones with supergene alunite, jarosite, and other limonite minerals. Panteleyev (1996):

Propylitic
Propylitic alteration occurs in the outer zone of alteration around a deposit. It can have regional extension.
HIGH SULFIDATION SYSTEMS

Definition of deposit type

High sulphidation Au + Cu ore systems develop from the reaction with host rocks of hot acidic magmatic fluids to produce characteristic zoned alteration and later sulphide and Au + Cu + Ag deposition. Ore systems display permeability controls governed by lithology, structure and breccias and changes in wall rock alteration and ore mineralogy with depth of formation. One of the challenges is to distinguish the mineralized systems from a group of generally non-economic acidic alteration styles including lithocaps or barren shoulders, steam heated, magmatic solfatara and acid sulphate alteration.

Original definitions (Heald et al., 1987) suggest that epithermal deposits formed at temperatures < 300°C and therefore at shallow depths, typically < 1 km.

The term epithermal is therefore used in field exploration studies to describe Au ± Ag ± Cu deposits formed in magmatic arc environments (including rifts) at shallow depths, most typically above the level of formation of porphyry Cu-Au deposits (typically < 1 km), although in many instances associated with subvolcanic intrusions. Higher level epithermal deposits are commonly formed later in a deposit paragenesis, and may be telescoped upon deeper earlier formed deposits, including in some instances porphyry systems which are older, or part of the same overall magmatic event (e.g., Lihir, Corbett et al., 2001b). In strongly dilational structural environments porphyry-related systems may be telescoped outwards into the deeper “epithermal” environment (e.g., Cadia, Maricunga Belt).

In brief, high sulphidation deposits develop in settings where volatile rich magmatic fluids rise to higher crustal levels without significant interaction with the host rocks or ground waters. Volatiles (SO$_2$) evolved from the depressurising fluids oxidize to form a two stage hot acid fluid, the initial stage of which reacts with the host rocks to produce the characteristic zoned acidic alteration at epithermal crustal levels (Corbett and Leach, 1998). A later liquid-dominated fluid phase deposits sulphides which are characterised by pyrite with enargite, or the latter’s low temperature polymorph luzonite.

Magmatic rocks are interpreted to represent the ultimate source of ore fluids as demonstrated by the commonly association of HS deposits with felsic domes and phreatomagmatic breccias within flow-dome complexes. Many HS deposits display associations with porphyry Cu-Au systems of similar ages (Nena), or are collapsed upon and form part of the porphyry ores. Fluids responsible for the formation of HS deposits rarely evolve to Au-rich lower sulphidation style ores.
Model for High Sulfidation Gold Systems

Figure 27 - Model for High Sulfidation Gold System showing alteration zones and feeder structure. Source: Henley and Ellis, 1983

List of Well Known Deposits - Distribution

GOLDFIELD, Nevada  ZIJINSHAN, China
YANACOCHA, Peru  SIPAN, Peru
SUMMITVILLE, Colorado  TAMBO, Chile
LEPANTO, Phillipines  TANTAHUATAY, Peru
PASCUA-LLAMA, Chile-Argentina  AQUA RICA, Argentina
PIERINA, Peru  QUIMSACOCHA, Ecuador
PUEBLO VIEJO, Dominican Republic  MARTABE, Indonesia
ALTA CHACAMA< Peru  VELADERO, Argentina
LA COIPA, Chile  RODIQUILAR, Spain
MULATOS, Mexico  FURTEI, Sardinia

ALTERATION TYPES  (Corbett 2000)

Acid alteration contains a suite of minerals formed in low-pH conditions (e.g., kaolin-dickite, alunite, pyrophyllite, dickite) and may be grouped as styles including: Barren high sulphidation alteration described as lithocaps or barren shoulders are a common source of difficulty for explorationists, especially as these may crop out in the vicinity of low and high-sulphidation epithermal Au and also porphyry Cu-Au mineralization. Barren shoulders comprise a class of non-mineralized advanced argillic (silica-alunite ± pyrophyllite etc.) alteration derived
from magmatic volatiles venting from intrusive sources at depth (Corbett and Leach, 1998). These alteration zones are characterized by high temperature minerals such as pyrophyllite-diaspore, zunyite, topaz, locally including corundum and andalusite. Silica characteristically occurs as a massive form rather than vuggy silica, typical of mineralized high sulphidation deposits. Examples include Lookout Rocks, Horse Ivaal at Frieda River, Vuda, and Alum Mountain. Lithocaps (Sillitoe (1995, 1999) may extend to higher crustal levels, include additional lower temperature argillic alteration, and are interpreted (Sillitoe, op cit) to locally conceal epithermal or porphyry mineralization.

Shoulders and lithocaps crop out as variably dipping bodies termed ledges developed by the exploitation of permeable lithologies (typically flat dipping), or dilatant structures (typically steep dipping). These alteration zones are distinguished from mineralized high sulphidation systems in the field by the lack of vuggy silica and generally coarse grained and higher temperature layered silicates (e.g., alunite, pyrophyllite, dickite). The presence of high temperature alteration minerals such as alunite or corundum is distinctive. PIMA, Terraspec and XRD studies are therefore useful for the delineation of alteration zonation and provision of vectors towards where mineralization should occur in ore systems.
Alteration Types Associated with High Sulfidation Systems

Vuggy Silica

This alteration type occurs in structural zones or as replacement bodies in permeable lithologies, usually in the core of zones of advanced argillic alteration. This extreme form of acid leaching occurs in the higher levels of epithermal deposits. It is also found in the upper parts of telescoped porphyry systems.

Silicic

Silicification adds silica to a vuggy, acid leached rock and fills the vugs created through intense leaching. Silicification is common in high sulfidation systems. It can be confused with intense quartz stockwork at the top of some porphyry deposits.

Advanced Argillic

This alteration type is found in the upper regions of a high sulfidation system. It can occur in the lithocap of a porphyry system.

