Magnetic petrology of igneous intrusions: implications for exploration and magnetic interpretation

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ABSTRACT

Magnetic petrology integrates rock magnetism and conventional petrology in order to define the processes that create, alter and destroy magnetic minerals in rocks. By relating magnetic mineralogy, bulk magnetic properties, petrology and geochemistry to observed magnetic anomalies an understanding of the geological factors that control magnetic signatures is obtained, which can be used to improve *geological* interpretation of magnetic surveys.

The magnetic properties of igneous intrusions, and hence the magnetic anomalies associated with them, reflect bulk rock composition, redox state, hydrothermal alteration and metamorphism. These geological variables are in turn controlled by tectonic setting, composition and history of the source region, depth of emplacement and nature of wall rocks. The fundamental control on magnetic mineralogy and bulk magnetic properties is partitioning of iron between silicate and oxide phases, which is strongly influenced by oxidation ratio. This paper reviews and synthesises information on relationships between the chemistry, mineralogy and metallogenic associations of igneous intrusions and their magnetic properties. Although links between magnetic properties and broad rock names are tenuous, refined rock classification enables magnetic properties to be predicted with reasonable confidence.

Oxidised, magnetite-series, and reduced, ilmenite-series granitoids have quite distinct metallogeny. Cu, Mo and Au are associated with oxidised granitoids and Sn with reduced granitoids. Fractional crystallisation, which has a distinctive magnetic expression, plays an important role in generating magmatic-hydrothermal ore deposits. Hydrothermal alteration profoundly affects magnetic properties, in a generally predictable fashion. Implications for interpretation of magnetic anomalies associated with igneous intrusions and recognition of magnetic signatures of potential intrusive-related ore deposits are adduced.

Keywords: magnetic petrophysics, magnetic petrology, magnetic anomalies, rock magnetism, interpretation of magnetic surveys, igneous intrusions, granitoids, granitoid-related mineralisation.

INTRODUCTION

The magnetic method has been widely used in mineral exploration for decades. Recent improvements in magnetic data acquisition, processing and presentation and reduced airborne acquisition costs have increased the utility and importance of magnetic surveys, particularly high-resolution aeromagnetic surveys. Increasingly, high quality surveys of large areas are becoming available at reasonable cost. This has led to increasing emphasis on magnetic methods in area selection and regional mapping, as well as prospect-scale mapping and drill targeting.

The mineral exploration industry has now reached a stage where the ability to acquire, process and present magnetic survey data far outstrips capacity to interpret the surveys. There is often far more geological information in these very large data sets than can be presently extracted in the time available for interpretation. Better understanding of the

relationships between magnetic signatures and geology can facilitate the interpretation process and produce more reliable geological interpretations.

A crucial limitation of interpretation of magnetic surveys arises from the fundamental non-uniqueness of potential field source distributions. This ambiguity in source geometry can only be addressed by constraining models. The most important control on the reliability of magnetic models is information on magnetic properties. Understanding of the factors that determine magnetisation intensities and directions for the geological units within the survey area is essential for resolving geological ambiguity in order to produce a reliable interpretation of subsurface geology.

Igneous intrusions comprise a substantial portion of many geological provinces and intrusive-related mineralisation is a major exploration target. Information on the magnetic petrology of igneous intrusions should therefore assist geological mapping and an understanding of the relationships between magnetic properties and metallogenic associations of intrusions is important in exploration for intrusive-related ore deposits. Extensive background material that cannot be included in this summary paper can be found in Clark et al. (1992a).

PRINCIPLES OF MAGNETIC PETROLOGY

What is magnetic petrology?

Magnetic petrology integrates rock magnetism and conventional petrology to characterise the composition, abundance, microstructure and paragenesis of magnetic minerals in order to define the processes that create, alter and destroy magnetic minerals in rocks. By relating magnetic mineralogy, bulk magnetic properties and petrology to observed magnetic anomalies an understanding of the geological factors that control magnetic signatures is obtained, which can be used to improve geological interpretation of magnetic surveys. Dunlop and Özdemir (1997) have provided a comprehensive and up-to-date overview of rock magnetism. There is no corresponding textbook on magnetic petrology. Useful reviews of magnetic petrological principles have been given by McIntyre (1980), Grant (1985) and Frost (1991a). Clark et al. (1992b) presented several magnetic petrological case studies. Clark and Emerson (1991) summarised magnetic properties of rocks and some principles of rock magnetism and magnetic petrology. Clark (1997) tabulated magnetic properties of rock-forming minerals, and reviewed general aspects of magnetic petrophysics and magnetic petrology.

Magnetic properties of rocks reflect the partitioning of iron in the rock between strongly magnetic oxides and/or sulphides and weakly magnetic phases (silicates, carbonates etc.). This partitioning depends on chemical composition, oxidation ratio of the iron, and petrogenetic conditions. Thus a host of geological factors influence magnetic properties and simplistic correlations between magnetic properties and lithotype are generally unreliable. It is dangerous to extrapolate empirical correlations between mapped geology and magnetics in one area to another area, ignoring changes in depositional environment, metamorphic grade or structural setting.

For the purposes of subsequent discussion, an informal classification scheme, based on rock susceptibility (k) is used. Igneous rocks are classified as

1. diamagnetic (DIA) if $k < 0$,
2. paramagnetic (PM) if $0 < k < 1260 \times 10^{-6}$ SI ($100\,\mu$G/Oe),
3. weakly ferromagnetic (WFM) if 1260×10^{-6} SI $\leq k < 3770 \times 10^{-6}$ SI ($300\,\mu$G/Oe),
4. moderately ferromagnetic (MFM) if 3770×10^{-6} SI $\leq k < 37{,}700 \times 10^{-6}$ SI ($3000\,\mu$G/Oe),
5. strongly ferromagnetic (SFM) if $k \geq 37{,}700 \times 10^{-6}$ SI ($3000\,\mu$G/Oe).

Diamagnetic igneous intrusions are extremely rare. The approximate magnetite contents corresponding to the ferromagnetic classes are: 0.02 vol % to 0.1 vol % for WFM intrusions, 0.1 vol % to 1 vol % for MFM intrusions and greater than 1 vol % for SFM intrusions. Rocks that have susceptibilities low enough to fall into the paramagnetic class contain at most trace amounts of ferromagnetic *(sensu lato)* minerals, such as magnetite or monoclinic pyrrhotite. In these rocks, the measured susceptibility is generally dominated by contributions from paramagnetic minerals. Because paramagnetic minerals do not carry any remanent magnetisation, the remanent magnetisation of PM intrusions is very weak. Ferromagnetic intrusions, on the other hand, may carry significant remanence.

The concept of oxygen fugacity

Standard textbooks on petrology treat the concept of oxygen fugacity in a geological context. Oxygen fugacity $(f\mathrm{O}_2)$ is measured in units of pressure and is formally defined as the chemical activity of oxygen. Apart from a small correction due to departures from ideal gas behaviour, $f\mathrm{O}_2$ is equal to the partial pressure of oxygen gas. It should be noted that the abundance of free oxygen is vanishingly small in magmas and hydrothermal fluids. Nevertheless, $f\mathrm{O}_2$ is a well-defined thermodynamic variable that can be controlled in the laboratory and can be deduced from mineral assemblages. Frost (1991b) has recently clarified some common misconceptions about oxygen fugacity and given an unusually clear and succinct treatment of the subject.

Iron, which is the fourth most abundant element in the Earth's crust, exists in three oxidation states: metallic (Fe°), ferrous (Fe^{2+}) and ferric (Fe^{3+}) iron. Oxygen fugacity is a variable that strongly influences the propensity for iron to occur in a particular oxidation state. At very low oxygen fugacities, such as in the Earth's core, in some serpentinised ultramafic rocks, and in a few exceptionally reduced lavas that have reacted with carbonaceous material, iron occurs as the native metal. Iron occurs in the divalent ferrous state at higher oxygen fugacities. In silica-bearing systems the ferrous iron is incorporated mainly into silicate minerals. With increasing oxygen fugacity, iron occurs in both the divalent and trivalent states and is incorporated into magnetite as well as silicates. At still higher oxygen fugacities, iron occurs in the ferric state and is incorporated into haematite. Note that the relative terms "low" and "high"

$f\mathrm{O}_2$ depend strongly on temperature (T). At 500°C an oxygen fugacity of 10^{-15} bar is strongly oxidising for most minerals, but at 1000°C the same $f\mathrm{O}_2$ would correspond to very reducing conditions.
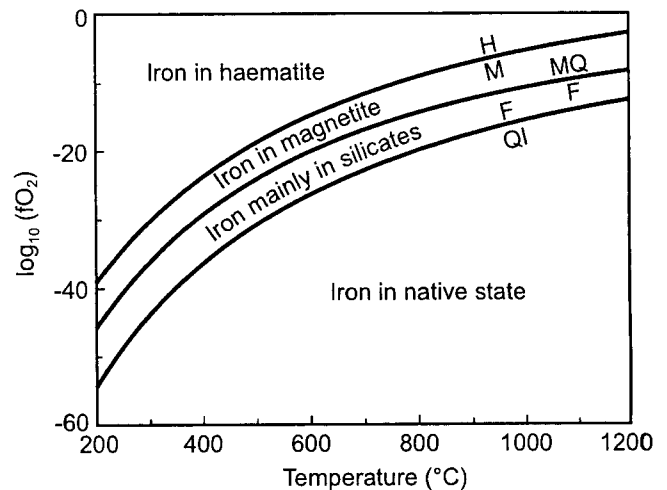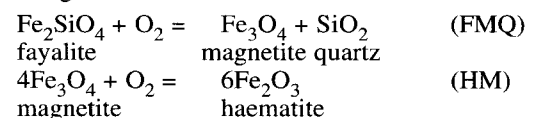

Figure 1. Plot of oxygen fugacity, expressed as $\log_{10}$ $(f\mathrm{O}_2)$, versus temperature showing the relative stabilities of the various oxidation states of iron in the system Fe-Si-O (after Frost, 1991b). Below the quartz-iron-fayalite (QIF) buffer iron is present as Fe°; between IQF and the fayalite-magnetite-quartz (FMQ) buffer iron occurs in the ferrous (Fe^{2+}) oxidation state; between FMQ and the haematite-magnetite (HM) buffer iron occurs in both ferrous and ferric (Fe^{3+}) oxidation states; and above HM iron is in the ferric state.

In the system Fe-O-SiO_2, the fayalite-magnetite-quartz (FMQ) buffer marks the lower oxygen fugacity limit for the stability of magnetite and the haematite-magnetite (HM) buffer marks the upper oxygen fugacity limit (Figure 1). The corresponding reactions are:

$$\mathrm{Fe}_2\mathrm{SiO}_4 + \mathrm{O}_2 = \quad \mathrm{Fe}_3\mathrm{O}_4 + \mathrm{SiO}_2 \qquad \text{(FMQ)}$$
fayalite   magnetite quartz

$$4\mathrm{Fe}_3\mathrm{O}_4 + \mathrm{O}_2 = \quad 6\mathrm{Fe}_2\mathrm{O}_3 \qquad \text{(HM)}$$
magnetite   haematite

Whether or not magnetite is precipitated from an igneous melt that is cooling along a particular T-$f\mathrm{O}_2$ path depends on the overall composition of the melt. For example, substitution of Mg for Fe in silicate minerals stabilises them to higher oxygen fugacity (Frost and Lindsley, 1991). In particular, addition of Mg reduces the activity of fayalite in olivine, thereby shifting the equilibrium in the FMQ reaction to the left. As a result, small amounts of magnetite and quartz react to produce fayalite, thereby partially restoring the fayalite activity, plus oxygen, which increases the oxygen fugacity. Thus the olivine-magnetite-quartz buffer is displaced upwards from FMQ and the stability field of magnetite is restricted. At higher Mg contents, this simple picture is complicated by reaction of Mg-rich olivine with quartz to produce orthopyroxene + magnetite. Thus, as shown in Figure 2, at high temperatures the oxygen fugacity of the Mg-rich Fe-O-SiO_2-MgO system is defined by either a quartz-orthopyroxene-magnetite buffer curve (if the melt is saturated in quartz) or an olivine-orthopyroxene-magnetite buffer curve (if the melt is olivine-saturated). It is evident from Figure 2 that higher Mg content of a melt tends to restrict the occurrence of magnetite to higher oxygen fugacities.

On the other hand, substitution of ferrous iron + titanium for ferric iron in titanomagnetite reduces the activity of magnetite and displaces the fayalite-titanomagnetite-quartz equilibrium downwards with respect to FMQ. Thus, titanomagnetite is stable in igneous rocks at lower oxygen fugacities than is end-member magnetite. Similarly, because
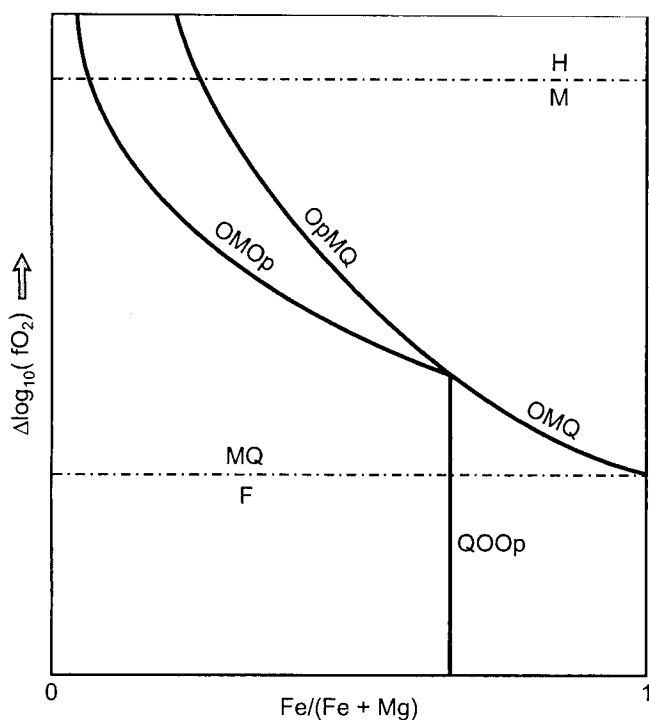
Figure 2. Schematic Δlog10 (fO_2) versus Mg/(Fe+Mg) diagram showing effects of adding MgO to the Fe-Si-O system at a fixed temperature (after Frost and Lindsley, 1991). OMQ = olivine-magnetite-quartz, QOOp = quartz-olivine-orthpyroxene, OpMQ = orthopyroxene-magnetite-quartz, OMOp = olivine-magnetite-orthopyroxene, OHQ = olivine-haematite-quartz, OHOp = olivine-haematite-orthopyroxene. The parameter $\Delta\log fO_2$ is defined as $\log_{10}$ (fO_2) - $\log_{10}$ $(fO_2$: FMQ). The values of $\Delta\log_{10}$ (fO_2) for the FMQ and MH buffers are shown for reference. Note that these equilibria become displaced towards higher absolute fO_2 as Mg/(Mg+Fe) increases, because Mg preferentially enters olivine over magnetite and magnetite over haematite.

Ti substitutes even more readily into haematite than into magnetite, addition of Ti to the system displaces the HM buffer to lower fO_2.

For many plutonic rocks that behave essentially as closed systems during their history, in particular many tholeiitic rocks, the Fe/Mg ratio of the silicates, plus the Ti content and the ferrous/ferric ratio of the oxides, monitor and, in effect, control the oxygen fugacity. In this case the cooling history of the rock is characterised by a path in fO_2 - T space that is approximately parallel to the standard mineral-buffered curves. In turn the oxygen fugacity influences the composition of the fluid phase and the stability of graphite and sulphides in such igneous rocks.

Fluid buffering, rather than mineral buffering, of oxygen fugacity is evidently important during hydrothermal processes that have large fluid-rock ratios. Fluid buffering may also play a role in some magmatic processes. If the initial volatile content of the magma is sufficiently high, the oxygen fugacity may be largely controlled by the fluid phase, rather than by the ferric and ferrous iron contents of the melt and the crystallising minerals. Takagi and Tsukimura (1997) suggest that the SO_2 - H_2S buffer may be important in the evolution of oxidised granitic rocks, provided the initial SO_2 content of the magma is greater than 250 ppm. Because this buffer curve lies below the FMQ buffer at high temperatures, but intersects FMQ at ~ 850°C and lies well above FMQ at lower temperatures, it represents a relatively oxidising cooling trend that in principle can oxidise ferrous iron in silicates to magnetite via the reaction:

$$9FeO + SO_2 + H_2O \rightarrow 3Fe_3O_4 + H_2S.$$
in silicates
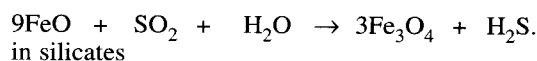
Takagi and Tsukimura (1997) calculate that initial SO_2 contents of 250-1900 ppm by weight as the dominant sulphurous species are required to precipitate 0.2-1.5 vol % magnetite from granitic melts and show that other fluid buffers, e.g. H_2 - H_2O, CO_2 - CH_4, or CO_2 - CO, cannot produce the oxidising trends that are inferred for many calc-alkaline granitic rocks. The general relevance of sulphur dioxide buffering of melts is still an open question, however, because the primary contents of sulphur species and other volatile phases in magmas is poorly known (P. Blevin, pers. comm). Reported sulphur contents in granitoids are lower than the values that are required to produce substantial magnetite, but this may reflect significant late-stage loss of sulphur carried away by hydrothermal fluids, which sometimes produce related sulphide ore deposits.