The progressive neutralization and cooling of high sulphidation fluids by rock reaction produces alteration moving away from the core in which zonation is characterized progressively outwards by mineral assemblages dominated by: alunite, pyrophyllite, kaolin, and illitic and chloritic clays (Corbett and Leach,
Mineralogies dominated by pyrophyllite-diaspore-dickite may be indicative, but not exclusively, of higher temperature (deeper) conditions, whereas lower temperature (opaline) pervasive silicification or alunite-kaolin dominate in cooler (higher level) settings. Similarly, thin alteration zones may suggest that fluid conditions have changed rapidly, possibly in a quenched higher level system (or distal to the fluid upflow), whereas wide zonation characterizes slower changing fluid conditions more typical of deeper levels (or more proximal to the fluid upflow).
Argilic — Intermediate Argillic

This alteration zone is found between propylitic and and advanced argillic. It can be progressive outward from illite → illite/smectite → montmorillonite. It may also show gradations from illite → ± illite/smectite into a chlorite envelope around the illite vein (Comstock Lode, Virginia City, Nv).

![Graphs showing mineral alteration](image)

**Figure 32** - Argillic alteration contains kaolinite, illite, illite/smectite, montmorillonite, quartz, and pyrite.

**Figure 33** - Propylitic Alteration shows Mg-Chlorite, Fe-chlorite, epidote, calcite, pyrite.

Propylitic Alteration

Propylitic can be regional, but it is usually found in the outer-most alteration zone. It will show different wavelengths for the chlorites as opposed to those chlorites in the intermediate argillic suite.
Porphyry Copper Deposits

Porphyry copper deposits form at 2-5 km below the earth’s surface at temperatures of >500-600 degree C from hypersaline (40-60 wt % NaCl equiv.) near neutral magmatic fluids (Hedenquist and Lowenstein, 1994). Aqueous magmatic vapors containing CO₂, SO₂, H₂S, and HCl separate from fluids and ascend to the surface. Ore bodies range in size from several hundred meters to more than a kilometer in diameter. Grades tend to be low, and the ore minerals are usually disseminated through a quartz stockwork within the upper part of the host porphyry intrusive.

Alteration around the ore body is usually concentric and may be described by the Lowell and Guilbert (1970) model for porphyry copper deposits. Zoned assemblages that were formed as reactive, metal-bearing magmatic fluid, which moved away from the intrusion, cooled and reacted with the country rock.

In the core of the deposit, over the mineralizing intrusion, there is a potassic zone (biotite + magnetite ± orthoclase ± albite ± anhydrite ± chalcopyrite ± bornite) which represents late magmatic K-metasomatism of the host intrusives. This passes upwards into extensive quartz stockwork and a transitional zone characterized by intermediate argillic alteration (chlorite + hematite/sulfides + quartz ± illite).

This is usually overprinted by the phyllic zone which dominates the stockwork. The phyllic zone is acid-leached and characterized by quartz + sericite + tourmaline + pyrite ± chalcopyrite. The sericite may be composed of either illite + montmorillonite or fine grained muscovite. Above the phyllic zone, the rock is capped by an advanced argillic zone where leaching is oxidized acidic water (generated by the absorption of magmatic vapor in to meteoric waters) has been particularly intense and may reach to the paleosurface. It is characterized by the assemblage quartz + alunite ± pyrophyllite ± kaolinite/dickite. Surrounding the potassic, intermediate/phyllic and advanced argillic zones is the deuterically (propylitic) altered country rock. The propylitic zone is characterized by green coloration and the assemblage: chlorite + epidote ± albite ± actinolite ± calcite ± hematite.

Hypogene copper sulphide mineralization is characterized by chalcopyrite [although bornite occurs when the sulfur activity is low as indicated by the absence of pyrite]. Magnetite is common in the potassic zone and pyrite in phyllic zone in the advanced argillic zone, sulphides have been decomposed and leached, or replaced by limonite.

Supergene leaching of hypogene alteration and mineralization occurs after the hydrothermal event. It is distinguished here from weak secondary alteration (due to later near neutral meteoric water percolation) which does not produce strong leaching, but may result in a modest overprint of montmorillonite, kaolinite, in the
hypogene and supergene zones. The principal mechanism of strong supergene leaching is the oxidation of pyrite in water, which produces limonite and generates sulphuric acid. The effect of this acid on acid stable hypogene minerals (quartz, alunite, kaolinite) is subtle. However, there may be decomposition of biotite and feldspars in the potassic zone. Feldspars and chlorite in the propylitic zone and illite in the phyllic zone, producing a supergene leached cap that is characterized by the assemblage quartz + alunite + kaolinite + gypsum. This is not always easy to distinguish from the hypogene leached cap of the advanced argillic zone.

![Figure 34](image)

**Infrared Active Alteration Minerals Associated with Porphyries**

Infrared-active alteration minerals associated with porphyries include quartz, sericite/muscovite, biotite, phlogopite, anhydrite/gypsum, magnetite, actinolite, chlorite, epidote, calcite, clay minerals (illite, kaolinite, smectite), tourmaline. Gangue minerals in mineralized veins are mainly quartz with lesser biotite, sericite, magnetite, chlorite, calcite, epidote, anhydrite and tourmaline. Leached cap and lithocap minerals can include jarosite, alunite, pyrophllite, diaspore, zunyite, topaz and dickite.
Alteration types associated with Porphyry Copper Deposits

ALTERATION MINERALOGY:
Early formed alteration can be overprinted by younger assemblages. Central and early formed potassic zones (K-feldspar and biotite) commonly coincide with ore. This alteration can be flanked in volcanic host rocks by biotite-rich rocks that grade outward into propylitic rocks. The biotite is a fine-grained, “shreddy” looking secondary mineral that is commonly referred to as an early developed biotite (EDB) or a “biotite hornfels”. These older alteration assemblages in cupriferous zones can be partially to completely overprinted by later biotite and K-feldspar and then phyllic (quartz-sericite-pyrite) alteration, less commonly argillic, and rarely, in the uppermost parts of some ore deposits, advanced argillic alteration (kaolinite-pyrophyllite) . Panteleyev (1995):

Potassic - biotite rich Alteration  (Thompson and Thompson, 1996)
This alteration type is found in the core of porphyry deposits. It may form large peripheral alteration zone in wall rocks (without K-spar) and zones out to propylitic alteration. Minerals include biotite, phlogopite, K-spar, magnetite, quartz, anhydrite, albite-sodic plagioclase, actinolite, rutile, apatite, sericite, chlorite, and epidote.

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Figure 35 - Potassic alteration shows Actinolite, biotite, phlogopite, epidote, iron-chlorite, Mg-chlorite, muscovite, quartz, anhydrite, magnetite.

Figure 36 - Potassic - K-silicate alteration includes "albite", anhydrite, quartz, muscovite and epidote.
Potassic - K-silicate

This alteration occurs in the core of porphyry systems, particularly when hosted by felsic intrusions - granodiorite, quartz monzonite, granite, seyenite. Minerals include K-feldspar, quartz, albite, muscovite, anhydrite, epidote. Feldspars do not have detectable bonds in the SWIR region and therefore their profiles do not have distinctive features.

Sodic, Sodic-Calcic

Occurs with minor mineralization in the deeper parts of some porphyry systems and is a host to mineralization in porphyry deposits associated with alkaline intrusions. Minerals include albite, actinolite, diopside, quartz, magnetite, titanite, chlorite, epidote, scapolite.

Figure 37 - Sodic-Calcic alteration plots
Albite, actinolite, Fe-chlorite, epidote, Mg-chlorite, marialite, mizzonite, diopside, quartz, magnetite.

Figure 38 - Phyllic alteration includes: illite, muscovite, iron chlorite, Mg-chlorite, gypsum, anhydrite, quartz, pyrite.
**Phyllic**

Phyllic alteration commonly forms a peripheral halo around the core of porphyry deposits. It may overprint earlier potassic alteration and may host substantial mineralization. Detectable minerals include muscovite, illite, quartz, pyrite, chlorite, hematite, and anhydrite.

**Intermediate Argillic**

Intermediate argillic alteration generally forms a structurally controlled to widespread overprint on other types of alteration in many porphyry systems. The detectable minerals include illite, muscovite, dickite, kaolinite, Fe-chlorite, Mg-chlorite, epidote, montmorillonite, calcite, and pyrite.

![Figure 39 - Intermediate argillic alterations](image)

*Figure 39 - Intermediate argillic alteration contains muscovite, illite, chlorite, kaolinite, dickite, montmorillonite, calcite, epidote, and pyrite.*

![Figure 40 - Advanced argillic alteration](image)

*Figure 40 - Advanced argillic alteration for a porphyry system will contain higher temperature minerals such as Topaz, pyrophyllite, andalusite, dumortierite, tourmaline, along with alunite, diaspore, quartz, hematite, and pyrite.*
**Advanced Argillic**

This is an intense alteration phase, often in the upper part of porphyry systems. It can also form envelopes around pyrite-rich veins that cross cut other alteration types. The minerals associated with this type are higher temperature and include pyrophyllite, quartz, andalusite, diaspore, corundum, alunite, topaz, tourmaline, dumortierite, pyrite, and hematite.

**Propylitic**

Forms the outermost alteration zone at intermediate to deep levels in the porphyry system. It can be mineralogically zoned from inner actinolite-rich to outer epidote-rich facies. Chlorite, epidote, albite, calcite, actinolite, illite, clay pyrite

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**Figure 41** - Propylitic alteration for porphyry systems includes; Fe-chlorite, Mg-chlorite, epidote, actinolite, calcite, illite, montmorillonite, albite, and pyrite.

**Figure 42** - Copper carbonates in porphyry systems can include azurite, malachite, aurichalcite, and rosasite.
Supergene
Secondary (supergene) zones carry chalcocite, covellite and other Cu₂S minerals (digenite, djurleite, etc.), chrysocolla, native copper and copper oxide, carbonate and sulphate minerals. Oxidized and leached zones at surface are marked by ferruginous “cappings” with supergene clay minerals, limonite (goethite, hematite and jarosite) and residual quartz. Panteleyev, A. (1995):

Minerals include: antlerite, bisbeeite, brochantite, chalcoalumite, claringbullite, conichalcite, crednerite, cuprite, delafossite, graemite, nantokite, paramelaconite, paratacamite, rosasite and spangolite, aurichalcite, connellite, cyanotrichite, malachite and shattuckite are additional supergene copper. Those available in the database are plotted in this section.

The best way to display the variations in the supergene environment is by group. The following plots are divided into copper carbonates, chlorides, arsenates, phosphate, silicates and sulfates.

**Figure 43** - Copper chlorides include atacamite, connellite, and cumengite.

**Figure 44** - Copper phosphates include libelhenite, pseudomalachite, and turquoise.

**Figure 45** - Copper arsenates include bayldonite, chenevixite, conichalcite, clinoclase and olivenite.
SILICATES

SULFATES

Figure 46 - Copper silicates include chrysocolla, dioptase, papagoite, and shattuckite

Figure 47 - Copper sulfates include antlerite, brochantite, chalcocite, cyanotrichite, kroehnite, spangolite.
LEACHED CAP

Minerals in the leached cap environment include alunite, kaolinite, illite, diaspore, iron oxides, copper sulfates, and copper hydroxides.

Exotic copper mineralization is a complex hydrochemical process linking supergene enrichment, lateral copper transport, and precipitation of copper oxide minerals in the drainage network of a porphyry copper deposit.
DIAMONDS - KIMBERLITES

DEPOSIT DEFINITION
Kimberlites are a group of volatile-rich (dominantly carbon dioxide) potassic ultrabasic rocks. Commonly, they exhibit a distinctive inequigranular texture resulting from the presence of macrocrysts (and in some instances megacrysts) set in a fine-grained matrix. The megacryst/macrocryst assemblage consists of rounded anhedral crystals of magnesian ilmenite, Cr-poor titanian pyrope, olivine, Cr-poor clinopyroxene, phlogopite, enstatite and Ti-poor chromite. Olivine is the dominant member of the macrocryst assemblage. The matrix minerals may include second generation euhedral primary olivine and/or phlogopite, together with perovskite, spinel (titaniferous magnesian aluminous chromite, titanian chromite, members of the magnesian ulvospinel-ulvospinel-magnetite series), diopside (Al- and Ti-poor), monticellite, apatite, calcite, and primary late-stage serpentine (commonly Fe rich). Some kimberlites contain late-stage phlogopites. The replacement of early-formed olivine, phlogopite, monticellite, and apatite by deuteric serpentine and calcite is common. Evolved members of the clan may be devoid of, or poor in, macrocrysts, and composed essentially of calcite, serpentine, and magnetite together with minor phlogopite, apatite, and perovskite. (From Mitchell 1986)

http://www.geology.ubc.ca/research/diamonds/kopylova/intro/morphology.html

Kimberlites can be divided into 3 distinct units based on morphology and petrology. These units are: 1) Crater Facies Kimberlite, 2) Diatreme Facies Kimberlite; 3) Hypabyssal Facies Kimberlite (Mitchell 1986). Kimberlite occurs in the Earth’s crust in vertical structures known as kimberlite pipes.