Frost (1991b) points out that there can be no *unique* correlation between fO_2 during rock formation and Fe^{3+}/Fe^{2+} of the rock. For example, rocks that contain the same mineral assemblages must have formed at similar fO_2, but if they have very different abundances of the iron-bearing minerals, they may have very different absolute and relative abundances of ferrous and ferric iron. However, oxygen fugacity of melts and glasses is simply dependent on chemical composition, in particular the relative abundance of ferrous and ferric iron. Similarly, in the case of volcanic rocks that are relatively free of cumulate minerals fO_2 can be calculated at a given temperature and pressure, corresponding to crystallisation conditions midway between the liquidus and solidus for the rock, from the whole rock chemical composition, including ferrous and ferric iron. Kress and Carmichael (1991) show that the Fe^{3+}/Fe^{2+} ratio is by far the most important term in the relationship between fO_2 and chemical composition of volcanic rocks.

Blevin (1994) has shown that ferric/ferrous iron ratios in granitoid rocks are also very highly correlated with oxygen fugacity as calculated from the chemical composition (and confirmed by mineral assemblages that are dependent on oxygen fugacity). Thus Fe^{3+}/Fe^{2+}, often measured as Fe_2O_3 /(FeO + Fe_2O_3), can in practice be used as a proxy for oxygen fugacity in granitoids, in spite of the theoretical possibility that the nexus between oxidation state and fO_2 might be broken for rocks that formed under very different conditions or that have exotic compositions.

Figure 3(a) plots isopleths in fO_2 - T space for various titanomagnetite and ilmenite compositions, with the FMQ and HM buffers shown for comparison. Titanomagnetites are solid solutions of magnetite, i.e. $Fe^{3+}[Fe^{2+}Fe^{3+}]O_4$, and ulvospinel, $Fe^{2+}[Fe^{2+}Ti^{4+}]O_4$, whereas natural ilmenites invariably incorporate some haematite, $Fe^{3+}_2O_3$, in solid solution with ilmenite, $Fe^{2+}Ti^{4+}O_3$. The square brackets indicate octahedral cations in the spinel phases. Note that the titanomagnetite isopleths are quite oblique to the oxygen buffer curves, whereas the ilmenite isopleths are subparallel to the buffers. This implies that as an igneous melt cools and solidifies along a trajectory that is approximately parallel to FMQ, the stable titanomagnetite composition evolves from very ulvospinel-rich at high temperatures to magnetite-rich at subsolidus temperatures. Corresponding changes in ilmenite composition are less pronounced. For example, a relatively oxidised melt at ~1000°C (point A in Figure 3(a)) may be in equilibrium with Fe-Ti oxide compositions of 50 mole % magnetite-50 mole % ulvospinel (Mt50) and 85 mole % ilmenite-15 mole % haematite (Ilm85). If this magma cools slowly along the Ilm85 isopleth, which almost parallels FMQ, the ilmenite composition remains unchanged, but the equilibrium titanomagnetite composition at ~600°C (point B) evolves to Mt90.
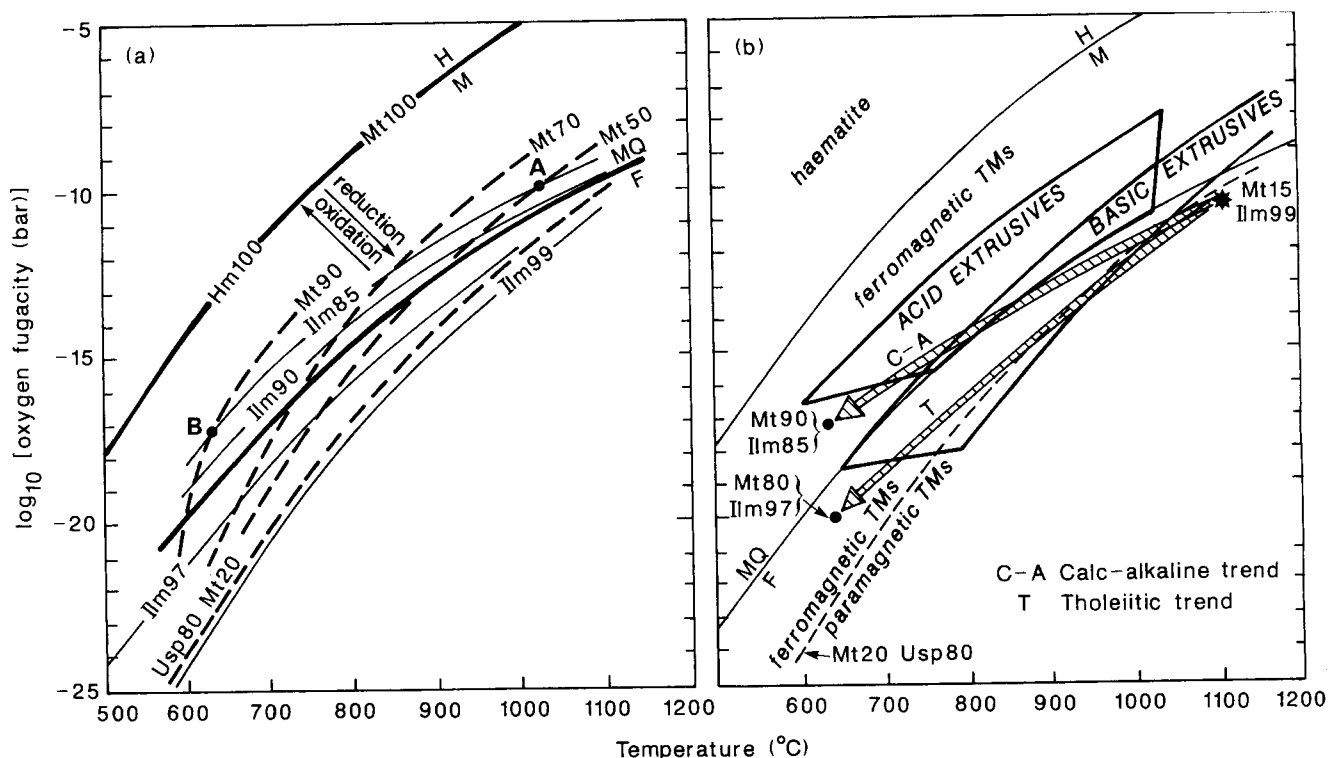
Figure 3. (a) Isopleths for various titanomagnetite and ilmenite compositions, plotted in fO_2 - T space. The FMQ and HM buffers are shown for comparison.

(b) Contrasting tholeiitic and calc-alkaline cooling trends in fO_2 - T space for an initially reduced ($\log_{10} [fO_2]$ = FMQ - 1), high temperature magma that is in equilibrium with Mt15 and Ilm99. The re-equilibrated Fe-Ti oxide compositions at ~600°C for the two cooling trends are indicated. The dashed line indicates the boundary between the stability fields of ferromagnetic (sensu lato) and paramagnetic titanomagnetites. The fields representing fO_2 - T conditions recorded by Fe-Ti oxides in basic and acid extrusive rocks (Haggerty, 1976) are also shown.

The final titanomagnetite composition found in the rock depends on the initial redox state of the magma (relatively oxidised magmas initially crystallise titanomagnetites with lower Ti than more reduced magmas) and the temperature at which the titanomagnetite composition is "frozen in", which depends on cooling rate. Rapidly cooled volcanic rocks quench in relatively titaniferous compositions that are metastable at low temperatures. On the other hand, in slowly cooled intrusions, oxide mineral compositions continue to re-equilibrate well below the solidus, producing titanomagnetites with progressively lower Ti, until the increasingly sluggish kinetics of Fe and Ti exchange between oxide phases inhibits further change. Furthermore, slowly cooled titanomagnetites tend to exsolve into intergrowths of magnetite-rich and ulvospinel-rich phases.

The magnetic properties of titanomagnetites depend on composition. Titanomagnetites with more than 80 mole % ulvospinel are paramagnetic at ambient temperatures and have very low susceptibility. Compositions with less Ti are ferromagnetic sensu lato. Consider a reduced ($\log_{10} [fO_2]$ = FMQ - 1), high temperature magma that is in equilibrium with Mt15 and Ilm99. If this magma were cooled very rapidly by being extruded onto the ocean floor, for example, the quenched titanomagnetite would be paramagnetic. However, if the magma is emplaced at depth and cools slowly, following a typical tholeiitic fO_2 - T cooling trend as shown in Figure 3(b), the Fe-Ti oxides re-equilibrate attaining compositions of Mt80 and Ilm97 by ~ 600°C. If this titanomagnetite composition is metastably stranded upon further cooling, the titanomagnetite is ferromagnetic at ambient temperature and greatly enhances the susceptibility of the rock. Thus, ferromagnetic titanomagnetites may form even under relatively reducing conditions, provided the cooling is sufficiently slow. Figure 3(b) also shows an alternative, more oxidised, cooling trend that is characteristic

of calc-alkaline magmas. Even though fO_2 decreases strongly with falling T, the calc-alkaline path falls more slowly than the FMQ buffer, so the system evolves to a relatively oxidised state that is in equilibrium with more oxidised mineral assemblages. In particular, the equilibrium Fe-Ti oxide compositions at ~ 600°C are Mt90 and Ilm85 for the calc-alkaline trend. Figure 3(b) is schematic, because initial magmatic conditions and cooling paths can vary substantially, but it serves to illustrate qualitative trends. Initial conditions of calc-alkaline magmas are generally more oxidising than those of tholeiitic magmas, so the final Fe-Ti oxide compositions may be even more oxidised than indicated in Figure 3(b). The fields representing fO_2 - T conditions recorded by Fe-Ti oxides in basic and acid extrusive rocks (Haggerty, 1976) are also shown in Figure 3(b).

Figure 4 plots the range of titanomagnetite compositions found in the major types of igneous rock. Note the tendency for decreasing Ti content of titanomagnetite, i.e. more magnetite-rich compositions, for more felsic compositions. There is also a clear tendency for lower Ti contents in titanomagnetites from intrusive rocks than for their extrusive analogues, reflecting greater re-equilibration during cooling for intrusive rocks. Paramagnetic titanomagnetite compositions are rare and are only found in a few mafic extrusive rocks with primitive compositions. The inferred primary magnetite composition of the Skaergaard gabbros, derived by reconstituting magnetite-ulvospinel intergrowths within single grains, is very Ti-rich and is close to a paramagnetic composition. However exsolution of primary titanomagnetite into relatively Ti-poor magnetite, which is ferromagnetic, and paramagnetic ulvospinel (or ilmenite, if oxidation-exsolution occurs) during slow cooling produces grains that are ferromagnetic overall. Although the saturation magnetisation of titanomagnetites depends strongly on composition, decreasing almost
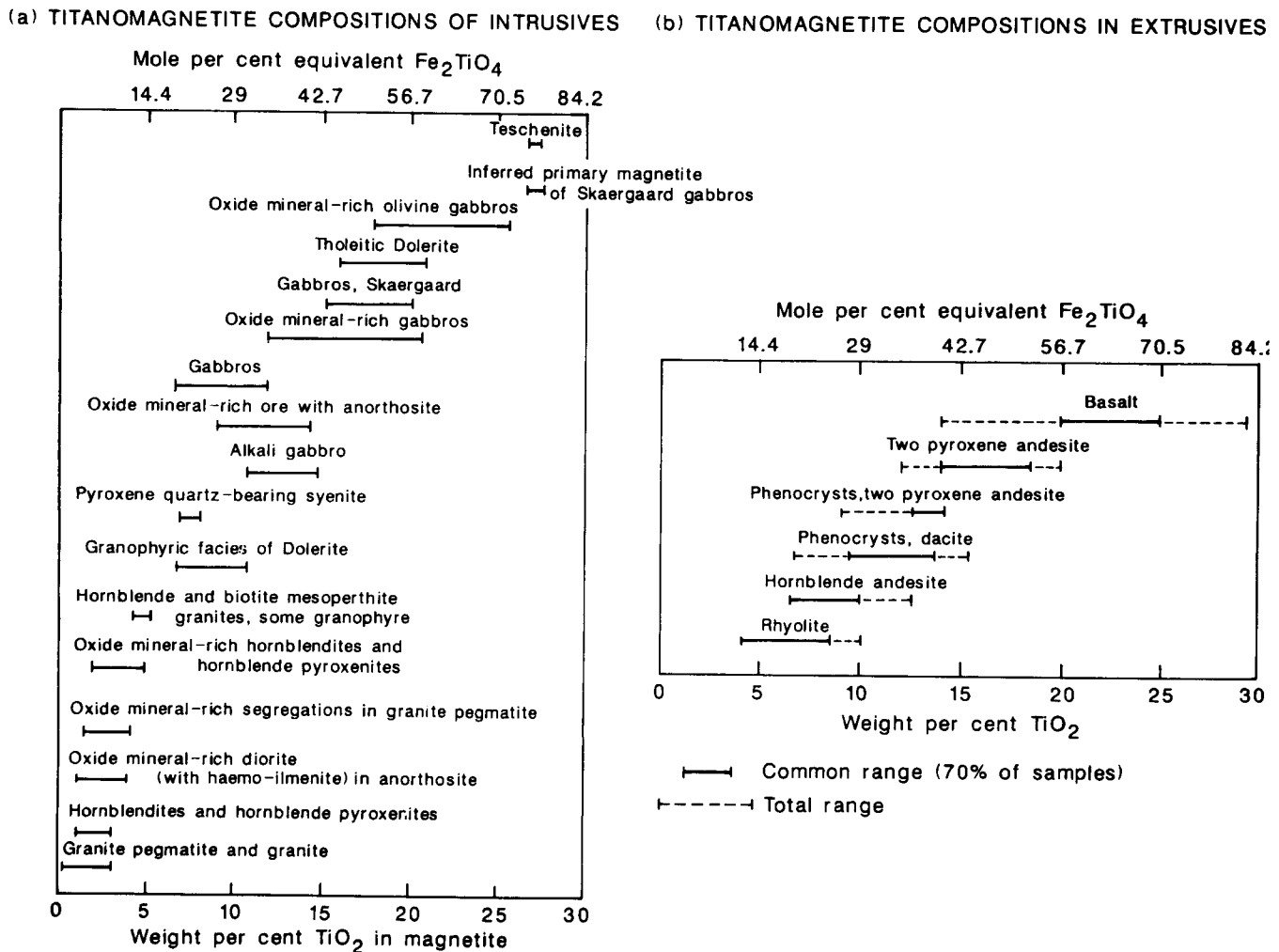
(a) TITANOMAGNETITE COMPOSITIONS OF INTRUSIVES   (b) TITANOMAGNETITE COMPOSITIONS IN EXTRUSIVES

Mole per cent equivalent Fe_2TiO_4

14.4   29   42.7   56.7   70.5   84.2

Teschenite

Inferred primary magnetite
of Skaergaard gabbros

Oxide mineral-rich olivine gabbros

Tholeitic Dolerite

Gabbros, Skaergaard

Oxide mineral-rich gabbros

Gabbros

Oxide mineral-rich ore with anorthosite

Alkali gabbro

Pyroxene quartz-bearing syenite

Granophyric facies of Dolerite

Hornblende and biotite mesoperthite
granites, some granophyre

Oxide mineral-rich hornblendites and
hornblende pyroxenites

Oxide mineral-rich segregations in granite pegmatite

Oxide mineral-rich diorite
(with haemo-ilmenite) in anorthosite

Hornblendites and hornblende pyroxenites

Granite pegmatite and granite

0   5   10   15   20   25   30
Weight per cent TiO_2 in magnetite

Mole per cent equivalent Fe_2TiO_4

14.4   29   42.7   56.7   70.5   84.2

Basalt

Two pyroxene andesite

Phenocrysts,two pyroxene andesite

Phenocrysts, dacite

Hornblende andesite

Rhyolite

0   5   10   15   20   25   30
Weight per cent TiO_2

Common range (70% of samples)
Total range

Figure 4. Range of titanomagnetite compositions found in the major types of igneous rock (after Buddington and Lindlsey, 1964).

linearly from 480 kAm^{-1} for pure magnetite to zero for Usp80, the susceptibility is only weakly dependent on Ti content for ulvospinel contents of less than ~70% (Clark, 1997). Thus the titanomagnetites carried by igneous rocks, ranging from gabbroic to granitic compositions, are almost invariably ferromagnetic and the susceptibility of the rock is essentially proportional to the modal titanomagnetite (allowing for intergrown paramagnetic phases in composite grains) and only weakly dependent on titanomagnetite composition. This conclusion differs from that of Grant (1985), who assumed that titaniferous magnetites have much lower susceptibilities than Ti-poor magnetite.

Relationship between lithology and magnetic properties

The data of Figures 5 and 6 are based on magnetic property measurements at the CSIRO Division of Exploration and Mining over the last 18 years and published studies and compilations. The systematic collection of petrophysical data by the geological surveys of Scandinavian countries, in particular, has greatly expanded the quantity and scope of the information available. It is evident from Figure 5 that each rock type exhibits a wide range of susceptibilities and that susceptibility values are not generally diagnostic of lithology. Classical rock names are in fact much too broad to be useful for classification of magnetic properties. This is because the susceptibility of most rocks reflects the abundance of accessory minerals, particularly magnetite (sensu lato), which are generally ignored in petrological classification.

Koenigsberger ratios (Q) can also vary quite widely (see Figure 6), but useful rules-of-thumb can be stated.

Ferromagnetic intermediate to felsic granitoid rocks contain multidomain magnetite, which is associated with Koenigsberger ratios less than unity (usually $Q < 0.5$, typically $Q \sim 0.2$). Furthermore, the remanence carried by such grains is generally unstable and is dominated by viscous remanence acquired in the recent field. However some, but not all, gabbros, norites and mafic diorites contain ultrafine pseudosingle domain to single domain magnetite hosted within silicate minerals, such as pyroxenes, olivine or plagioclase, as well as discrete multidomain grains. The ultrafine (<10 µm) grains are capable of carrying intense remanence and these rocks may accordingly exhibit Q values substantially greater than unity. Thus, magnetisation by induction can be assumed as a first approximation for the more felsic granitoids, whereas remanent magnetisation, possibly oblique to the present field, may be significant for mafic plutonic rocks.

Bimodal susceptibility distributions reflect ferromagnetic and paramagnetic populations

A notable feature of Figure 5 is that the magnetic susceptibilities of a number of rock types have distinctly bimodal distributions. Puranen (1989) presented results from very large petrophysical sampling programs in Finland. His data confirmed that all broad field names, such as "granite", "gabbro", "mica schist", "amphibolite" etc., exhibit distinctly bimodal susceptibility distributions. Figure 7 shows frequency distributions of susceptibility for major intrusive rock types, based on Puranen's data. The
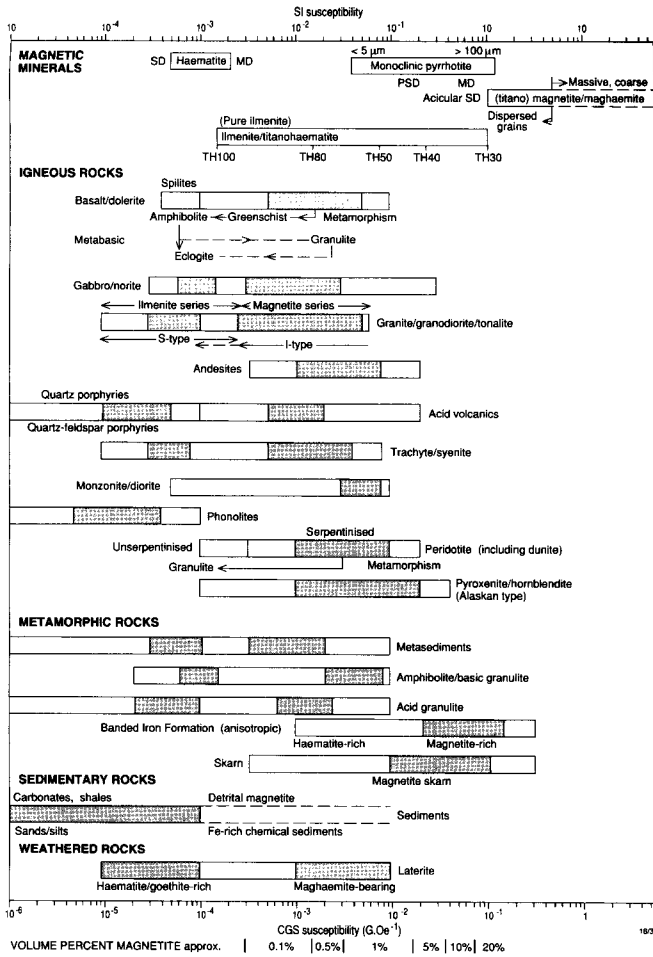
Figure 5. Range of magnetic susceptibilities for important magnetic minerals and major rock types. Stippled portions of bars indicate common susceptibility ranges for various lithologies. Note the bimodal susceptibility distributions for many rock types.

Figure 6. Range of Koenigsberger ratios for common rock types. Stippled portions of bars indicate common ranges.

two modes of the frequency distribution correspond to distinct paramagnetic and ferromagnetic populations, with a pronounced intervening gap. Iron in the weakly magnetic subpopulation is incorporated into paramagnetic silicate minerals, predominantly as Fe^{2+}, whereas similar rocks that are moderately to strongly magnetic contain significant Fe^{3+}, which is incorporated into magnetite.

Very highly oxidised rocks, however, tend to contain haematite rather than magnetite and are therefore also weakly magnetic. Within each of the subpopulations, the modal and mean values of susceptibility are much more closely related to rock type than for the total susceptibility distribution. For the paramagnetic subpopulation, in particular, the susceptibility is directly related to the chemical composition, which tends to have a restricted range for each lithology. Clark and Emerson (1991) give the relationship between iron content and susceptibility for paramagnetic rocks and between magnetite content and susceptibility for rocks that contain more than ~ 0.1% magnetite by volume.

When varietal mineralogy is incorporated into a refined rock classification, the bimodal susceptibility distribution tends to resolve into a paramagnetic subpopulation and a ferromagnetic subpopulation, each associated with a distinct mineralogy. Bimodality often also reflects the fact that important geological factors, such as geochemical affinity, alteration and metamorphic grade are not considered in the simple classification schemes used for most petrophysical summaries. A truly meaningful magnetic petrological
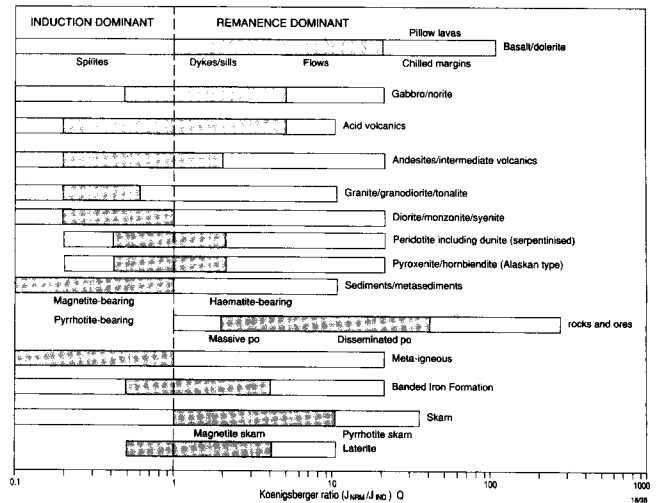
classification scheme must include chemical and/or mineralogical data for protoliths, plus information on metamorphic grade and environment, and/or alteration.

Similarly, when susceptibility distributions are considered on progressively smaller scales, the range of susceptibilities becomes more restricted. Within different geological provinces, the relative proportions of paramagnetic and ferromagnetic subpopulations differ from those of other provinces. It is often found that within sufficiently small areas, e.g. within a particular geological environment or simply within a single outcrop, all susceptibilities fall exclusively within one of the subpopulations. Thus, the distinct susceptibility subpopulations tend to reflect differing geological conditions, which are not considered in the primary rock classification schemes.

CLASSIFICATION OF INTRUSIVE IGNEOUS ROCKS

IUGS classification of plutonic rocks

The internationally accepted IUGS classification of mafic to felsic plutonic rocks (Le Bas and Streckeisen, 1991) is simply based upon the relative proportions of three major rock forming minerals: plagioclase (> An5); alkali feldspar (K-feldspar and albite); and either quartz (in oversaturated rocks) or a feldspathoid mineral, most commonly nepheline, in the case of an undersaturated rock. Figure 8 shows the fields and rock names on the QAPF double triangle. Ultramafic rocks, for which mafic minerals constitute 90% to 100% of the rock, are classified separately.

Given the fact that, in extreme cases, up to 90% of the mineral content of the rock may be ignored in the first-order classification, it is little wonder that magnetite abundance, for instance, is weakly correlated with rock name. It is also clear that there can be no unique correlation between rock name and bulk chemistry, given the wide range and variety of minor minerals that can be present within any one of the rock type fields. Of course, the classification is so useful and widely accepted because there are coherent patterns of mineralogical and chemical variation among plutonic rocks. Figure 9 illustrates some aspects of this coherency. A generalised plot of mineral composition for the full range of plutonic rock types is shown in Figure 9(a). Figure 9(b) shows average trends in plagioclase composition, mafic mineral contents and hornblende/biotite ratio in granitoid rocks, showing systematic variation with position in the QAP diagram.
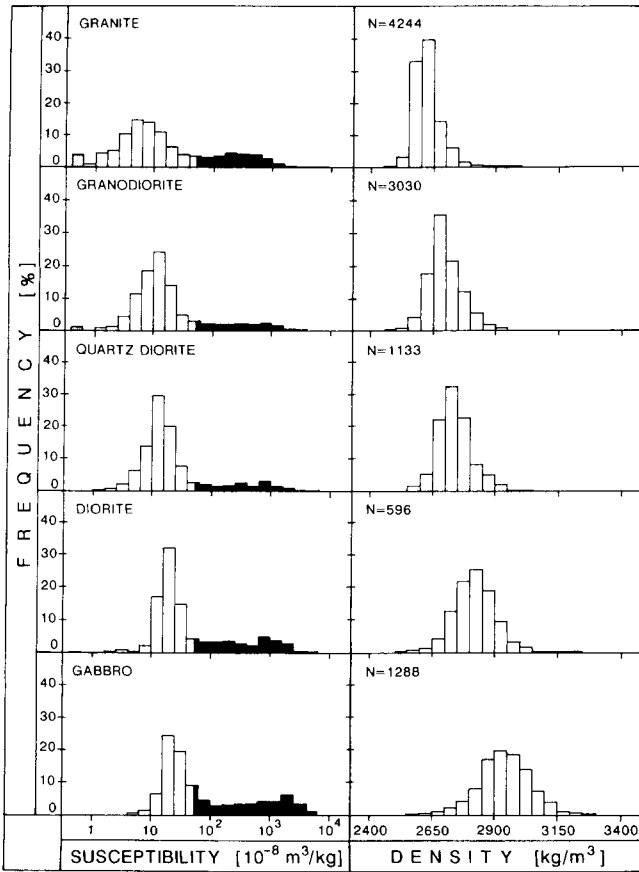
Figure 7. Histograms of SI mass susceptibility and density for plutonic rock types from Finland (after Puranen, 1989). Note the unimodal density distribution contrasting with the bimodal susceptibility distribution. The ferromagnetic subpopulation is shown as black; the small proportion of diamagnetic samples is shown hatched. To convert mass susceptibility to SI volume susceptibility, multiply by the density in kg/m^3.
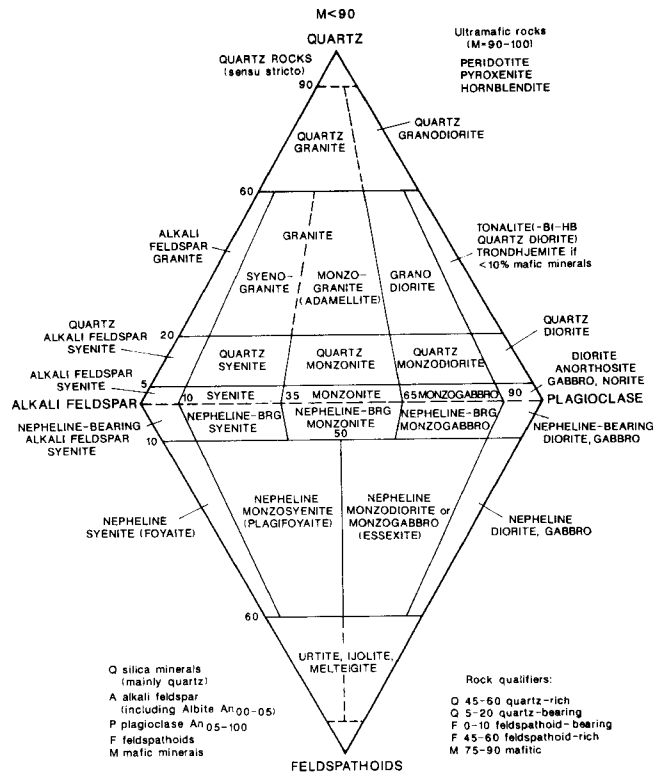

Figure 8. IUGS classification of plutonic rocks, based on the QAPF double triangle (Le Bas and Streckeisen, 1991). Q = quartz , A = alkali feldspar, P = plagioclase, F = feldspathoid (foid), M = mafic minerals, BI = biotite, HB = hornblende. Rock qualifiers are specified in terms of modal percentages. Ultramafic rocks, for which mafic minerals constitute 90% to 100% of the rock, are classified separately.

Spatially related plutonic rock series show clear mineralogical and chemical correlations with tectonic environment and relative time of emplacement, as shown in Figure 10. These various rock associations are characterised by different metallogeny and can be related to magnetic petrology much more reliably than to the broad IUGS rock names. This has implications for exploration, as use of magnetic methods for locating intrusion-hosted or intrusion-related mineralisation requires better understanding of the relationship between rock magnetisation and the geological factors that influence mineralisation.

Chemical classification of plutonic rocks

The following summary of chemical classification schemes for plutonic rocks is largely based on the excellent textbook by Hughes (1982).

Feldspars are the commonest minerals in igneous rocks, in which they constitute more than 50%, on average. Alumina occurs in a 1:1 ratio with oxides of the alkali metals or alkaline earth elements in feldspars. Thus departures from this ratio cannot be accommodated by varying the feldspar compositions or relative proportions, but must be expressed in the varietal mineralogy. Peraluminous rocks are oversaturated with respect to alumina, i.e. molar Al_2O_3 (A) exceeds the sum of $Na_2O + K_2O + CaO$ (denoted A/CNK>1), and are characterised by aluminous minerals, such as corundum (rarely), andalusite, sillimanite or kyanite, almandine garnet or, most commonly, muscovite. Peralkaline rocks, on the other hand, contain insufficient

alumina to consume all of the sodium and potassium in feldspars, i.e. molecular Al_2O_3 is less than $Na_2O + K_2O$ (A/NK < 1). Such rocks are characterised by minerals of the aegirine, riebeckite, arfvedsonite or aenigmatite classes. Metaluminous rocks are intermediate in alumina saturation, such that all the alumina, soda and potash can be accommodated in feldspars, with excess calcium appearing in the norm as diopside and in the mode as calcium-bearing pyroxene, amphibole etc. Peraluminous chemistry may result either from high Al content, or from low levels of Na, K or Ca. For example, mature sedimentary rocks, their metamorphic equivalents, and granitic rocks derived from partial melting of the metasediments are peraluminous because of the severing of the nexus between alumina and Na+Ca during the sedimentary cycle. Sodium is partitioned strongly into seawater and calcium into carbonates, leaving sedimentary rocks with excess alumina.

Quartz is a major constituent of many igneous rocks, and its presence or absence is a very significant petrological characteristic. Many minerals exhibit a clear sympathetic or antipathetic association with quartz. Oversaturated rocks contain free quartz, together with oversaturated (compatible) minerals (e.g. Al- and Ti-poor pyroxenes, feldspars, amphiboles, micas, fayalitic olivine). Undersaturated rocks contain undersaturated minerals that are antipathetic to quartz (e.g. nepheline, magnesian olivine, sodalite, leucite, Al- and Ti-rich augite).

The abundance of Na and K exerts a strong influence on the silica saturation state. In feldspars, every molecule of soda or potash in feldspars consumes six molecules of silica, whereas CaO only consumes two. Thus alkaline rocks, with relatively high Na and/or K for their silica content, have no excess silica to form a free silica phase and are undersaturated. The thermodynamic parameter, silica

(a)   MINERAL COMPOSITION OF MAJOR PLUTONIC ROCK TYPES

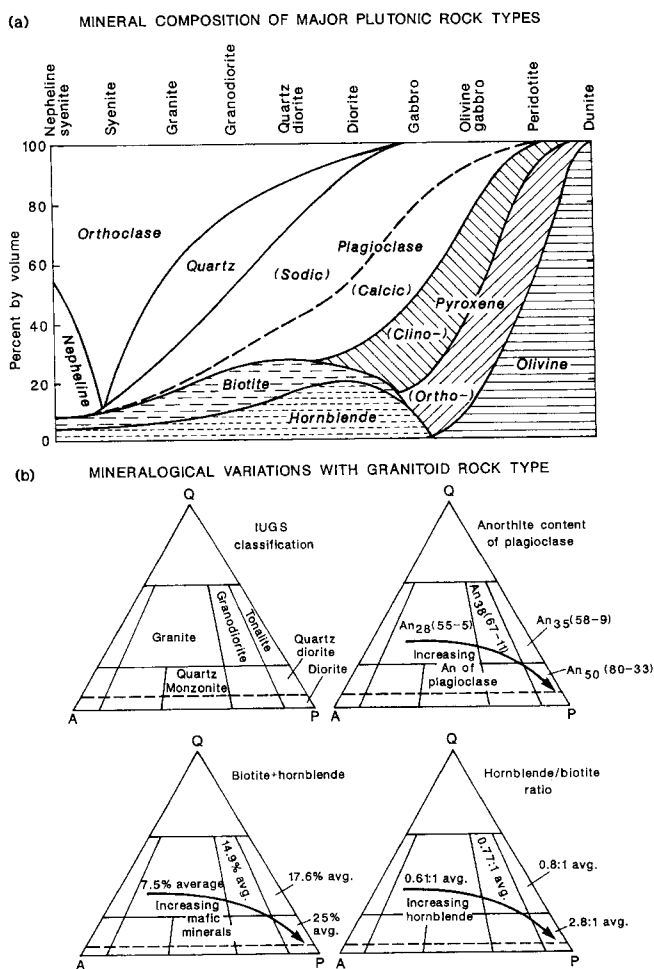(b)   MINERALOGICAL VARIATIONS WITH GRANITOID ROCK TYPE

Figure 9. (a) Generalised plot of mineral composition for the full range of plutonic rock types (after Washington and Adams, 1951). (b) Average trends in plagioclase composition (expressed as average and range of anorthite contents (%)), mafic mineral contents and hornblende/biotite ratio in granitoid rocks, showing systematic variation with position in the QAP diagram (after Hyndman, 1972).

activity, is strongly dependent on the alkali content for this reason. As an example, alkali basalts are silica undersaturated and are characterised chemically by normative nepheline. Undersaturated magnesian olivine is relatively abundant in these basalts, whereas tholeiitic basalts are silica saturated, with hypersthene in the norm. Olivine, if present, is in a reaction relationship to ferromagnesian pyroxene and was therefore out of equilibrium with the tholeiitic magma.