Kimberlites are divided into Group I (basaltic) and Group II (micaceous) kimberlites, along mineralogical grounds, the difference in mineralogy being caused by the preponderance of water versus carbon dioxide. Group I kimberlites are of CO₂-rich ultramafic potassic igneous rocks dominated by a primary mineral assemblage of forsteritic olivine, magnesian ilmenite, chromian pyrope, almandine-pyrope, chromian diopside, phlogopite, enstatite and of Ti-poor chromite. Group-II kimberlites (or orangeites) are ultrapotassic, peralkaline rocks rich in volatiles (dominantly H₂O). The distinctive characteristic of orangeites is phlogopite macrocrysts and microphenocrysts, together with groundmass micas that vary in composition from phlogopite to Fe-rich phlogopite.
Diamonds are mined in about 25 countries. India was the initial source, followed by Brazil and then South Africa. Africa is the richest continent for diamond mining, accounting for roughly 49% of world production. The major sources are in the south with lesser concentrations in the west-central part of the continent. The major producing countries are Congo Republic (Zaire), Botswana, South Africa, Angola, Namibia, Ghana, Central African Republic, Guinea, Mali, Sierra Leone, Tanzania and Zimbabwe. Canada is starting to come on line now with Ekati and Diavik in the northern part of the country. Australia has one producing mine, the Argyle. There are major resources in Russia. The United States has very minor deposits in Colorado, Wyoming and Arkansas.

(http://www.amnh.org/exhibitions/diamonds/other.html)
ALTERATION MINERALS

Common alteration minerals associated with kimberlites include: Carbonates, Phlogopite, Biotite, chlorites, Serpentine, richterite, other amphiboles, talc, sepechiorlitie, smectite, saponite, apophyllite, zeolites (analcime), vermiculite.

INDICATOR MINERALS

Several minerals, when found in glacial sediments, are useful indicators of the presence of kimberlite, and to a certain extent, an evaluation of the diamond potential of kimberlite. These minerals are far more abundant in kimberlite than diamond, survive glacial transport, and are visually and chemically distinct. Cr-pyrope (purple colour, kelyphite rims), eclogitic garnet (orange-red), Cr-diopside (pale to emerald green), Mg-ilmenite (black, conchoidal fracture), chromite.
(reddish-black, irregular to octahedral crystal shape), and olivine (pale yellow-green) are the most commonly used kimberlite indicator minerals in drift prospecting. Kimberlite indicator minerals are recovered from the medium to very coarse sand-sized fraction of glacial sediments, and analyzed by electron microprobe to confirm their identification.

(source: http://sts.gsc.nrcan.gc.ca/page1/miner/kirk/indic.htm)

Garnets

- Pyrope, Almandine, Grossular

Diopside

- Chrome-diopside, diopside

Others to consider

- Zircon, Olivine, Orthopyroxenes (Harzburgite?), Chromite??, Ilmenite??, Perovskite??

Figure 53 - This plot includes almandine, grossularite, pyrope garnets; diopside, Cr-diopside, and olivine.

Figure 54 - Garnet comparisons

The indicator garnet for the presence of diamonds in kimberlites is a G-10. This has a diagnostic chrome feature in the visible, indicated by the arrow.
SKARNS

DEPOSIT TYPE DEFINITION

http://www.wsu.edu:8080/~meinert/aboutskarn.html#Definitions

There are many definitions and usages of the word "skarn". Skarns can form during regional or contact metamorphism and from a variety of metasomatic processes involving fluids of magmatic, metamorphic, meteoric, and/or marine origin. They are found adjacent to plutons, along faults and major shear zones, in shallow geothermal systems, on the bottom of the seafloor, and at lower crustal depths in deeply buried metamorphic terranes. What links these diverse environments, and what defines a rock as skarn, is the mineralogy. This mineralogy includes a wide variety of calc-silicate and associated minerals but usually is dominated by garnet and pyroxene.

The composition and texture of the protolith tend to control the composition and texture of the resulting skarn. In contrast, most economically important skarn deposits result from large-scale metasomatic transfer, where fluid composition controls the resulting skarn and ore mineralogy.

One of the more fundamental controls on skarn size, geometry, and style of alteration is the depth of formation.

Major skarn types:
Fe Skarns , Au Skarns, W Skarns, Cu Skarns, Zn Skarns, Mo Skarns.

The vast majority of skarn deposits are associated with magmatic arcs related to subduction beneath continental crust. Plutons range in composition from diorite to granite although differences among the main base metal skarn types appear to reflect the local geologic environment (depth of formation, structural and fluid pathways) more than fundamental differences of petrogenesis (Nakano et al., 1994). In contrast, gold skarns in this environment are associated with particularly reduced plutons that may represent a restricted petrologic history.
ALTERATION TYPES ASSOCIATED WITH SKARNS

RETROGRADE ALTERATION
The term retrograde alteration refers to an alteration stage in which higher temperature, generally anhydrous, minerals are replaced by lower temperature, generally hydrous, minerals. In some cases, this process is largely metamorphic, i.e., the bulk composition stays roughly the same but mineral "A" is replaced by mineral "B". In other cases, retrograde alteration is concurrent with strong metasomatism, such as during mineralization with various sulfide minerals.

Retrograde skarn mineralogy, in the form of epidote, amphibole, chlorite, and other hydrous phases, typically is structurally controlled and overprints the prograde zonation sequence. Thus, there often is a zone of abundant hydrous minerals along fault, stratigraphic, or intrusive contacts. This superposition of later phases can be difficult to discriminate from a spatial zonation sequence due to progressive reaction of a metasomatic fluid.