An important geochemical classification of igneous rock series is based upon Peacock's (1931) alkali-lime index (ALI), which is a measure of the relative alkalinity of a rock series derived by igneous differentiation from a parental magma. With increasing differentiation, accompanied by increasing silica content, CaO decreases while Na_2O and K_2O increase. There is a value of silica content, therefore, where the trend of CaO plotted against SiO_2 intersects the trend of Na_2O+K_2O versus SiO_2. This SiO_2 value (in weight per cent) is the alkali-lime index, and is lower for more alkaline rock series. Rock series are classified on the basis of their alkali-lime index into one of four categories: alkalic (ALI < 51), alkali-calcic (ALI = 51-56), calc-alkalic or calc-alkaline (ALI = 56-61) and calcic (ALI > 61), as shown in Figure 11(a).

Examples of igneous rock series representing each of the ALI categories include: tholeiitic basalts (calcic); basalt-andesite-rhyolite series (calc-alkalic); alkali basalt-phonolite

series (alkali calcic), and alkali syenite complexes (alkalic). The ALI provides a measure of the maturity of volcanic arcs, with igneous rock series tending to evolve from early mantle-derived, calcic magmatism, through calc-alkalic orogenic magmatism, reflecting crust-mantle interactions, to post-orogenic alkali-calcic or anorogenic alkalic magmatism.

Figure 11(b) shows a major difference in the behaviour of iron during differentiation of tholeiitic and calc-alkaline magmas. On a ternary plot of MgO, total iron and alkalis (AFM diagram) tholeiitic magmas show a pronounced initial iron enrichment trend, reflecting early crystallisation of Mg-rich olivine and pyroxenes. This trend is typical of many layered mafic complexes (e.g. the Skaergaard, Stillwater and Bushveld Complexes), for which the parental mantle-derived magma is anhydrous and relatively reduced. The initial oxygen fugacity in such magmas is too low to precipitate magnetite. The early-crystallising spinel phase in these intrusions is chromite. As fractional crystallisation proceeds, ferrous iron is increasingly sequestered in silicates and removed from the residual melt, whereas nearly all the ferric iron remains in the melt as the magma composition evolves along the iron-enrichment trend. Thus the Fe^{3+}/Fe^{2+}ratio steadily increases in the melt until the point is reached when magnetite can precipitate. The differentiation trend then turns towards the alkali apex of the AFM diagram as iron is removed from the melt in magnetite. In layered complexes, therefore, the primary ultramafic rocks near the base and the overlying lower gabbros are magnetite-free and have susceptibilities in the paramagnetic range. Overlying, more differentiated, gabbros, norites and anorthosites have higher susceptibilities, increasing upwards, due to the presence of intercumulus magnetic and the upper ferrogabbros and ferrodiorites are very strongly ferromagnetic due to copious amounts of cumulus magnetite. Although the unaltered ultramafic cumulates are paramagnetic, serpentinisation, particularly of olivine-rich layers, frequently produces secondary magnetite and produces susceptibilities in the MFM range.

Calc-alkaline series, typified by orogenic andesites and their plutonic equivalents, and minor related mafic and silicic rocks, show a quite different trend, with early depletion in iron and pronounced silica enrichment. This is thought to reflect more hydrous magmas associated with crust-mantle interactions in a subduction zone, with more oxidised parental magma and early, and continuing, crystallisation of Fe-Ti oxides and hydrous phases, such as hornblende. This leads to a pronounced depletion in iron in the more evolved members of a calc-alkaline series, whereas fractionated rocks derived from tholeiitic magma are relatively iron-rich. While calc-alkaline volcanics are subduction-related, calk-alkaline granitoids are not necessarily directly associated with subduction, but are often derived from partial melting of calc-alkaline source rocks produced during an earlier tectonic cycle. Tholeiitic magmas are associated with a variety of tectonic settings. These mainly, but not always, correspond to tensional regimes and include: mid-ocean ridges; mantle plume-related intraplate oceanic islands; and anorogenic continental settings, including flood basalts, major dolerite dyke or sill swarms and layered gabbroic complexes.

Source rock classification of granitoids

Chappell and White (1974) recognised two categories of calc-alkaline granitoids with very distinctive mineralogical, chemical and geological features, which were interpreted as reflecting different source rocks. S-type granitoids are derived from partial melting of (meta)sedimentary rocks, and I-type granitoids from igneous source material. S may also stand for "Supracrustal" and I may represent "Infracrustal". S-type granitoids are characterised by metasedimentary
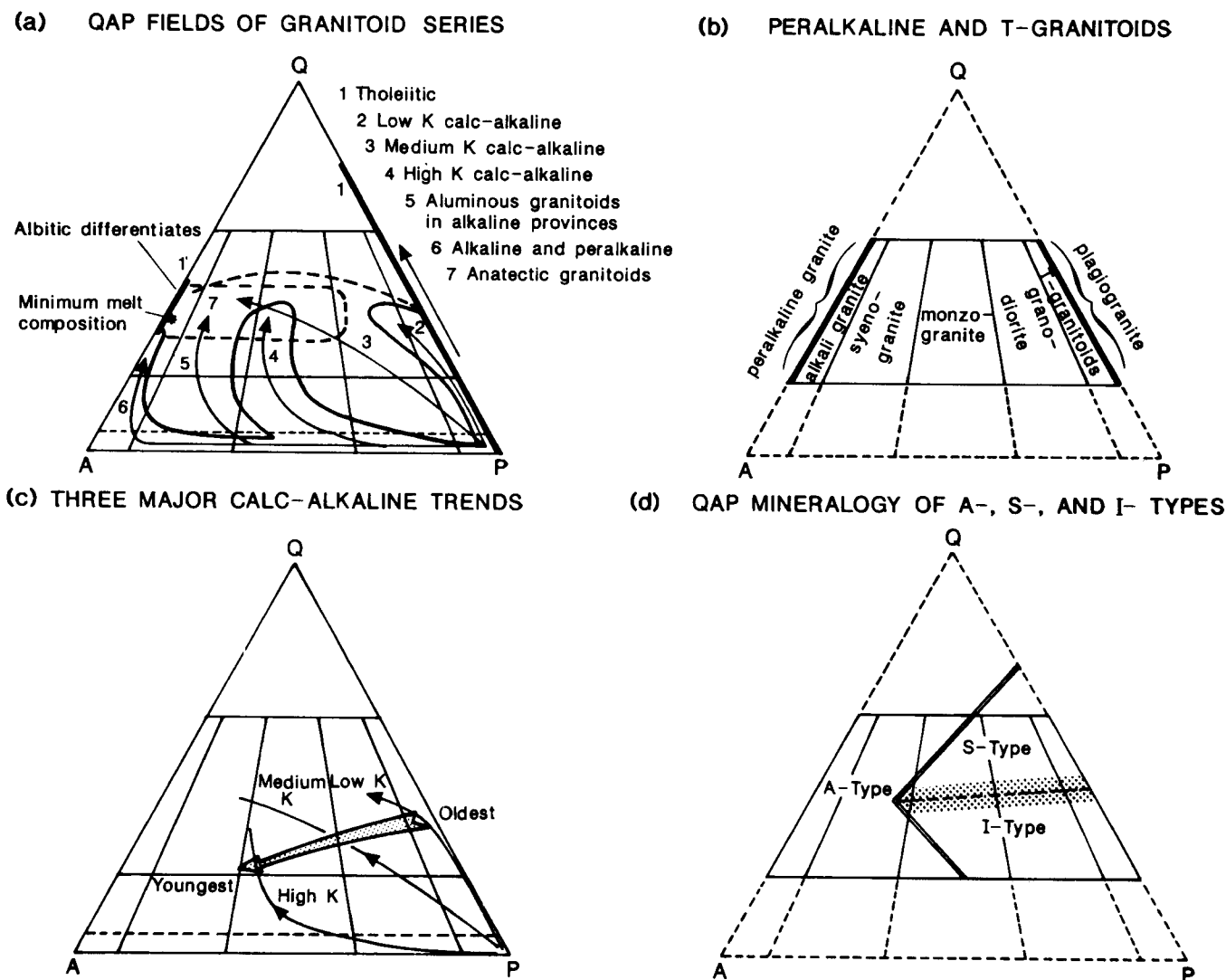
(a) QAP FIELDS OF GRANITOID SERIES

1 Tholeiitic
2 Low K calc-alkaline
3 Medium K calc-alkaline
4 High K calc-alkaline
5 Aluminous granitoids in alkaline provinces
6 Alkaline and peralkaline
7 Anatectic granitoids

Albitic differentiates

Minimum melt composition

(b) PERALKALINE AND T-GRANITOIDS

peralkaline granite
alkali syeno-granite
monzo-granite
grano-diorite
plagiogranite
T-granitoids

(c) THREE MAJOR CALC-ALKALINE TRENDS

Medium K
Low K
Oldest
Youngest
High K

(d) QAP MINERALOGY OF A-, S-, AND I- TYPES

A-Type
S-Type
I-Type

Figure 10. (a) QAP fields and differentiation trends for seven distinctive plutonic rock series.

(b) QAP fields of peralkaline granites, plagiogranite and T-granitoids (tonalite/trondhjemite).

(c) Spatially related plutonic rock series show clear mineralogical and chemical correlations with tectonic environment and relative time of emplacement e.g. evolution of calc-alkaline series in orogenic belts from the oldest (low-K tonalitic series) through the medium-K granodioritic series to the youngest (high-K monzonitic series).

(d) Fields of the QAP plot typically occupied by I-, S- and A-type granitoids (after Bowden et al., 1984).

inclusions (microgranitoid enclaves), whereas I-types contain hornblende-rich, mafic inclusions of igneous appearance. Chappell and White interpret these inclusions as "restite", residual source material. Linear inter-element variation trends are regarded as due to restite unmixing. Alternative interpretations involving magma mixing have been suggested, but are not relevant to the present topic. This first order classification based on source rock has been extended to include M-type (mantle-derived) granitoids and A-type (anorogenic, alkaline, anhydrous and, somewhat cynically, "ambiguous") granitoids, with distinctive characteristics. A-type granitoids are inferred to be derived by partial melting of F and/or Cl-enriched dry granulitic residue remaining in the lower crust after earlier extraction of an orogenic granitic melt (Whalen et al., 1987).

Selected characteristic features of these four granitoid types can be drawn from Pitcher (1983) and Bowden et al. (1984). They include:

I-type: metaluminous; calc-alkaline to alkali-calcic, relatively quartz-poor monzogranites, granodiorites and tonalites; 53% to 76% SiO_2; high Na/K, high Ca for mafic varieties; hornblende-bearing (except most felsic members).

S-type: strongly peraluminous; alkali-calcic to calc-alkaline, relatively quartz-rich monzogranites, granodiorites and tonalites; 65% to 74% SiO_2; low Na/K, Ca and Sr; with peraluminous minerals (muscovite, cordierite, garnet or andalusite); often biotite-rich.

A-type: peralkaline to metaluminous; alkalic to alkali-calcic syenogranites, alkali granites and quartz syenites; mostly 70% to 78% SiO_2; high Na+K, Fe/Mg, F+Cl and low Ca, Sr; accessory minerals such as fayalite, hedenbergite, ferrohastingsite, annite, fluorite, sodic pyroxenes, perthitic or rapakivi-textured feldspars.

M-type: metaluminous; calcic gabbros, diorites, quartz diorites, tonalites and plagiogranites; 45% to 78% SiO_2; little or no K-feldspar.

Refinements of the I-S classification have been suggested. Pitcher (1983) recognised I-Cordilleran and I-Caledonian granitoids, each with distinctive composition and mineralogy, based on their tectonic setting. His data suggest that the tonalite-dominant I-Cordilleran granitoids tend to be relatively magnetite-rich, whereas the granodiorite-dominant I-Caledonian type granitoids tend to have less magnetite. Chappell and Stephens (1988) proposed that

(a)   DEFINITION OF PEACOCK'S ALKALI-LIME INDEX
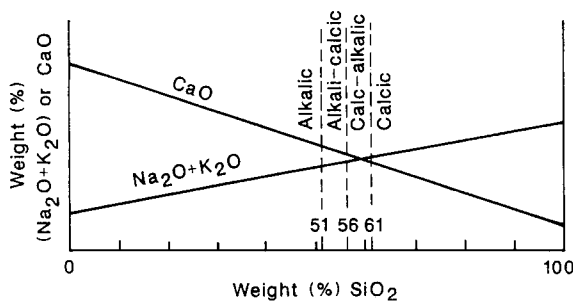
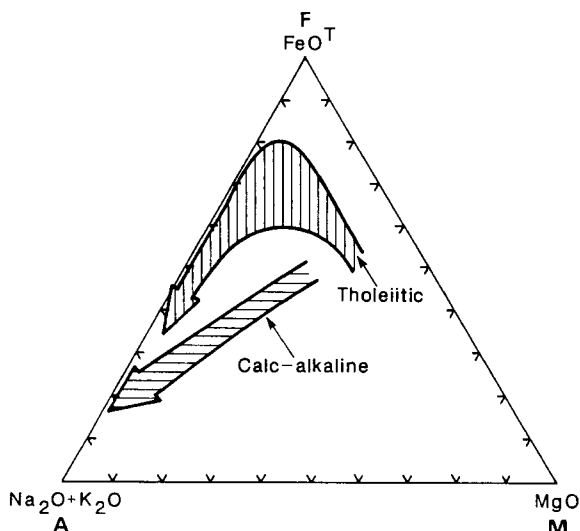(b) CONTRASTING THOLEIITIC AND CALC-ALKALINE TRENDS

Figure 11. (a) Classification of comagmatic igneous rock series as alkalic, alkali-calcic, calc-alkalic and calcic on the basis of the alkali-lime index (Peacock, 1931).

(b) Contrasting trends on the ternary plot of alkalis, total iron and MgO (AFM diagram) for differentiating tholeiitic and calc-alkaline magmas.

progressively more felsic and chemically evolved I-type granitoids result from successive remelting of older mafic rocks that have underplated the crust. M-types comprise gabbros to mafic granites derived directly from the mantle or mantle wedge, I-tonalite types are derived from fusion of M-type material and I-granodiorite types represent remagmatised products of I-tonalite rocks. This classification allows for both I-tonalite and I-granodiorite types to occur in the one tectonic setting, although typically one subtype will predominate.

Granitoid classification based on tectonic setting

Pitcher (1983) has related other granite classification schemes to tectonic environment. Maniar and Piccoli (1989) have proposed an independent granitoid classification scheme, based on tectonic setting. A first order orogenic category is subdivided into island arc granitoids (IAG), continental arc granitoids (CAG), continental collision granitoids (CCG) and post-orogenic granitoids (POG). Anorogenic granitoids fall into three categories: Rift-Related Granitoids (RRG); Continental Epeirogenic Uplift Granitoids (CEUG); and Oceanic Plagiogranites (OP). Although the occurrence and abundance of magnetite was not noted by these authors, the detailed information on chemistry and mineralogy allows broad conclusions on likely magnetite contents to be inferred by comparison with other studies.

Granitoid classification based on Fe-Ti oxide mineralogy

Ishihara (1977) instigated a descriptive classification of calc-alkaline granitoids into a magnetite-series and an ilmenite-series, based on their characteristic iron-titanium oxide mineralogy. This classification can be directly related to magnetic properties and has important exploration implications, because of the association between metallogeny and the magnetite-series/ilmenite-series classification (Ishihara, 1981). The characteristic accessory mineralogies of the two categories of granitoid are:

magnetite-series — 0.1-2 vol % magnetite ± ilmenite; plus haematite, pyrite, sphene, oxidised Mg-rich biotite; and

ilmenite series — magnetite absent, ilmenite (< 0.1 vol %) + pyrrhotite, graphite, muscovite, reduced Fe-rich biotite.

Thus magnetite-series granitoids are ferromagnetic (MFM to SFM), with susceptibilities in the approximate range $3800\text{-}75,000 \times 10^{-6}$ SI (300-6000 µG/Oe), whereas ilmenite-series granitoids are paramagnetic. Pyrrhotite is present in ilmenite-series granitoids in very minor amounts and cannot contribute significantly to the susceptibility, particularly since much of the pyrrhotite present in ilmenite-series granitoids is the hexagonal variety (Whalen and Chappell, 1988), which is weakly magnetic. Magnetite-series granitoids are significantly more oxidised than ilmenite-series granitoids. This is thought to reflect upper mantle/lower crustal generation of the magnetite-series, involving minimal interaction with carbonaceous material, whereas the ilmenite-series is interpreted to have been generated in the middle to lower crust and to be significantly contaminated by C-bearing crustal rocks.

Fershtater et al. (1978) and Fershtater and Chashchukhina (1979) have devised a "Ferrofacies Classification" of granitoids. The ferrofacies concept has been applied to a wide range of granitoids from the former USSR and can be regarded as an extension of the magnetite-series and ilmenite-series classification. The categories in that classification are: the magnetite ferrofacies; the magnetite-bearing ferrofacies; the magnetite-"free" ferrofacies; and the titanomagnetite ferrofacies — each with distinctive mineralogical characteristics. This classification does not appear to have been used by other workers, but may form the basis for a refined magnetic petrological classification with metallogenic implications.

Suites, supersuites and basement terranes

Hine et al. (1978) showed that granitoids of the Lachlan Fold Belt can be grouped into suites using petrographic, chemical and isotopic criteria. Members of a suite are interpreted to be derived from similar source rocks. Suites with similar character can be grouped into supersuites. Chappell et al. (1988) demonstrated that granitoids within specific provinces tend to exhibit common geochemical character. Since the compositions of the granitoids largely reflect compositions of their source regions, the distribution of granitoid suites and supersuites can be used to define terranes, within each of which the lower crust has distinctive geochemical characteristics. These basement terranes are often poorly correlated with the tectonostratigraphic terranes that are defined from the surface geology.