SKARN MODEL

Figure 55 - Zonation of most skarns reflects the geometry of the pluton contact and fluid flow. Such skarns are zoned from proximal endoskarn to proximal exoskarn, dominated by garnet. More distal skarn usually is more pyroxene-rich and the skarn front, especially in contact with marble, may be dominated by pyroxenoids or vesuvianite. Meinert, L.D., 1992
In general, retrograde alteration is more intense and more pervasive in shallower skarn systems. In some shallow, porphyry copper-related skarn systems, extensive retrograde alteration almost completely obliterates the prograde garnet and pyroxene skarn (Einaudi, 1982a,b)

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<th>Skarn</th>
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</table>

<table>
<thead>
<tr>
<th>Vein system</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>A vein: biotite vein</td>
<td></td>
<td>Dominant in shallower part of skarn system</td>
</tr>
<tr>
<td>B vein: Quartz-molybdenite (no alteration halo along vein)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C vein: chlorite (or green smectite bearing mixed layer clay) + sulfides (partly pyrite +molybdenite)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>D vein: phyllitic alteration (sericite alteration along Quartz vein)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Related intrusion</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Generally, porphyry, sometimes depend on depth of formation</td>
<td></td>
<td>Depend on depth of formation</td>
</tr>
</tbody>
</table>

Figure 56 - From Murakami (2005)

The main differences between amphiboles in different skarn types are variations in the amount of Fe, Mg, Mn, Ca, Al, Na, and K. Amphiboles from Au, W, and Sn skarns are progressively more aluminous (actinolite-hastingsite-hornblende), amphiboles from Cu, Mo, and Fe skarns are progressively more iron-rich in the tremolite-actinolite series, and amphiboles from zinc skarns are both Mn-rich and Ca-deficient, ranging from actinolite to dannemorite. For a specific skarn deposit or group of skarns, compositional variations in less common mineral phases, such as idocrase, bustamite, or olivine, may provide insight into zonation patterns or regional petrogenesis (e.g. Giere, 1986; Silva and Siriwardena, 1988; Benkerrou and Fonteilles, 1989).
PROGRADE ALTERATION
The minerals found in prograde systems include: Scapolite-meionite, scapolite-marialite, scapolite-mizzonite, diopside, hedenbergite, olivine, rhodonite, grossularite, andradite, wollastonite, vesuvianite.

Figure 57 - Scapolite-meionite, scapolite-marialite, scapolite-mizzonite, diopside, hedenbergite, olivine, rhodonite, grossularite, andradite, wollastonite, and vesuvianite.
RETROGRADE SKARN ALTERATION

Retrograde alteration commonly replaces earlier skarn alteration, but may also affect adjacent wallrock, which is many cases is limestone.

The minerals observed in these suites include illite/smectite, montmorillonite, nontronite, pyrite, epidote, clinozoisite, tremolite ->actinolite, prehnite, chlorite, vesuvianite, biotite, phlogopite, quartz, calcite, dolomite, ankerite, siderite, Fe-dolomite. These are plotted in Figures 58, 59, and 60.

CLAY SERIES

Figure 58 - Illite, Illite/smectite, montmorillonite, nontronite, and pyrite.

CARBONATE SERIES

Figure 59 - Ankerite, calcite, dolomite, Fe-dolomite, and siderite.
Magnesium skarns are developed as metasomatic replacements of dolomitic limestone. High temperature magnesium skarns are characterized by forsterite, and diopside and low temperature magnesium skarns contain serpentine and talc. Both of which occur as retrograde minerals after forsterite and clinopyroxene.

Figure 61 plots forsterite, serpentine, talc, calcite, tremolite, and magnetite.
IRON OXIDE COPPER GOLD (IOCG)

Definition of deposit type

Iron oxide copper-gold (IOCG) deposits encompass a wide spectrum of sulphide-deficient low-Ti magnetite and/or hematite ore bodies of hydrothermal origin where breccias, veins, disseminations and massive lenses with polymetallic enrichments (Cu, Au, Ag, U, REE, Bi, Co, Nb, P) are genetically associated with, but either proximal or distal to large-scale continental, A- to I-type magmatism, alkaline-carbonatite stocks, and crustal-scale fault zones and splay}s. The deposits are characterized by more than 20% iron oxides. Their lithological hosts and ages are non-diagnostic but their alteration zones are, with calcic-sodic regional alteration superimposed by focused potassic and iron oxide alterations. The deposits form at shallow to mid crustal levels in extensional, anorogenic or orogenic, continental settings such as intracratonic and intra-arc rifts, continental magmatic arcs and back-arc basins. Margins of Archean craton where arcs and successor arcs were developed appear to be particularly fertile.

http://gsc.nrcan.gc.ca/mindep/synth_dep/iocg/index_e.php

The 3810 Mt Olympic Dam deposit in Australia is the deposit that guides most mineral exploration worldwide (Hitzman et al., 1992; Western Mining Corporation, 2003). Currently known metallogenic IOCG districts occur in Precambrian shields worldwide as well as in Circum-Pacific regions (e.g., Carlon, 2000; Fourie, 2000; Porter, 2000, 2002a and papers therein; Strickland and Martyn, 2002; Gandhi, 2004c; Goff et al., 2004).

The recognition of this deposit type was triggered by the 1975 discovery of the giant Olympic Dam deposit in Australia (Roberts and Hudson, 1983) followed by the discovery of Starra (1980), La Candelaria (1987), Osborne (1988), Ernest-Henry (1991), and Alemao (1996). The early discoveries served to define this group of deposits as the IOCG deposit type in the 1990s (Hitzman et al., 1992).

Source: Mineral Deposits of Canada

Source: Iron Oxide Copper-Gold (+/-Ag,+/-Nb,+/-REE,+/-U) Deposits: A Canadian Perspective

Louise Corriveau  Geological Survey of Canada

http://gsc.nrcan.gc.ca/mindep/synth_dep/iocg/index_e.php

This appears to be a website publication only.
IOCG PROPOSED MODELS

Figure 62 - Schematic illustrations of flow paths and hydrothermal features for alternative models for IOCG deposits. Shading in arrows indicates predicted quartz precipitation (veining) for different paths in different quartz-saturated rocks which provides a useful first-order indication of path (cf., Barton el al., 1997, Barton and Johnson, 2000).
ALTERATION
Following is a summary of alteration zoning associated with many IOCG deposits. Please note that reflectance spectroscopy does not detect the silicate bonds found in the feldspars.

The principal ore minerals are hematite (specularite, botryoidal hematite and martite), low-Ti magnetite, bornite, chalcopyrite, chalcocite and pyrite. Subordinate ones include Ag-, Cu-, Ni-, Co-, U-arsenides, autunite, bastnaesite, bismuthinite, brannerite, britholite, carrolite, cobaltite, coffinite, covellite, digenite, florencite, loellingite, malachite, molybdenite, monazite, pitchblende, pyrrhotite, sulphosalts, uraninite, xenotime, native bismuth, copper, silver and gold, Ag-, Bi-, Co-telluride, and vermiculite.

Gangue mineralogy consists principally of albite, K-feldspar, sercite, carbonate, chlorite, quartz (crypto-crystalline in some cases), amphibole, pyroxene, biotite and apatite (F- or REE-rich) with accessory allanite, barite, epidote, fayallite, fluorite, ilvaite, garnet, monazite, perovskite, phlogopite, rutile, scapolite, titanite, and tourmaline. The amphibole includes Fe-, Cl-, Na- or Al-rich hornblende (edenite), actinolite, grunerite, hastingsite, and tschermakitic or alkali amphibole. Carbonates include calcite, ankerite, siderite and dolomite. Late-stage veins contain fluorite, barite, siderite, hematite and sulphides.

There are 3-6 alteration types associated with IOCG deposits. These include sodic, sodic-calcic, potassic, quartz-tourmaline, and albite-calcite. Figure 63 shows the minerals associated with the different alteration types.

<table>
<thead>
<tr>
<th>MINERAL</th>
<th>Sodic</th>
<th>Sodic-calcic</th>
<th>Potassic</th>
<th>Qzt-Tourmaline Albite-Calcite</th>
</tr>
</thead>
<tbody>
<tr>
<td>albite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K-spar</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>amphiboles</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>bastnaesite (REE carbonates)</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>biotite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>calcite</td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>chlorite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>epidote</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>garnet (andradite, Fe-rich garnet)</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>hematite</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>monazite</td>
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<tr>
<td>muscovite</td>
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</tr>
<tr>
<td>quartz</td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>scapolite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>tourmaline</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>magnetite</td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

Figure 63 – Minerals associated with alteration types for IOCG deposits.
**Sodic Alteration.** The minerals are albite, magnetite, and epidote.

**Sodic Calcic Alteration early stages**

There is strong albitization (± clinopyroxene, titanite, calc-silicate (clinopyroxene, amphibole, garnet) - alkali feldspar (K-feldspar, albite) ± Fe-Cu sulphides, and various assemblages including albite, actinolite, magnetite, apatite and late epidote. Ca-rich host-rocks are present, with Fe-rich garnet-clinopyroxene ± scapolite skarn.

Scapolite-actinolite-magnetite
Albite - actinolite-magnetite
Albite-chlorite-epidote-calcite

This is a very complex system or systems.

**Iron Stage**

This stage is largely sterile except for iron ore and apatite. There is an influx of magnetite ± biotite.
The minerals include **iron oxide** (magnetite and hematite), **iron carbonate** (siderite and ankerite) and **iron silicate** (chlorite and amphibole, in particular grunerite and hastingsite).

**Potassic Alteration**

The main ore stage is associated with potassium-iron alteration zones in which either sericite (at Olympic Dam and Prominent Hill) or K-feldspar (most common) prevail as the potassic-bearing phase.

K-feldspar-biotite-pyrite-chalcopyrite

At **Olympic Dam**, Potassic alteration occurs with hematite, sericite, chlorite, carbonate ± Fe-Cu sulphides ± uranium, pitchblende and REE (bastnaesite) minerals prevail. It is locally superimposed on a magnetite-biotite alteration (Reynolds, 2000; Skirrow et al., 2002). The intense chloritization destroys the biotite. Quartz, fluorite, barite, carbonate, orthoclase, epidote and tourmaline can also be present. Titanite, monazite, perovskite and rutile may occur. The main uranium mineral is uraninite (pitchblende) with lesser coffinite and brannerite.
Quartz-Tourmaline Alteration

The minerals in this suite include tourmaline-magnetite-quartz-chalcopyrite-pyrite, chlorite, and chalcopyrite.

Albite-Calcite Alteration

This phase is post ore - retrograde.

The minerals include albite-calcite-bornite-chalcocite and dolomite in quartz veining.

Figure 68 - Quartz-tourmaline alteration plot with magnetite, quartz, tourmaline, and Fe-chlorite.

Figure 69 - Albite-calcite alteration is post-ore and contains calcite, dolomite, and quartz.
UNCONFORMITY URANIUM DEPOSITS

Unconformity-associated uranium (U) deposits consist of pods, veins, and semi-massive replacements of uraninite which are overlain by zoned, quartz dissolution, kaolinite, sudoite, illite, dravite, quartz alteration halos and located near unconformities between Paleo- to Mesoproterozoic siliciclastic basins and metamorphic basement.

In major producing basins (e.g. Athabasca, Kombolgie) the 1-2 km, flat-lying, and apparently un-metamorphosed, mainly fluvial strata include red to pale tan quartzose conglomerate, arenite and mudstone deposited from ~1730 to <1540 Ma and pervasively altered. Below the basal unconformity, red hematitic and bleached clay-altered regolith grades down through chloritic altered to fresh basement gneiss. The highly metamorphosed, interleaved Archean to Paleoproterozoic granitoid and supracrustal basement gneiss includes graphitic metapelite that preferentially hosts shear zones and ore deposits. A variety of deposit shapes, sizes and compositions ranges from monometallic U and generally basement-hosted moderately dipping veins, to U-dominated polymetallic sub-horizontal uraninite lenses just above or straddling the unconformity, including variable Ni, Co, As, Pb >> Au, Pt, Cu, REE and Fe.


The paleo-weathered basement immediately below the Athabasca Group has a vertical profile ranging from a few centimeters up to 70 meters thick, with much deeper pockets and slivers developed along fault zones. The following description is after Macdonald (1980, 1985). The regolith is variably affected by diagenetic iron reduction resulting in a bleached zone at the top, interpreted as hematite removal from the upper "red zone" of the paleo-weathered basement section. This bleached zone, always present immediately below the unconformity, is composed of buff-colored clay and quartz. It crosscuts and therefore post-dates the red zone beneath. Below this, the upper portion of the regolith profile exhibits a strong red hematitic alteration that grades into underlying greenish chloritic alteration. In profiles developed on the basement meta-arkose, a white zone of white clay replacement of feldspars and mafic minerals separates the red and green zones. A downward progression from this kaolinite to illite and chlorite is common through the regolith profile.