GEOLOGICAL FACTORS THAT CONTROL MAGNETISATION OF INTRUSIONS

Iron content and oxidation ratio

Many petrological studies of intrusive igneous rocks have been made that are relevant to the problem of defining the geological controls on magnetic properties in these rocks. To a good approximation, the magnetic susceptibility of
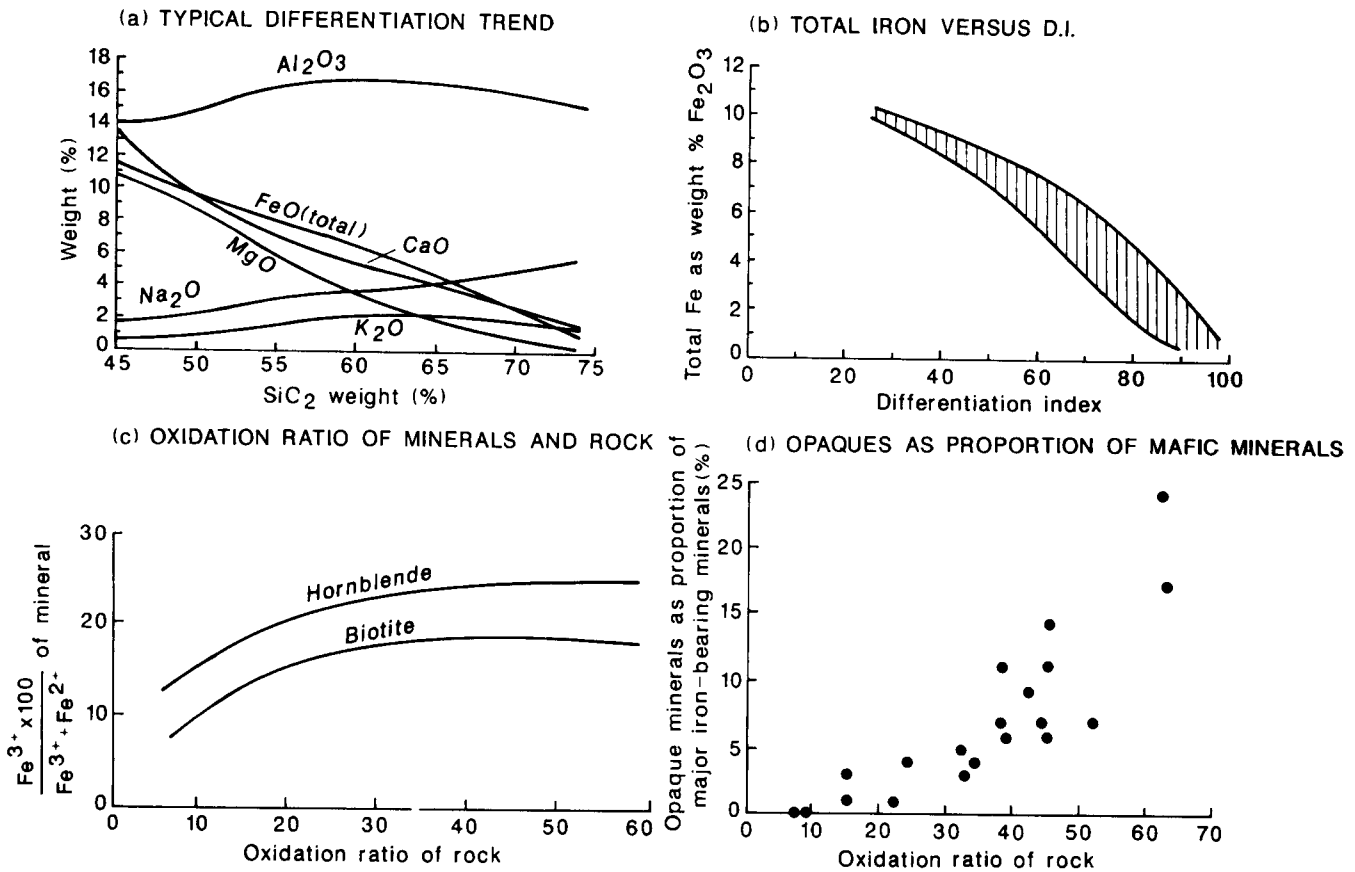
(a) TYPICAL DIFFERENTIATION TREND

Weight (%)

Al_2O_3

$FeO_{(total)}$

MgO

CaO

Na_2O

K_2O

SiO_2 weight (%)

(b) TOTAL IRON VERSUS D.I.

Total Fe as weight % Fe_2O_3

Differentiation index

(c) OXIDATION RATIO OF MINERALS AND ROCK

$\frac{Fe^{3+} \times 100}{Fe^{3+} + Fe^{2+}}$ of mineral

Hornblende

Biotite

Oxidation ratio of rock

(d) OPAQUES AS PROPORTION OF MAFIC MINERALS

Opaque minerals as proportion of major iron-bearing minerals (%)

Oxidation ratio of rock

Figure 12.(a) Harker diagram showing typical differentiation trends for major elements in a comagmatic igneous rock suite.
(b) Total iron versus differentiation index for a suite of granitoids from the Sierra Nevada Batholith (after Dodge, 1972).
(c) Oxidation ratio (%) for hornblende and biotite from the Sierra Nevada granitoids versus oxidation ratio (%) of whole rock.
(d) Opaque mineral (essentially magnetite) contents as proportion of total iron-bearing mineral assemblage of the Sierra Nevada granitoids versus oxidation ratio of rock.

intrusive igneous rocks is simply proportional to their magnetite content. The directly relevant chemical parameters are the total iron content of the rock, which constrains the theoretical maximum attainable susceptibility, and the oxidation ratio (ferric/total iron), which essentially determines the partitioning of iron between silicates and oxides (mainly magnetite, in fresh igneous rocks). Figure 12(a) shows the typical trend for major elements with increasing silica for a series of igneous rocks derived from a basic parental magma. Total iron tends to decrease steadily, but it is important to note that even the most felsic members of common igneous rock series would contain sufficient iron to make them at least moderately to strongly ferromagnetic, provided that all the iron was contained in magnetite. Figure 12(b) gives an example of a total iron versus differentiation index trend for a comagmatic suite of granitoids, showing that even the most evolved members of this suite have at least 0.5 wt %, and generally more than 1 wt %, total iron.

Much of the iron, however, is always sequestered within paramagnetic silicate minerals. If the rocks are paramagnetic, the susceptibility decreases monotonically with increasing silica content. This occurs if the iron oxidation ratio of the rocks is low, particularly in the more evolved rocks. In that case, silicates take up the predominantly ferrous iron and the relatively small amounts of ferric iron can also be accommodated in silicates, mainly in hydrous phases. As the Fe oxidation ratio of the rocks increases, the silicates are obliged initially to take up more ferric iron. Once the oxidation ratio exceeds the maximum amount of ferric iron that can be accommodated in silicates, the excess ferric iron is forced to appear as magnetite.

Figure 12(c) shows, for the same suite of granitoids considered in Figure 12(b), how the Fe oxidation ratios of hornblende and biotite are correlated with oxidation ratio of the whole rock, indicating that these phases start to become saturated with ferric iron at rock oxidation ratios above ~20%. Maximum ferric iron contents in these silicates are attained when the oxidation ratio of the rock is ~30%. When this ratio is exceeded, there is a steady increase in magnetite content, until it constitutes ~20% of the mafic minerals, as the oxidation ratio increases up to ~70%. Above this value, the whole rock ferric iron would be in surplus for forming magnetite, especially when the large proportion of ferrous iron in silicates is considered, and haematite or maghemite would be present in addition to magnetite.

In mafic anhydrous rocks without amphibole or mica, however, the anhydrous silicates can accommodate much less ferric iron than hornblende and biotite, and magnetite appears in such rocks at lower oxidation ratios. This explains why many gabbros and norites are strongly magnetic, in spite of lower oxidation ratios than for the granitoids considered in Figure 12(b)-(d).

Data on intrusive rocks from Finland, taken from Puranen (1989), are plotted in Figure 13. Figure 13(a) shows the average total iron, oxidation ratio, and percentage of ferromagnetic rocks for compositions ranging from gabbro to granite. There is a systematic increase in oxidation ratio with silica content, offsetting the effect on susceptibility of the decrease in total iron. This produces an increased proportion of ferromagnetic rocks at the felsic end of the spectrum, with a slight increase in average susceptibility for granites, compared to granodiorites, as a result (see Figure
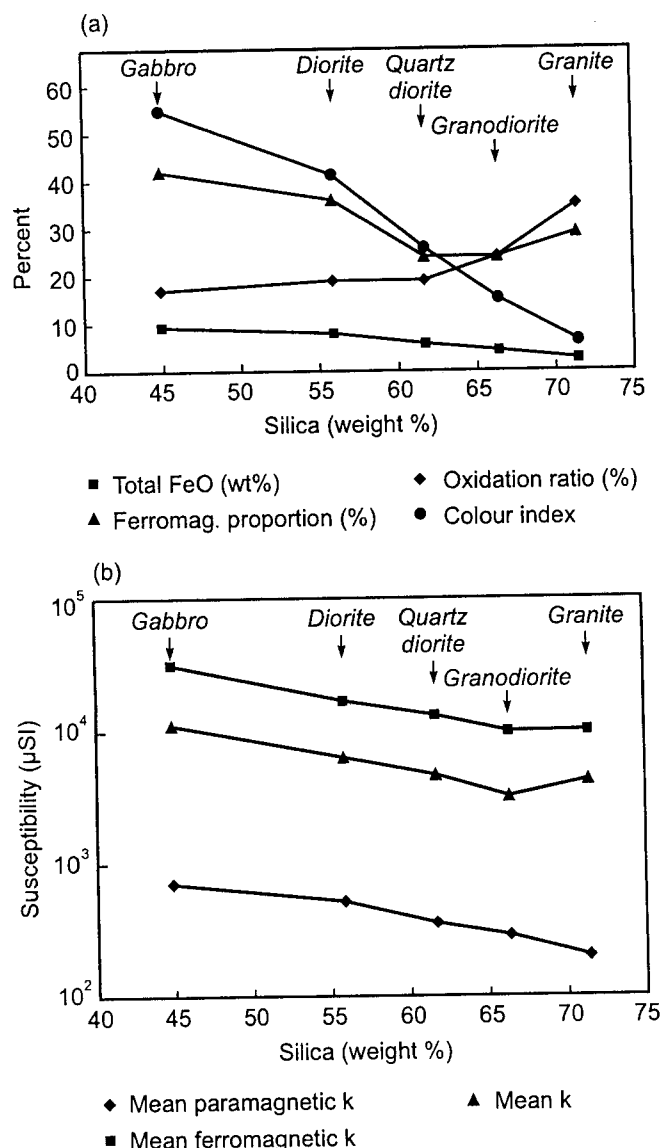
(a)

(b)

Figure 13.(a) Total iron as FeO (wt %), oxidation ratio (%), and percentage of the total sample for each rock type that is ferromagnetic versus SiO_2 for Finnish gabbros, diorites, granodiorites and granites. (b) Mean susceptibility of paramagnetic subpopulations, mean susceptibility of ferromagnetic subpopulations, mean susceptibility of total population, and colour index (volume % mafic minerals) versus SiO_2 for Finnish gabbros, diorites, granodiorites and granites (data from Puranen, 1989).

13(b)). Note that these data, which are derived from a very large petrophysical sampling program by the Finnish Geological Survey, refer to all sampled units within the appropriate QAP field, irrespective of geological setting, metamorphic grade, varietal mineralogy etc. Although systematic trends tend to be smoothed out by this geological and petrological variability, a clear correlation between chemistry and magnetic petrology is still evident. This indicates that there are strong underlying trends, when specific geological provinces, tectonic settings, geochemical characteristics or mineralogical varieties are considered.

Blevin (1994, 1996) has analysed a large collection of samples for relationships between susceptibility, oxidation state, granitoid type and composition. There is a distinct susceptibility gap between two main trends in the susceptibility-SiO_2 plot. The ferromagnetic trend, representing magnetite-bearing granitoids, exhibits a gradual decrease in susceptibility in both the maximum and average suscepti-

bility values with increasing SiO_2, up to ~72 wt % SiO_2, and then plunges rapidly at higher silica contents. The paramagnetic trend lies two to three orders of magnitude below the ferromagnetic trend and also exhibits a gentle systematic decrease in susceptibility with increasing SiO_2. Oxidised granitoids generally have susceptibilities greater than 2000×10^{-6} SI (160 µG/Oe), with a maximum of 80,000 $\times 10^{-6}$ SI (~ 6000 µG/Oe), whereas reduced granitoids have susceptibilities ranging from ~ 500×10^{-6} SI (~ 40 µG/Oe) at the low silica end to ~ 130×10^{-6} SI (~ 10 µG/Oe) at the highest silica contents.

For a given silica content, which generally implies similar total iron contents, there is pronounced increase in susceptibility with increasing oxidation ratio. Granitoids that plot in the gap between the main ferromagnetic and paramagnetic trends are either so felsic ($SiO_2 > 72$ weight %) that the iron content is too low to crystallise significant magnetite, irrespective of oxidation state, or show evidence of alteration of magnetite. There is little correlation between susceptibility, SiO_2 and Fe_2O_3 /FeO for the latter group of granitoids, indicating that the processes of magnetite alteration are not systematically related to granitoid composition.

Geochemical and mineralogical associations with magnetite

The clearest correlations between geochemical or mineralogical factors in granitoids and magnetite content are the general increase in magnetite abundance with increasing oxidation ratio (except for the most oxidised haematite-bearing rocks), for a given iron content, and the increase in maximum magnetite content with increasing total iron, for a given oxidation ratio. However, the occurrence and abundance of magnetite is clearly correlated with other geochemical characteristics. Metaluminous granitoids are much more likely to be ferromagnetic than peraluminous or peralkaline granitoids, and igneous rocks with extreme alumina saturation are almost always paramagnetic. Within each Ishihara series, there is a general correlation of decreasing susceptibility with increasing silica content.

Hornblende + pyroxene or olivine (except fayalite) in mafic varieties is favourable for the presence of magnetite, as is hornblende + biotite in more felsic rocks. Fayalitic olivine, however, indicates reducing conditions and is found only in magnetite-poor granitoids. Mg-rich hornblende and biotite indicate relatively oxidising conditions, with removal of iron into magnetite and consequent enrichment of the mafic silicates in magnesium, particularly when the Mg/Fe ratio increases with increasing rock SiO_2. When ilmenite is present, its composition is correlated with magnetite content. Granitoids without magnetite have relatively reduced ilmenite (< 8 mol % Fe_2O_3), whereas magnetite-bearing granitoids have either more oxidised ilmenite, or Mn-rich ilmenite. High Mn-ilmenite is favoured by oxidising conditions because Fe is preferentially incorporated into magnetite rather than ilmenite.