Discussion continues regarding the respective contributions of paleo-weathering, diagenesis and hydrothermal processes to the alteration preserved at the unconformity. Cuney et al. (2003) stated that the regolith may represent a redox front controlled by the degree of percolation of the oxidized diagenetic brines in the basement, and that the lack of Ce-anomalies is not in favour of a lateritic
origin. It is here preferred that the red-green alteration was indeed regional, a result of lateritic weathering as proposed by Macdonald (1980), and was overprinted by hydrothermal alteration (the white zone of Macdonald) that increases in intensity close to and is genetically related to the bleaching alteration associated with U deposits (Macdonald, 1985). This interpretation is based on field relationships noted by Macdonald (the white zone transects down across the red and green along fractures).

**Deposit Distribution**

The new uranium production is likely to come from deposits in Canada, Australia, Niger, the United States, and Kazakhstan.

**Table III - Global Distribution of Uranium Deposits**

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Location</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olympic Dam</td>
<td>Australia</td>
<td>breccia</td>
</tr>
<tr>
<td>Jabiluka</td>
<td>Australia</td>
<td>unconformity</td>
</tr>
<tr>
<td>Ranger</td>
<td>Australia</td>
<td>unconformity</td>
</tr>
<tr>
<td>Kintyre</td>
<td>Australia</td>
<td></td>
</tr>
<tr>
<td>Athabasca, Thelon</td>
<td>Canada</td>
<td>unconformity</td>
</tr>
<tr>
<td></td>
<td>Kazakhstan</td>
<td>sandstone</td>
</tr>
<tr>
<td></td>
<td>Niger,</td>
<td>sandstone</td>
</tr>
<tr>
<td></td>
<td>Uzbekistan,</td>
<td>sandstone</td>
</tr>
<tr>
<td>Franceville Basin</td>
<td>Gabon</td>
<td>sandstone</td>
</tr>
<tr>
<td>Karoo Basin)</td>
<td>South Africa</td>
<td>sandstone</td>
</tr>
<tr>
<td>Wyoming, New Mexico</td>
<td>USA</td>
<td>Sandstone</td>
</tr>
</tbody>
</table>

**Alteration mineralogy and geochemistry**

Alteration mineralogy and geochemistry of Australian and Canadian unconformity associated deposits and their host rocks in the Athabasca and Thelon basins, and the Kombolgie Basin have been compared by Miller and LeCheminant (1985), Kotzer and Kyser (1995), Kyser et al. (2000) and Cuney et al. (2003).

Early work on alteration mineralogy in the Athabasca Basin is exemplified by Hoeve and Quirt (1984), and Wasilyuk (2002) has set the modern framework of exploration clay mineralogy. Intense clay alteration zones surrounding deposits such as Cigar Lake constitute natural geological barriers to U migration in ground waters (Percival et al., 1993) and are important geotechnical factors in mining and ore processing (Andrade, 2002) Figure 70.

Alteration halos are developed mainly in the siliciclastic strata and range between two distinctive end member types: 1) quartz dissolution + illite (NE Athabasca)
Ore-related illitic alteration is reflected in anomalously high illite proportions (Hoeve et al. 1981a, b, Hoeve and Quirt 1984, Percival et al., 1993) and resultant anomalous K$_2$O/Al$_2$O$_3$ ratios in the sandstone (Earle and Sopuck 1989). Sudoite, an Al-Mg-rich di-tri, octahedral chlorite (Percival and Kodama, 1989), is present in both alteration types. Local silicification fronts are also present in some of the larger de-silicified alteration systems, e.g. Cigar Lake (Andrade 2002).

Deposit-related silicification (Q2 event), recorded notably by drusy quartz-filled fractures (McGill et al. 1993) and pre-ore tombstone-style silicification (Q1; Yeo et al. 2001b; Mwenifumbo et al., 2004, 2005), are particularly intense in Athabasca sandstone sequences above or proximal to basement quartzite. The tombstone silicification of the Athabasca Group preserves diagenetic hematite and dickite (Mwenifumbo et al. 2005) and microbial laminae (Yeo et al. 2005a) very close to the McArthur River deposit.

Illite-kaolinite-chlorite alteration halos are up to 400 m wide at the base of the sandstone, thousands of meters in strike length, and several hundreds of meters vertical extent above deposits (Wasilyuk, 2002, for McArthur River; Kister et al.,
2003, for Shea Creek; and Bruneton, 1993, for Cigar Lake). This alteration typically envelops the main ore-controlling structures, forming plume-shaped or flattened elongate bell-shaped halos that taper gradually upward from the base of the sandstone (Hoeve and Quirt, 1987) Figure 70. Potassium, magnesium and aluminum ratios are diagnostic of the different clay types.

Hydrothermal, ore-related alteration effects are superimposed on pre-existing alteration assemblages, including the paleo-weathering alteration recorded by the red-green profile below the basal unconformity, and the initial detrital clay mineralogy that is locally preserved in silicified zones (Mwenifumbo et al., 2005). This has led to confusion as to which process generated the red-green transition at the basement-sandstone unconformity. Ore-related white clay alteration is superimposed on the red hematitic paleo-weathering alteration (McDonald, 1980). Very dark coarse grained hematite alteration forms a cap over ore deposits such as Cigar Lake (Andrade 2002). This very dense, crystalline, hydrothermal hematite alteration contrasts with the brick red, fine-grained hematite that is interpreted as a detrital or very early diagenetic mineral, and is intimately interlayered with clay minerals to form micro-laminae in oncoidal hematite beds (Yeo et al., 2005a).

Figure 71 - Simplified mineral paragenesis of the Paleo- to Mesoproterozoic Athabasca, Thelon and Kombolgie basins, after Figure 10.21 of Kyser et al. (2000). The ages of primary uraninite, the first deposition of U in a given deposit, may be different for different deposits, hence U1 (~1500-1600) and U1' (~1350-1400) might both be primary in the Athabasca Basin. Depositional ages in the Athabasca Group are Zv: U-Pb on volcanic zircon in Wolverine Point Formation, and Os-Re: primary organic matter in Douglas Formation.
INFRARED ACTIVE ALTERATION MINERALS ATHABASCA BASIN

Figure 72 - The common alteration minerals found in the Athabasca Basin deposits include dravite (magnesiofoitite Mg end member), illite, kaolinite and dickite.