Figure 14(a) shows the correlation between opaque mineral content and susceptibility for Japanese granitoids and gabbroic rocks, showing an essentially proportional relationship (Ishihara, 1981). This reflects the dominance of magnetite over other opaque phases in magnetite-series plutons and the proportionality of susceptibility and magnetite content for normal ferromagnetic rocks. Figure 14(b) indicates that there is a wide range of susceptibilities for mafic magnetite-series granitoids, with many strongly ferromagnetic examples, but the maximum magnetite content, and hence the maximum susceptibility, decreases linearly with increasing quartz + K-feldspar, so that the most felsic members of the series (syenogranites) are only weakly
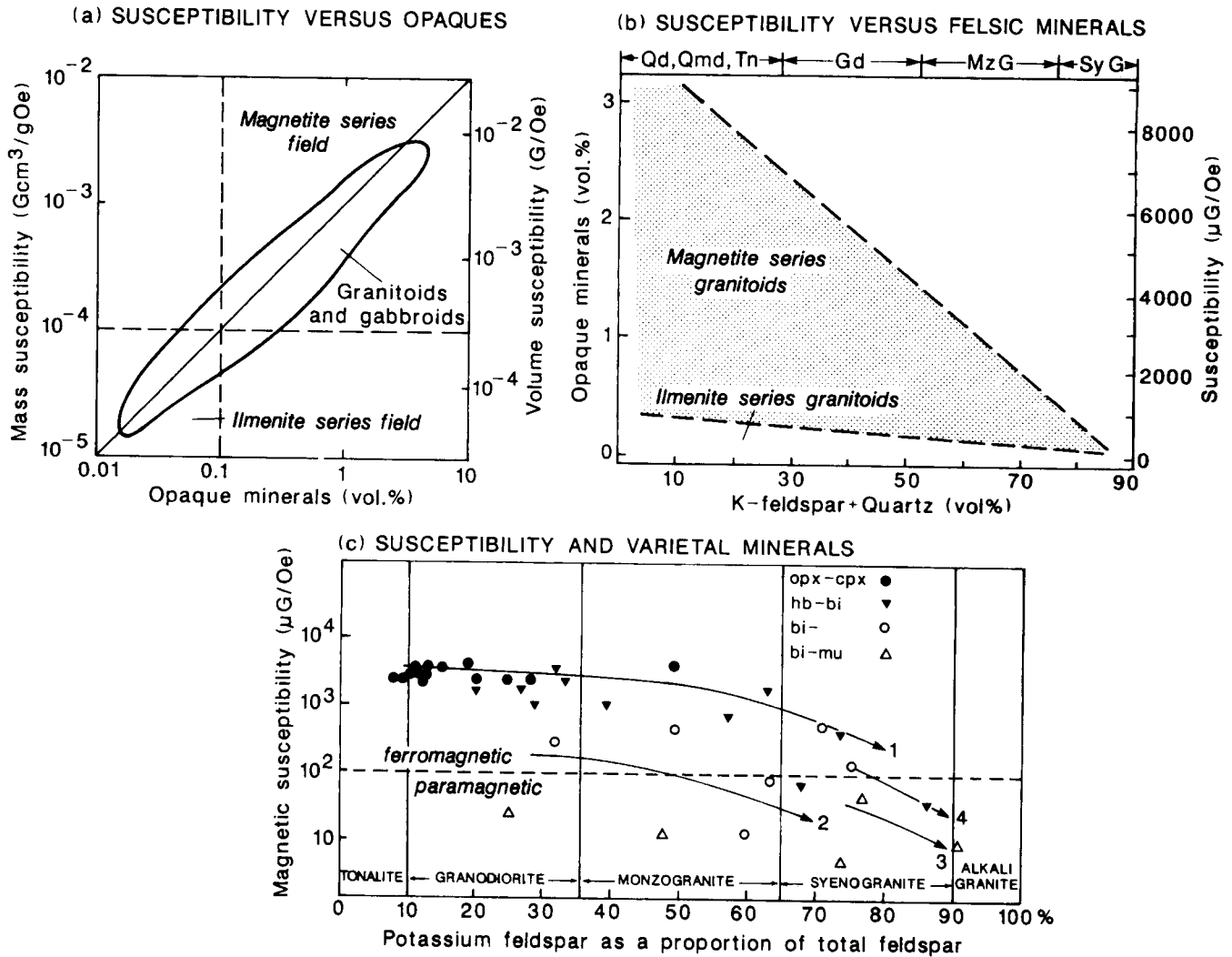
(a) SUSCEPTIBILITY VERSUS OPAQUES

Magnetite series field

Granitoids and gabbroids

Ilmenite series field

Opaque minerals (vol.%)

(b) SUSCEPTIBILITY VERSUS FELSIC MINERALS

Qd,Qmd,Tn — Gd — MzG — Sy G

Magnetite series granitoids

Ilmenite series granitoids

K-feldspar+Quartz (vol%)

(c) SUSCEPTIBILITY AND VARIETAL MINERALS

opx-cpx ●
hb-bi ▼
bi- ○
bi-mu △

ferromagnetic

paramagnetic

TONALITE — GRANODIORITE — MONZOGRANITE — SYENOGRANITE — ALKALI GRANITE

Potassium feldspar as a proportion of total feldspar

Figure 14. (a) Correlation between opaque mineral content and susceptibility for Japanese granitoids and gabbroids, showing an essentially proportional relationship (after Ishihara, 1981). Ilmenite-series granitoids have very low opaque mineral contents (< 0.1 vol %).
(b) Range of susceptibilities for magnetite-series and ilmenite-series granitoids versus quartz + K-feldspar (after Ishihara, 1981).
(c) Relationships between susceptibility, lithology and varietal mineralogy for granitoids from central Australia (after Mutton and Shaw, 1979).

to moderately ferromagnetic. Ilmenite-series granitoids have very low opaque mineral contents (< 0.1 vol %) and there is still a distinctly lower average susceptibility for the ilmenite-series syenogranites than for their magnetite-series equivalents. There is also a general trend to decreasing paramagnetic susceptibility in more felsic ilmenite-series granitoids, as expected.

Figure 14(c) shows the relationships between susceptibility, lithology and varietal mineralogy for granitoids from central Australia (Mutton and Shaw, 1979). This confirms the decrease in maximum susceptibility for more felsic rocks, and the association of magnetite with pyroxene and hornblende (indicative of M- or I-type affinities) and the apparent antipathetic relationship of magnetite with muscovite, which is a mineral characteristic of peraluminous, usually S-type, granitoids.

Blevin (1994, 1996) has established a useful relationship between feldspar colour and susceptibility for calc-alkaline granitoids. For granitoids with white plagioclase there is a distinct increase in average oxidation ratio and susceptibility with increasing pinkness of K-feldspar. Salmon pink K-feldspars tend to be most oxidised and have the highest susceptibilities. White K-feldspar indicates a reduced rock with low susceptibility. Brick red K-feldspars, on the other hand, which are most common in very felsic rocks, indicate hydrothermal alteration and are generally associated with

lower susceptibilities than pink K-feldspars. Green plagioclase is generally indicative of alteration that tends to be magnetite-destructive and is correlated with variable, generally lower, susceptibilities that are poorly correlated with rock composition and oxidation state. Yellowish feldspars usually indicate weathering and such samples are not representative of the fresh rock. Overall, there is a reasonably predictable relationship between susceptibility and the field-observable features: colour index (percentage of mafic minerals), which provides a proxy estimate of silica and iron contents, and K-feldspar colour (provided the plagioclase is white).

Source rock

Whalen and Chappell (1988) showed that most I-type granitoids of the Lachlan Fold Belt are magnetite-series and most S-types are ilmenite-series, although exceptions to the rule are found. Blevin (1994, 1996) has shown that ~80% of I-type granitoids from the Lachlan and New England Fold Belts have susceptibilities greater than 1000 × 10^{-6} SI (80 μG/Oe), mostly greater than 2000 × 10^{-6} SI (160 μG/Oe), whereas nearly all S-types have susceptibilities less than 1000 × 10^{-6} SI (80 μG/Oe). Exceptions to these generalisations occur within specific suites or supersuites and are confined to particular basement terranes.
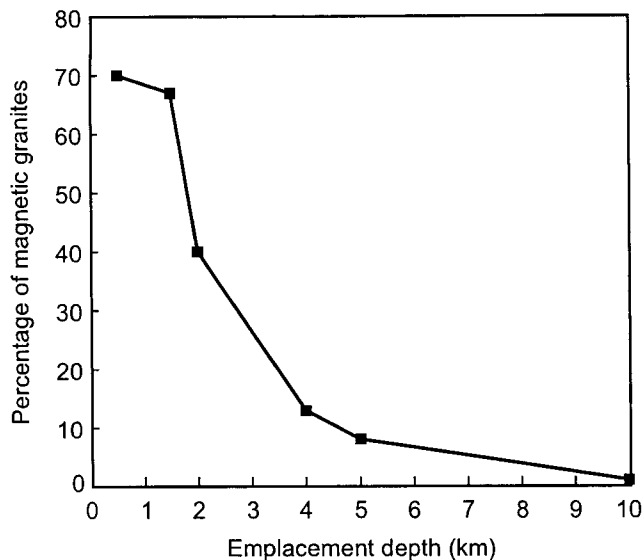
Figure 15. Proportion of ferromagnetic granitoids versus estimated emplacement depth for granitoids from NE Russia (Pecherskiy, 1965).

Blevin (1994, 1996) has also shown that within each granitoid suite there is generally a systematic decrease in susceptibility with increasing SiO_2. This decrease becomes very rapid above ~74 wt % SiO_2. M-types from the southwest Pacific have the highest susceptibilities in this extensive collection of granitoid rocks from eastern Australia and Oceania, unless they are altered. Both carbonate and pyrite-pyrrhotite-chalcopyrite alteration are magnetite-destructive in the M-type rocks. For silica contents of 70% to 74% by weight, A-type granitoids exhibit a bimodal distribution of susceptibilities, similar to those of oxidised and reduced I-types with equivalent SiO_2. Above ~74 wt % SiO_2 the A-types exhibit a broad unimodal susceptibility distribution, reflecting a rapid decrease from WFM to PM levels as SiO_2 increases from 74 wt % to 78 wt%.

Average susceptibilities for M-, I- and S-type granitoids are $40,000 \times 10^{-6}$, 8900×10^{-6} and 410×10^{-6} SI (3200 μG/Oe, 700 μG/Oe and 30 μG/Oe) respectively. The corresponding medians are $32,500 \times 10^{-6}$, 5600×10^{-6} and 270×10^{-6} SI (2600 μG/Oe, 450 μG/Oe and 20 μG/Oe). The median susceptibility of the limited set of A-type granitoids studied by Blevin (1994) is ~ 1000×10^{-6} SI (~ 80 μG/Oe). For the ferromagnetic subpopulation of the slightly less felsic (70 wt % to 74 wt % SiO_2) varieties of A-type, the median susceptibility is an order of magnitude greater.

Magnetite contents of granitoids show distinct provinciality, along with other mineralogical and chemical characteristics, reflecting distinctive compositions of lower crustal source regions (Chappell et al., 1988; Blevin, 1994). For example, in most basement terranes I-type granitoids are relatively oxidised magnetite-series rocks. In the Melbourne Basement terrane, however, the I-type granitoids are reduced and belong to the ilmenite-series. The infracrustal protolith from which these rocks have been derived is therefore inferred to be more reduced than elsewhere in the Lachlan Fold Belt. Granitoids belonging to individual suites, which are derived from fairly homogeneous source rocks, exhibit a systematic correlation between magnetic susceptibility and composition that is much better defined than global relationships between these variables.

Overall, mantle-derived granitoids, I-types derived from mafic crustal underplates and second-generation I-types derived from oxidised I-type source rocks are magnetite-series, whereas I-types derived from reduced igneous rocks
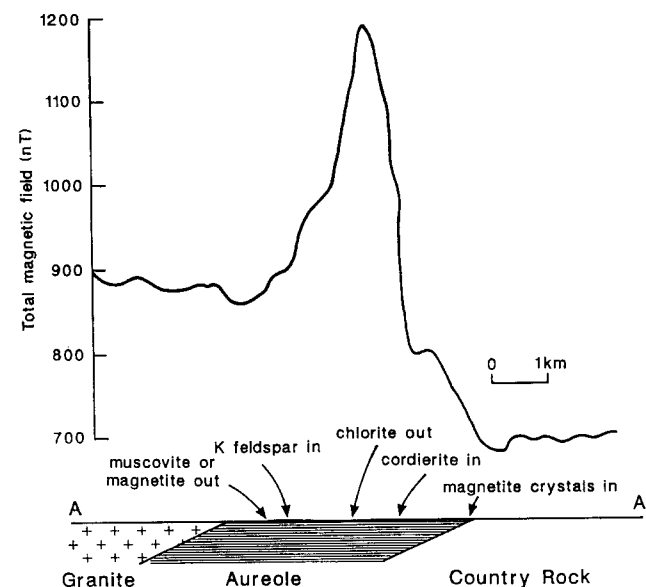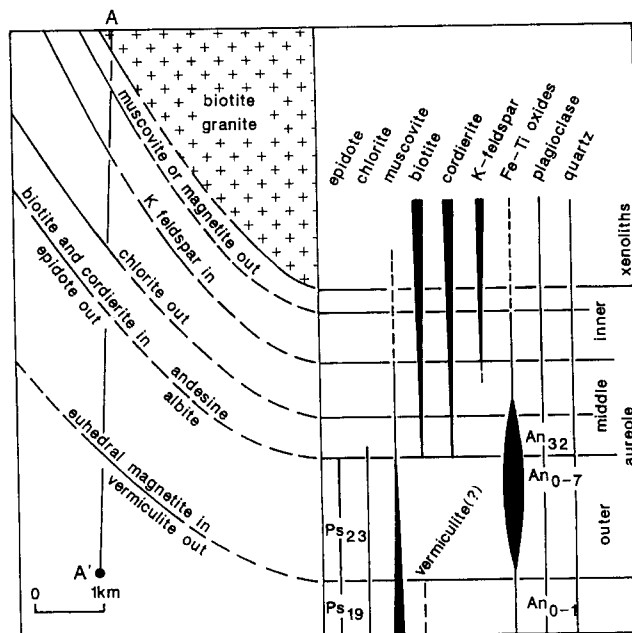

Figure 16. Magnetic expression of metamorphic magnetite formation in the contact aureole of the Liberty Hill pluton, South Carolina (after Speer, 1981).

are ilmenite-series. A-types resemble felsic I-types and have subequal magnetite-series and ilmenite-series populations. Most S-types are reduced ilmenite series granitoids, probably reflecting presence of carbon in their lower to middle crustal source material.

Lithology

The overall proportion of ferromagnetic rocks within a given geological province or within a particular igneous rock series decreases from gabbro through to granite. This trend is apparent from Figure 13(a), which combines data from a wide range of areas and rock series, but is more clearly expressed within particular provinces or rock series. Mafic to felsic and intermediate to felsic associations are much more likely to be magnetite-series throughout, than compositionally restricted felsic associations. Alkaline intrusive rocks are often magnetite-series, with the exception of extreme compositions, such as peralkaline granites and agpaitic (peralkaline, undersaturated) nepheline syenites. In
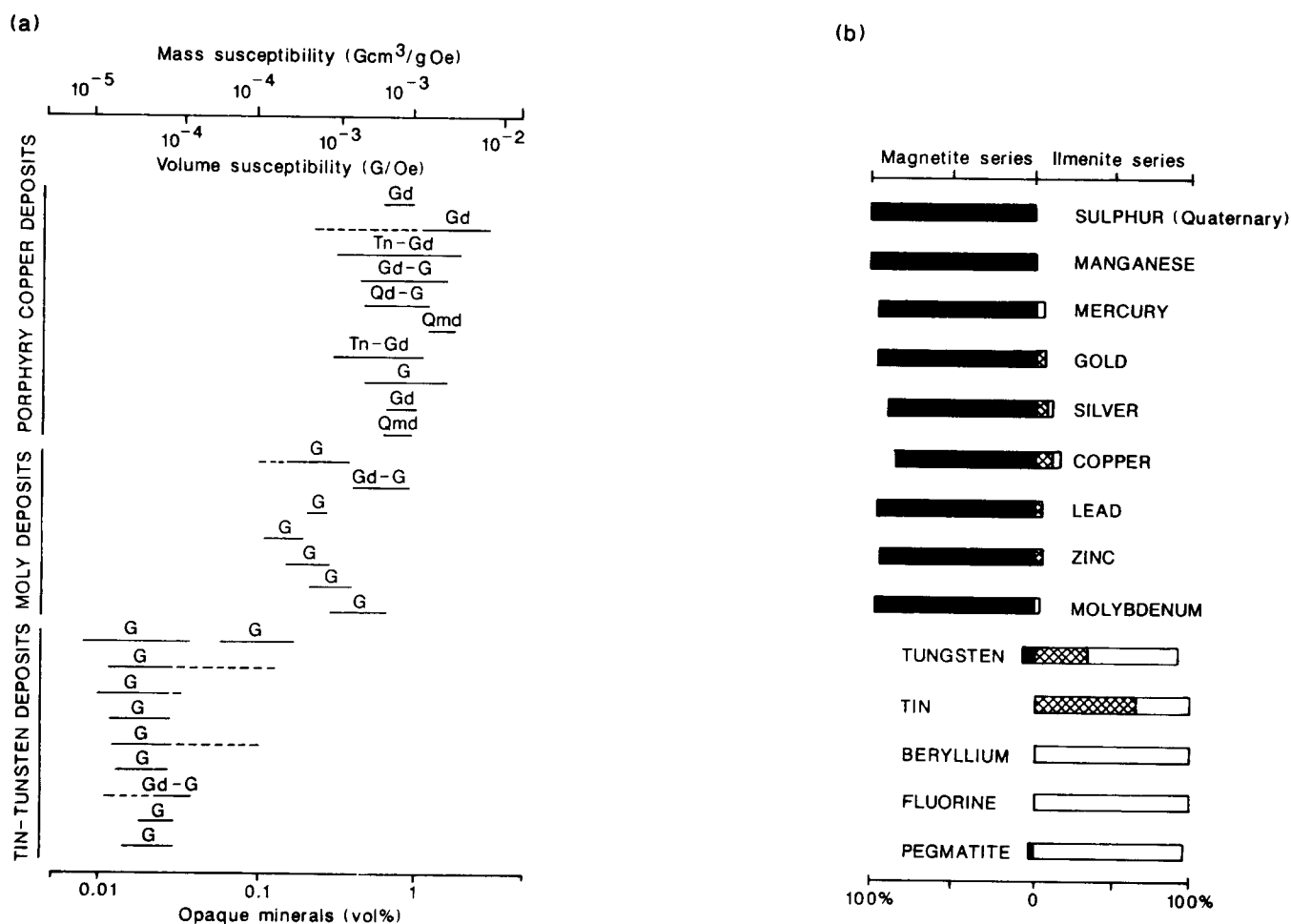
Figure 17. (a) Range of CGS mass and volume susceptibilities and opaque mineral contents for granitoids associated with porphyry Cu, granitoid-related Mo deposits and granitoid-related Sn-W deposits (Gd = granodiorite, Tn = tonalite, G = granite, Qmd = quartz monzodiorite).
(b) Proportions of mineral deposits, of a variety of commodities that occur within magnetite-series and ilmenite-series granitoid belts. Hatched regions represent WFM magnetite-series granitoids. The mineral deposits are inferred to be genetically related to granitoids or to their associated volcanics. Pegmatite refers to stanniferous pegmatite deposits. SI volume susceptibility = CGS volume susceptibility (G/Oe) × 4π; mass susceptibility = volume susceptibility/density (after Ishihara, 1981).

tholeiitic layered complexes, less evolved lower gabbros are paramagnetic to weakly ferromagnetic, whereas sufficiently evolved upper ferrogabbros and ferrodiorites, and associated granophyres, are usually strongly ferromagnetic.

Emplacement depth

Pecherskiy (1965) noted a strong correlation between shallow emplacement depth and occurrence of magnetite for a wide variety of granitoids in northeastern Russia. Figure 15 shows the percentage of moderately and strongly ferromagnetic granitoids versus estimated depth of emplacement for a large number of plutons. There is a systematic increase in ferromagnetic proportion with decreasing depth of emplacement. The ferromagnetic proportion rises to 70% for subvolcanic/epizonal granitoids. This interesting observation does not appear to have attracted much attention, but other studies lend some indirect support. Czamanske et al. (1981) explain a similar correlation in Japan by invoking onset of second boiling in the residual melt in epizonal plutons. Dissociation of water and preferential diffusion of hydrogen out of the pluton into fractured country rock is the oxidation process that is postulated to produce the high magnetite contents of these plutons. However, Candela (1986) has shown that dissociation of water can only be an important oxidising process for iron-poor ($<<1$ wt % FeOT) granitoids. This mechanism may operate in Climax-type

Mo porphyries, which are very felsic, but interaction with oxygenated meteoric waters is a more probable explanation for the relatively oxidised nature of at least some epizonal granitoids.

There is also a general correlation between the source rock, depth of generation and depth of emplacement of granitoids, which probably explains much of the empirical trend shown in Figure 15. Deep-seated, high temperature, anhydrous magmas rise to shallow crustal levels, whereas lower temperature, hydrous magmas (produced by partial melting of muscovite-rich pelitic metasediments, for example) do not rise very far from their source regions, producing catazonal granitoids. The former type of magma is more likely to produce magnetite-series granitoids, whereas the latter generally produces ilmenite-series granitoids, for reasons already explained.

Tectonic setting

Referring to Maniar and Piccoli's (1989) classification, most island arc and oceanic plagiogranites, and more mafic continental arc granitoids, are ferromagnetic. Nearly all continental collision and post-orogenic granitoids are paramagnetic. Rift-related granitoids and continental epeirogenic uplift granitoids have an inferred bimodal distribution of susceptibilities, with the mafic compositions tending to be ferromagnetic and the felsic compositions generally paramagnetic.
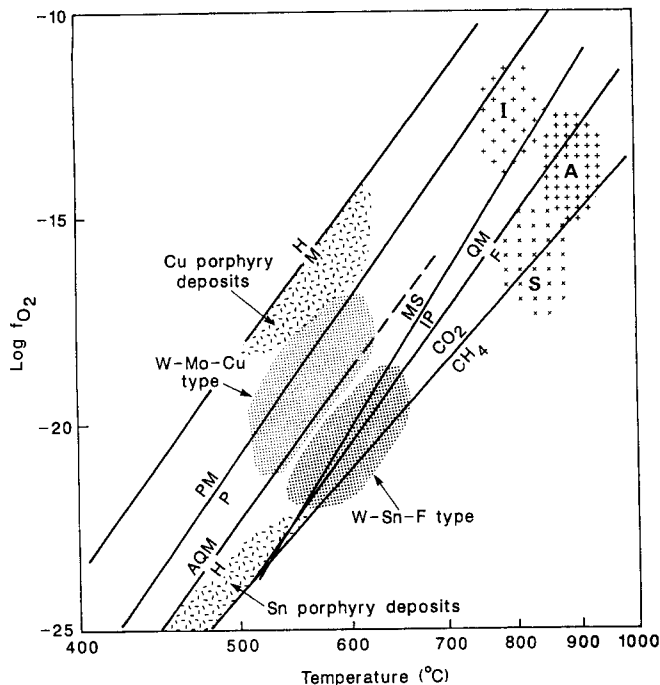
Figure 18. Oxygen fugacity versus temperature fields for typical I-, S- and A-type granitoids and for porphyry Cu, W-Mo-Cu skarn, W-Sn-F skarn and porphyry Sn mineralisation, together with a number of standard oxygen fugacity buffers (after Kwak and White, 1982). Mineral and fluid oxygen fugacity buffers that may be important controls on the magnetic mineralogy of the igneous intrusions and their associated mineralisation include (CO_2-CH_4; PMP = pyrite-magnetite-pyrrhotite; AQMH = andradite-quartz-magnetite-hedenbergite; MSIP = magnetite-sphene-ilmenite-pyroxene) as well as the FMQ and HM buffers.

Crustal contamination and contact aureoles

Ishihara et al. (1987) presented evidence that local contamination of a I-type tonalitic pluton by sulphur and carbon derived from sedimentary country rocks produced a vertical zonation from low magnetic susceptibility ilmenite-series tonalite at lower levels to magnetite-series tonalite at higher levels. There was less contamination at higher levels. Similar effects have also been observed in the Lachlan Fold Belt, but only within a few metres of the granitoid margin (Blevin, 1994). More generally, Blevin (1994) argues that crustal contamination effects on oxidation state of Lachlan Fold Belt granitoids are negligible and that in the vast majority of cases the oxidation state of these granitoids is inherited from the source region.

On the other hand, Ague and Brimhall (1988) suggest that substantial contamination of granitoid magmas by country rocks has occurred in Californian batholiths. Where strongly contaminated by graphitic pelites, the I-type tonalites are reduced ilmenite-series, otherwise they are magnetite-series. Pecherskiy (1965) estimates that the nature of the country rocks has significant effects on the magnetic properties of granitoids in NE Russia in at most 20% of cases (probably much less).

As well as the country rocks affecting the magnetic properties of the granitoid, emplacement of granitoids frequently has a pronounced effect on the magnetic properties of country rocks that are metamorphosed and metasomatised by the intrusion. Often the magnetic signature of the contact aureole is more pronounced than that of the granitoid itself. Speer (1981) studied the mineralogical changes, including production of secondary magnetite, within the contact aureole of the Liberty Hill pluton, South Carolina. For that granitoid, there is a very

clear relationship between the detailed magnetic signature of the aureole and changes in metamorphic grade of the metapelitic country rocks, as shown in Figure 16. There are smooth variations in magnetite content, correlated with changes in mineral modes and mineral chemistry, within metamorphic zones, with inflections at metamorphic isograds. The susceptibility of the metamorphic magnetite zone is substantially greater than that of the magnetite-series pluton. Outside the aureole, the susceptibility of the country rocks is very low. The magnetic signature of the granite and aureole comprises a relative magnetic high (~ 200 nT above regional background) over the granite, rimmed by a strong, narrow high (~ 500 nT above background) centred on the magnetite-rich middle to outer aureole, dropping to the regional background level outside the aureole.

Contact aureoles around granitoids that intrude pyritic sediments may exhibit substantial magnetic anomalies due to breakdown of pyrite to monoclinic pyrrhotite (the ferromagnetic variety of pyrrhotite). Monoclinic pyrrhotite generally carries a relatively strong remanent magnetisation, characterised by $Q \gg 1$.

MAGNETIC PETROLOGY AND METALLOGENY OF INTRUSIONS

Mineralisation in layered mafic/ultramafic complexes

Differentiation of reduced mafic magmas within large, essentially closed system, slowly cooled magma chambers proceeds according the tholeiitic trend discussed above, with initial iron-enrichment and late iron-depletion, producing zoned complexes with basal ultramafic layers, overlain by paramagnetic mafic rocks, then by increasingly more magnetic mafic rocks, grading finally to MFM granophyres. Cr-mineralisation occurs as chromite bands towards the top of the ultramafic zone, which contains no primary magnetite, but may be MFM due to secondary magnetite produced by serpentinisation of olivine. The paramagnetic mafic rocks that overly the ultramafic zone may host platinum group element and Cu-Ni mineralisation. In the upper portions of the zoned complex, the SFM upper ferrogabbro and ferrodiorite zones host bands of titanium- and vanadium-bearing cumulus magnetite, which may constitute economic ore deposits of Ti and V. The Bushveld Complex, which hosts the world's greatest repository of magmatic ore deposits, may be regarded as the type example of such mineralised layered mafic/ultramafic intrusions.

An idealised model of the magnetic stratigraphy can be developed from this generalised picture of layered mafic/ultramafic complexes. Clark et al (1992a, 1992b) have used this model to predict magnetic anomalies over tilted Bushveld-type complexes.

Oxidation state and metallogenic associations

More than 30 years ago Pecherskiy (1965) noted an empirical association of granitoid-related gold deposits with ferromagnetic granitoids and tin deposits with paramagnetic granitoids. Ishihara (1981) established the important correlation between his magnetite- and ilmenite-series granitoid classification and granitoid-related mineralisation. For example, copper and molybdenum porphyries are almost always magnetite-series, whereas tin granites are invariably ilmenite-series. It has become apparent in recent times that this relationship is not just empirically based, but can be related to redox conditions in the magma. Ishihara's data on metallogenic associations with granitoid series and with susceptibility are reproduced in Figure 17.

The compatible or incompatible behaviour of multivalent metals such as Cu, Mo, W and Sn in the melt depends on their valency, which is a function of redox conditions. For
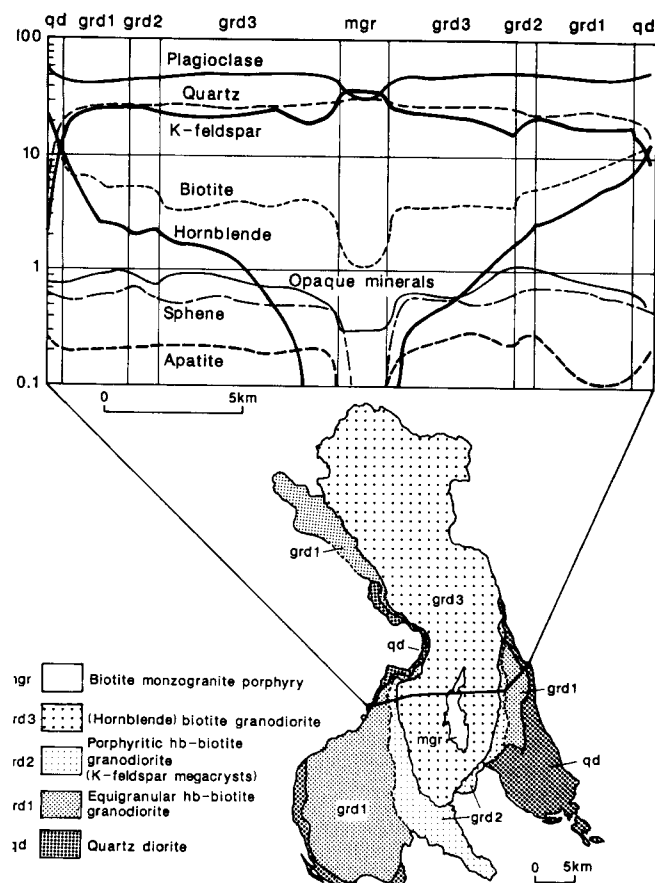
Figure 19. Compositional variations across the oxidised I-type Tuolumne intrusion, California, which shows a classic normal zonation from relatively mafic margin to felsic core, produced by fractional crystallisation (after Bateman and Chappell, 1979). Modal amounts of major, varietal and accessory minerals are plotted as a function of position along the profile indicated, which passes through all mapped phases of the pluton. Magnetite is the dominant opaque mineral.

example, tin occurs in two oxidation states in magmas: stannous (Sn^{2+}) and stannic (Sn^{4+}). The oxidised stannic species fits easily into the structures of minerals such as magnetite and sphene, which are diagnostic of oxidising conditions in the magma, and is therefore dispersed throughout an oxidised granitoid. On the other hand, the reduced stannous ion is too large to be accommodated readily within mineral structures and is accordingly concentrated in the residual melt. Thus, reduced, and therefore paramagnetic, granitoids are potential sources of tin mineralisation, whereas magnetite-bearing granitoids are too oxidised to be associated with tin deposits. Development of an exsolved fluid and partitioning of ore elements into hydrothermal liquids or vapour phases also depends strongly on the nature and concentrations of volatile species. The ratios SO_2/H_2S and CO_2/CH_4 depend on oxygen fugacity and therefore oxygen fugacity exerts a major influence on hydrothermal evolution, concentration of ore elements into mineralising fluids, and transport and deposition of ore elements.

Khitrunov (1985) has attempted a general explanation of the empirical relationships between oxidation state of granitoids and associated Cu, Mo, W or Sn mineralisation, concluding that magmatic conditions are progressively more reduced for Mo, Cu, W and Sn mineralisation. Cameron and Carrigan (1987) and Hattori (1987) have pointed out the association between oxidised felsic magmas, magnetic granitoids and Archaean gold deposits. They have

given a detailed discussion of the factors favouring incorporation of gold from sulphide minerals in source rocks, concentration into a CO_2-rich melt at mesothermal levels and deposition after development of an immiscible fluid phase. Sillitoe (1979) pointed out the association between gold-rich porphyry copper deposits and oxidised, magnetite-rich plutons with a magnetite-rich potassic alteration zone. Kwak and White (1982) distinguished between reduced porphyry tin and W-Sn-F skarn deposits and more oxidised W-Mo-Cu skarn deposits and Cu porphyries (Figure 18).

Blevin and Chappell (1992, 1995) have published thorough analyses of the metallogenic implications of granitoid chemistry, oxidation state and magmatic differentiation, based mainly on studies of the Lachlan Fold Belt. Sn mineralisation is associated with both S- and I-type granites that are reduced and have undergone fractional crystallisation. Such granites contain negligible magnetite and are paramagnetic. On the other hand, Cu and Au mineralisation is associated with oxidised, magnetite- and/or sphene-bearing, intermediate I-type suites. Mo is associated with similar granites that are more fractionated and oxidised. W does not appear to show a close relationship to granitoid type and is an opportunistic ore element, occurring in association with a number of other metals. Within mineralised granitoid suites, ore element ratios are simply related to relative oxidation state and degree of fractionation.

The importance of fractional crystallisation

Blevin and Chappell (1992, 1995) point out that fractional crystallisation of magmas is a powerful mechanism for concentration of ore elements into the residual melt, which is a prerequisite for formation of intrusive-related mineralisation. Late stage fractional crystallisation leads to quasi-exponential increases of concentration for incompatible elements in the residual melt, which may then partition the ore elements into late stage fluids and ultimately deposit them in a suitable trap to form economic mine-ralisation. Fractionated granitoids can be recognised, for example, by high Rb content and high Rb/Sr, which are sensitive indicators of fractional crystallisation. Other causes of chemical variation within granitoid suites, such as restite unmixing, magma mingling or crustal contamination cannot produce the enormous concentration factors required to form an ore deposit. Magmatic differentiation by fractional crystallisation is characteristic of melt-rich magmas. Thus mineralisation is associated with granitoids that are derived from hot magmas (very hot magmas if the source region is relatively anhydrous) or with felsic granitoids that have undergone extensive fractionation after all restite has separated from the melt.

The Tuolumne intrusion, California, represents a classic zoned pluton that grades from quartz diorite at the margin through progressively more felsic hornblende- and biotite-bearing granodiorite phases, to a core of biotite monzogranite porphyry (Bateman and Chappell, 1979). The normal zoning pattern, from mafic margin to felsic core, represents fractional crystallisation within the magma body as it cooled from the outside in. Figure 19 shows the compositional variations across the Tuolumne intrusion. Note the general slow decrease in opaque mineral (mainly magnetite) content from the margins towards the centre, with a pronounced dip in modal magnetite within the felsic core.

The expected magnetic signature of fractional crystallisation is gradation or zoning of susceptibility, where the most fractionated phase, which is most likely to be intimately associated with the mineralisation, has the lowest susceptibility. Figure 19 provides an example of this pattern.

In the case of an oxidised comagmatic suite, the susceptibility contrast between less evolved and more evolved phases should be large, whereas the effect will be subtle for a reduced paramagnetic suite. Large increases in radioelement concentrations and changes in radioelement ratios in the most fractionated rocks may also be detectable radiometrically. Airborne radiometric data can complement magnetic survey data in this environment, because the radiometric signal is best developed over the most felsic and fractionated intrusive phases, which are the phases that have the most subdued magnetic signature.

The Tuolumne pluton is unmineralised, probably because this intrusion, at least at the current level of exposure, was a sealed system during emplacement and cooling, precluding escape of late metal-bearing hydrothermal fluids, and because fractionation of the magma did not quite proceed to the stage required to concentrate metals into an ore-bearing fluid. However, a slightly more evolved variant of the Tuolumne intrusion, emplaced at a shallower depth or in a more favourable structural setting for tapping off hydrothermal fluids should be quite favourable for development of Cu-Au mineralisation.

Effects of sulphur saturation and halogen contents of magmas

Wyborn and Sun (1994) suggested that most magma types are generally sulphur-saturated and are unlikely to produce gold or copper-rich fluids after fractional crystallisation. Au and Cu partition strongly to sulphide phases and sulphur saturation leads to precipitation of sulphides and early removal of these metals from the melt. For development of a magmatic Cu-Au deposit, the magma must remain sulphur-undersaturated throughout most or all of its magmatic evolution. Oxygen fugacity has a large effect on sulphur saturation. Under oxidising conditions, sulphur becomes more soluble in the magma, dissolving as an anhydrite component. Thus oxidised magmas are more likely to be sulphur-undersaturated and are more likely to generate Cu-Au mineralisation.

The most favourable magma source for formation of high-gold sulphur-undersaturated magmas is lithospheric mantle that has already been depleted in sulphur by removal of sulphur-saturated basaltic melt, leaving behind small amounts of sulphide enriched in Cu, Au and other precious metals. If this refractory mantle is metasomatised, its liquidus temperature is lowered and it can subsequently undergo partial melting more readily, in appropriate tectonic conditions. The magmas generated often have shoshonitic affinities and have characteristics that are favourable for generation of magmatic-hydrothermal mineralisation. Less potassic magmas (i.e. those within the normal K-SiO_2 field for calc-alkaline magmas) that are derived from less metasomatised mantle can also give rise to large Cu-Au deposits, provided the mantle source is sulphur-undersaturated. Relatively low water and sulphur contents of these mantle-derived magmas produce rather inconspicuous alteration halos with restricted potassic zones and little iron sulphide. However, the oxidised nature of the magmas and high K^+, which boosts Fe^{3+} content, encourages formation of both magmatic and hydrothermal magnetite.