Figure 73 - Chlorites found in the Athabasca suite. Mg-chlorite and the unique sudoite are plotted.

Figure 74 - Other minerals observed are quartz, hematite, monazite (REE), dolomite and siderite.
DISCUSSION

Starting with the PIMA™, it took about four hard years to introduce the technology to the industry. It was very alien to users initially, as most universities do not teach infrared analytical methods. Those that do, usually treat the subject very briefly, as most departments do not have expensive infrared instrumentation. Once applications, case studies, spectral libraries, some simple mineral identification software, and a well-documented training course and manuals were developed, it became easier to sustain the method.

As more and more companies utilized the technology, word spread via the "grapevine". Most companies would not cite PIMA™ analyses in their public presentations and papers. They all considered reflectance spectroscopy their "secret weapon".

With the introduction of the TerraSpec™, some of the secrecy has dissipated and companies are more willing to acknowledge the application of reflectance spectroscopy to their exploration programs. Additionally, VNIR analysis has been used for over 10 years and is no longer so exotic.

Spectroscopy is used for field and core reconnaissance, core logging, target generation, vectoring, blast hole logging, and remote sensing ground truth. It has its biggest successes in epithermal gold systems, porphyry, IOCG, skarns, kimberlites and unconformity uranium.

Reflectance Spectroscopy has changed the way exploration is done, how core is targeted, drilled and logged and alteration systems modeled. It has made some impact on mine metallurgy, grade control and production. It has opened the important world of alteration mineralogy to a rapid, non-invasive, identification of alteration minerals in-situ and instantly, in the field, in the core shack, on the pit benches and faces and underground in the tunnels. It has expanded the geologist’s knowledge about the minerals in alteration systems, their associations, and the environments in which they form, therefore leading to a better and more predictive understanding of mineral deposits. Now, clays, phyllosilicates, and carbonates, minerals that have similar physical appearances and have always been difficult to identify by hand lens, are known in 3, 5, or 10 seconds. This analysis can be in-situ in the field, at the outcrop, and/or in the core shack.

Reflectance spectroscopy has been used by all the major and many smaller companies in some capacity. There are nearly 400 instruments in the industry between the different manufacturers. Following is a partial list of companies who own or have owned spectrometers and many have more than one (apologies to those left off the list) : Alamos Gold, Almaden, Anglo American, Anglo-Gold-Ashanti, Barrick, BHP, Balkan Minerals, Battle Mountain, Billiton, Bolnisi, Buenaventura, Codelco, Collahausi, Cornerstone, CRA, Crystalllex, Cyprus, De Beers, Depromisa, El Dorado, Formicruz, Gencor, GoldCorp, Goldfields, Gryphon Gold, Hecla, Hemlo, Homestake, Hochschilds, Iamgold, Incor, Inmet,

Many more companies have either rented instruments or used consultants who have done spectral analysis for them.

The uranium companies are where spectroscopy makes a major contribution to drill target and core logging of alteration to vector to mineralization. The alteration minerals and clays occur in unique relationships to ore and spectroscopy is the major way to define these relationships and establish vectors. The companies include Areva, Cameco, Can-Alaska, Forum Uranium, Jabiru, JNR, Triex, Uranium City, Uravan, Uex. There are over 30 instruments in Canada and Australia dedicated to this task alone.

DeBeers funded the original PIMA development and then used close to 20 PIMAs to characterize alteration in kimberlite pipes globally. G-10 garnets, as diamond indicators, have a distinctive chrome feature in the visible, easily analyzed by the Terraspec™.

The alteration model for Pierina was done with the PIMA. Newmont at Yanacocha evolved from PIMA™ core logging into using the Terraspec™ to define their alteration and drill targets. Novagold used the FieldSpec Pro (FR) at Donlin Creek, as did Barrick at Alta Chicama and Ivanhoe at Oyu Tolgoi. It goes on with Lac and Barrick in the El Indio belt where they redefined alteration and mineralization episodes finding major advanced argillic alteration throughout the belt; Placer Dome logged core at La Coipa. At Valedaro, in Argentina, no hole was drilled before adjoining holes had been checked with the spectrometer as they had discovered a unique silica signature associated with gold zones. At Tres Cruces in Peru, Oro Peru defined mineralized zones and degrees of alteration using the presence of ammonia in illite or buddingtonite. Noranda and Rio Tinto use their FR's for remote sensing ground truth, as has Buenaventura, Newmont, Teck-Cominco, WMC and Barrick.

Perhaps the greatest use for spectroscopy is in core logging. One of the major success stories, as mentioned, is in the unconformity uranium deposits where companies have to drill blind targets through glacial till and sandstone down through the unconformity, where the ore lies, to the basement. The alteration mineralogy zoning is very diagnostic relative to proximity to ore. The spectrometers are the most rapid, economic way to vector to the ore bodies and define them.

Reflectance spectroscopy has become a robust method, routinely used in a wide variety of applications in the industry,
Acknowledgements

This paper is dedicated to all those field geologists who persevered from PIMA to Terraspec; who collected thousands of spectra and applied their new found knowledge to better understanding their deposits and alteration systems. I acknowledge all of you for your individual and collective contributions to this wonderful discipline of field spectroscopy.

A special acknowledgement to Graham Hunt, who gave me the first taste of reflectance spectroscopy years ago at the U.S. Geological Survey. Also to Alexander F.H. Goetz, who has not only given me knowledge and opportunity, but taught me a great deal about instrumentation and was supportive in the creation of the fabulous TerraSpec™. And to the Team of engineers, scientists, and technical support people at Analytical Spectral Devices who created the Terraspec™.

Acknowledgement as well to Terry Cocks of Australia, who with his CSIRO colleagues, created the PIMA, the first innovative step in this mini-revolution.

Another special salute to Anne Thompson, whose tireless and absolutely top notch efforts through the years has made reflectance spectroscopy an accepted analytical tool for all those Canadian Juniors doing very grassroots exploration, pioneers in their own rights.


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