Halogens are important complexing agents for metals in hydrothermal fluids and Cl and F contents of magmas influence development of intrusive-related mineralisation. Cl decreases and F increases with fractional crystallisation in both I- and S-type granitoids (Blevin and Chappell, 1992). Cl contents of I-type granitoids are higher than for S-types. When a hydrothermal fluid exsolves from a silicate

melt Cl partitions strongly to the aqueous phase, accompanied by chloride-complexed metals such as Fe, Mn, Cu, Mo, Pb, Sn and S. However, if a sulphur-enriched magmatic vapour phase forms, sulphide-complexed Cu and Au preferentially partition into the low viscosity vapour, which can travel considerable distances before deposition, whereas Fe, Mn, Pb and Zn preferentially partition into the brine and tend to be deposited closer to the intrusion from which they emanate.

Effects of hydrothermal alteration

Studemeister (1983) pointed out that the redox state of iron in rocks is a useful indicator of hydrothermal alteration. Large volumes of fluid or high concentrations of exotic reactants, such as hydrogen or oxygen, are required to shift Fe^{3+}/Fe^{2+} ratios. When reactions associated with large water/rock ratios occur, the change in redox state of the rocks produces large changes in magnetic properties due to creation or destruction of ferromagnetic minerals.

Criss and Champion (1984) studied the southern Idaho batholith. They showed that intense hydrothermal alteration around Tertiary plutons generally reduced the susceptibility of magnetite-series Mesozoic tonalites and granodiorites over substantial areas. However, hydrothermal alteration with a lower water/rock ratio locally produced secondary magnetite within ilmenite-series granitoids, enhancing their susceptibility.

Hollister (1975) distinguished between the Lowell and Guilbert (1970) quartz monzonite model of porphyry copper deposits and a diorite model. The Lowell-Guilbert model incorporates a core potassic zone, surrounded successively by phyllic, argillic and propylitic zones arranged in concentric but incomplete shells. This model is typically most applicable to calc-alkaline granodiorite-quartz monzonite porphyries (often associated with quartz diorite intrusions) with copper and molybdenum mineralisation, but negligible gold. In the diorite model the phyllic and argillic zones are absent and the propylitic zone adjoins the core potassic zone. Sulphides are less developed in the diorite model and lower pyrite contents in the altered host rocks allow some of their magnetite to survive alteration. The diorite model is applicable usually to syenite-monzonite porphyries associated with diorites and often contain gold, as well as copper, but no economic molybdenum mineralisation.

Phyllic alteration, argillic alteration and intense propylitic alteration associated with porphyry intrusions tend to destroy magnetite within the intrusion and in surrounding rocks. Weak to moderate, but pervasive, propylitic alteration may leave most of the magnetite in host rocks relatively unaffected. On the other hand, the potassic alteration zone associated with oxidised, magnetic felsic intrusions is often magnetite-rich. This is commonly observed for Au-rich porphyry copper systems (Sillitoe, 1979). It is evident from the above descriptions of the Lowell-Guilbert and Hollister's diorite models that the magnetic signatures of the two types of system should differ substantially. Clark et al. (1992a; 1992b) presented a theoretical magnetic signature of an idealised gold-rich porphyry copper deposit, based on the Sillitoe (1979) model and magnetic petrological concepts. Early potassic (biotite-rich) alteration around the gold-mineralised Mount Leyshon Complex (Queensland), which is comprised of intrusive breccias and trachytic to rhyolitic porphyry plugs and dykes, produced abundant magnetite in metasedimentary and doleritic host rocks that

adjoin the southern half of the Mount Leyshon Complex. That alteration is therefore largely responsible for the Mount Leyshon magnetic anomaly (Sexton et al., 1995). However, the equivalent alteration within felsic, iron-poor, granitic host rocks, around the northern portion of the complex, produces K-feldspar alteration with little or no secondary magnetite. Thus, the Mount Leyshon magnetic anomaly is centred over the southern portion of the complex and its adjoining metasomatised aureole, rather than being symmetrically distributed around the complex.

In most porphyry systems, both primary magmatic magnetite and hydrothermal magnetite are generally in the multidomain size range. Multidomain magnetite boosts susceptibility but is not an efficient or stable carrier of remanent magnetisation. Thus, most of the magnetic signature of such porphyry systems is attributable to induced magnetisation, perhaps slightly enhanced by viscous remanence. However, alteration of certain country rocks can produce substantial quantities of fine-grained magnetite that is capable of carrying intense and stable remanence.

The magnetic anomaly at Mount Leyshon is a pronounced low that arises from reversed remanent magnetisation (Q >>1) of the biotite-magnetite altered metasediments and dolerites. Magnetite-bearing skarns with reversed remanence are also responsible for negative anomalies at the Red Dome Au deposit in NE Queensland (Collins, 1987). Monoclinic pyrrhotite may also carry intense remanence and produce large magnetic anomalies. Therefore, pyrrhotite skarns may produce strong magnetic anomalies that are dominated by remanent magnetisation.

Magnetite-rich alteration zones around calc-alkaline porphyry copper deposits have been extensively discussed by Clark and Arancibia (1996). These authors argue that magnetite-rich vein systems in and around some porphyry systems are often early (pre-mineralisation) and are distinct from magnetite-biotite potassic alteration that is associated with sulphides and Cu-Au mineralisation. The magnetite ± amphibole ± plagioclase alteration, with very little sulphide, represents the initial stage in the evolution of a subclan of porphyry copper deposits. Deposition of this assemblage is favoured by host rocks of mafic-intermediate composition. Host rocks influence deposition of magnetite around these systems, but iron metasomatism effected by magmatic conditions is also demonstrably important. The early strongly magnetic alteration appears to be associated with strongly oxidised intrusions that contain magnetite + sphene rather than the less oxidised assemblage magnetite + ilmenite.

Wall and Gow (1996) recognise a magnetite-rich Cu-Au class and a haematite-rich Cu-Au (U, REE) class of deposits associated with Proterozoic felsic plutons. Magnetite precipitation may be an important chemical control on sulphide precipitation in granitoid roof zones. The haematite association overprints the magnetite-rich bodies and results from highly oxidised lower temperature fluids with major meteoric component. La Candelaria-type magnetite mineralisation in Chile and Peru has some similarities and may be related to Mesozoic granitoids. The granitoids associated with these types of mineralisation are oxidised, high temperature, magnetite-series plutons.

Influence of country rocks

The nature of the country rock is crucial in the case of magmatic-hydrothermal skarn deposits, which develop in carbonate rocks that have been metamorphosed and metasomatised by the mineralising intrusion. In most cases

emplacement of the intrusion into non-carbonate rocks would not have resulted in economic mineralisation. The review of Einaudi et al. (1981) contains much useful information relevant to magnetic petrology of skarn deposits. Magnetite contents of magnesian skarns developed in dolomite are generally higher than those of calcic skarns developed in limestone, because Fe-rich calc-silicates are not stable in a high-Mg system. However both island arc-type calcic skarns (associated with gabbros and diorites in volcano-sedimentary sequences) and Cordilleran-type magnesian skarns (associated with quartz monzonites or granodiorites intruding dolomites) have been mined for magnetite. Such deposits are evidently associated with very large magnetic anomalies.

Cu skarns (mostly associated with epizonal quartz monzonite and granodiorite stocks in continental settings) are associated with oxidised assemblages, including magnetite + haematite, with the less common magnesian skarns exhibiting higher magnetite and lower sulphide contents than calcic skarns. Tungsten-bearing skarns (associated with mesozonal calc-alkaline quartz monzonite to granodiorite intrusions) have a more reduced calc-silicate and opaque mineralogy than Cu-skarns, but typically contain minor magnetite and/or pyrrhotite and would therefore be expected to exhibit a relatively weak, but nevertheless detectable, magnetic signature in most cases. Calcic Zn-Pb skarn deposits associated with granodioritic to granitic magmatism and Mo skarns associated with felsic granites appear to contain relatively little magnetite. Sn skarns are associated with reduced ilmenite-series granites and have relatively low sulphide contents. The skarns themselves contain magnetite ± pyrrhotite and exhibit a substantially larger susceptibility than the paramagnetic granite and unaltered host rocks. Massive sulphide replacement tin orebodies in dolomite (e.g. Renison and Cleveland deposits, Tasmania) are rich in monoclinic pyrrhotite and have high susceptibilities, with substantial remanent magnetisation. This type of orebody may represent the low temperature distal analogue of magnesian Sn skarns.

Webster (1984) analysed magnetic patterns over a number of granitoids associated with tin mineralisation in the Lachlan Fold Belt and contrasted these with unmineralised and Cu-Mo-W mineralised granitoids. The characteristic magnetic signature of granitoid-associated tin mineralisation is: a granitoid with low magnetic relief, surrounded by a more magnetic aureole, with significant magnetic anomalies associated with the mineralisation.

Wyborn and Heinrich (1993) and Wyborn and Stuart-Smith (1993) have suggested that particular host rocks favour deposition of Au mineralisation from oxidised fluids that emanate from felsic granitoids and move up to 5 km from the granitoid contact. Graphite-, sulphide- and magnetite-bearing lithologies are capable of reducing the fluids and depositing Au and Cu, whereas Pb and Zn are preferentially deposited in carbonate rocks. Au-only mineralisation will preferentially be deposited within graphite-bearing but magnetite- and sulphide-poor rocks, whereas magnetite and or iron sulphide-rich rocks tend to precipitate Cu and Au together. These relationships appear to have been observed in the eastern Mount Isa Inlier and the Pine Creek Inlier. Thus a rock unit that is strongly magnetic, indicative of high magnetite content, may be a favourable site for deposition of Au-Cu mineralisation sourced from a nearby granitoid.

Table 1. Petrological and geological characteristics of ferromagnetic and paramagnetic igneous intrusions

FERROMAGNETIC INTRUSIONS	PARAMAGNETIC INTRUSIONS
Source Rock	
Mantle Mafic crustal underplate Oxidised intermediate-felsic igneous rocks	Metasediments (particularly pelites) Reduced igneous rocks
Lithology	
Gabbro>diorite>tonalite>granodiorite>granite Hornblende ± biotite granitoids and biotite granites with high Mg, low Al biotite Pyroxene ± hornblende granitoids Many monzonites, quartz monzonites, syenites, quartz syenites and miaskitic nepheline syenites Most alkali gabbros, essexites, ijolites etc. Ferrogabbros, ferrodiorites and granophyres, within upper levels of layered mafic complexes	Predominantly granite and granodiorite Muscovite and two mica granitoids, most leucogranites, biotite-rich granitoids Cordierite, corundum or aluminosilicate-bearing granitoids Peralkaline granites, syenites or nepheline syenites Lower gabbros in layered mafic complexes
Mineralogy	
Generally higher colour index Biotite ± hornblende in felsic calc-alkaline granitoids; hornblende ± pyroxene ± olivine (except fayalite) in more mafic varieties Fe^{3+} and Mg-rich biotite and hornblende; biotite colour is brown, black or olive green	Generally lower colour index Biotite + muscovite, cordierite, garnet or aluminosilicate in calc-alkaline granitoids.
Sphene (> 0.1 vol %) ± hemoilmenite (8 mole % to 20 mole % Fe_2O_3) or Mn-rich ilmenite (up to 30 mole % $MnTiO_3$) ± epidote ± allanite ± pyrite as accessories White plagioclase + pink K-feldspar	Fe^{2+} and Al-rich (annite/siderophyllite-rich) biotite, often with "foxy red" colour; occasionally Fe^{2+}-rich hornblende or fayalite "Reduced" ilmenite (< 8 mole % Fe_2O_3, usually Mn-poor) ± pyrrhotite (predominantly hex po) ± spinel ± graphite as accessories, primary sphene absent White plagioclase + white K-feldspar or sometimes brick red K-feldspar; sometimes green plagioclase + pink K-feldspar
Intermediate to Fe-rich olivine and pyroxenes + intermediate to sodic plagioclase ± hornblende ± apatite in upper ± middle levels of tholeiitic layered intrusions Zoned plagioclase (> 60 %) + quartz + biotite and/or hornblende in M-type oceanic/ophiolitic plagiogranites	Mg-rich olivine and pyroxenes, calcic plagioclase ± chromite in lower to middle zones of tholeiitic layered intrusions
Nepheline + alkali feldspar + plagioclase + calcic pyroxene + hastingsite + biotite in silica-undersaturated metaluminous (miaskitic) rocks Nepheline + alkali feldspar + sodic pyroxene and amphibole ± biotite without aenigmatite or astrophyllite in mildly peralkaline undersaturated rocks	Quartz + alkali feldspar + sodic pyroxene and/or sodic amphibole ± aenigmatite ± astrophyllite ± biotite in oversaturated peralkaline (ekeritic) rocks Nepheline + sodic pyroxene + aenigmatite ± astrophyllite in silica-undersaturated peralkaline (agpaitic) rocks
Chemistry	
Predominantly metaluminous, but also weakly peraluminous or weakly peralkaline granitoids (0.9 ≤ A/NK < A/CNK ≤ 1.1)	Strongly peraluminous (A/CNK > 1.1) and strongly peralkaline (A/NK < 0.9) granitoids, some metaluminous granitoids
Moderate-high ferric iron (Fe_2O_3 > 0.8 wt %, typically 1-3 wt%) and moderate-high total iron (> 2 wt% FeO^T) Moderate oxidation ratio (mean Fe^{3+}/Fe^{2+} ~ 0.6 at 60% SiO_2, Fe^{3+}/Fe^{2+} ~ 0.9 at 75% SiO2, i.e. molar Fe_2O_3/(FeO+ Fe_2O_3) ~0.2-0.3) Normative (diopside ± olivine ± acmite) plus > 1 wt % normative magnetite ± haematite Relatively anhydrous	Low ferric iron (Fe_2O_3 < 0.8 wt %) or very low total iron (< 1 wt % FeO^T) Low oxidation ratio (mean Fe^{3+}/Fe^{2+} ~ 0.1 at 60% SiO_2, Fe^{3+}/Fe^{2+} ~ 0.4 at 75% SiO_2, molar Fe_2O_3 /(FeO+ Fe_2O_3) = 0.05.-0.2) or very high oxidation ratio Normative corundum or normative acmite+sodium metasilicate Relatively hydrous
Emplacement depth	
Predominantly epizonal, particularly subvolcanic, some mesozonal and catazonal	Predominantly mesozonal or catazonal, some epizonal
Associated rocks	
Associated volcanics common Gabbro-diorite-trondhjemite associations (Gabbro)-diorite-granodiorite-monzogranite associations Diorite-monzonite-quartz monzonite-monzogranite associations Syenite-alkali syenite-alkali granite associations	Associated volcanics uncommon Syenogranite-monzongranite-granodiorite associations Quartz syenite-syenogranite associations
Tectonic setting	
Andinotype (subduction of oceanic plate beneath continental margin, generating Cordilleran I-type batholiths) Island arc plagiogranites, gabbros and quartz diorites Alpinotype (tectonically emplaced serpentinised peridotites, gabbros and plagiogranites) Caledonian-type post-closure uplift and tensional regimes with major faulting Anorogenic, rifting-associated moderately evolved granitoids	Hercynotype (continental collision, e.g. Himalayan and Hercynian leucogranites) Encratonic ductile shear belts with thickened continental crust Late tectonic/post tectonic catazonal migmatites and mesozonal granitoids associated with regional metamorphism Compressional regimes Anorogenic, rifting-associated highly evolved granitoids

A/NK = atomic Al/(Na + K); A/CNK = atomic Al/(Ca + Na + K)

Table 2. Magnetic petrophysical classes of intrusive rocks and alteration zones associated with mineralisation

TYPE OF MINERALISATION	MAGNETIC PETROPHYSICAL CLASSIFICATION
Au-rich (> 0.4 g/t) porphyry Cu within MFM mafic-intermediate igneous host rocks	SFM granitoid (M- or I-type) + SFM potassic alteration zone ± PM phyllic zone + WFM to MFM propylitic zone
Porphyry Cu within MFM mafic-intermediate igneous host rocks	MFM to SFM granitoid ± SFM potassic alteration zone + PM phyllic zone + PM argillic zone + WFM to MFM propylitic zone
Porphyry Mo	WFM granitoid directly associated with mineralisation, zoned to MFM less fractionated granitoid phase WFM to MFM granitoid
Au-scheelite-quartz exogranitic plutonic vein (scheelite contains Mo) W-Mo-Cu skarn W-Cu-Sn veins. Tungsten mineral is wolframite or Mo-free scheelite W-Sn-F skarn Sn-W greisen	MFM to SFM granitoid + SFM skarn PM granitoid
	PM granitoid + WFM to SFM skarn PM granitoid
Cr, PGEs, Ni-Cu in lower levels of layered mafic complex	PM to WFM gabbros overlying PM unserpentinised ultramafics or, more commonly, MFM to SFM serpentinised ultramafics
Ti, V in upper levels of layered mafic complex Sn-W, Be, Li and U associated with peraluminous two-mica granites	SFM gabbros PM granitoid
Nb-Ta, REE mineralisation associated with peralkaline anorogenic ring complexes	PM granitoid ± MFM to SFM carbonatite

PM = paramagnetic; WFM = weakly ferromagnetic; MFM = moderately ferromagnetic; SFM = strongly ferromagnetic; VSFM = very strongly ferromagnetic (see text)

CONCLUSIONS

Relationships between magnetic properties of igneous intrusions and their mineralogy, chemical composition, geological setting and history are complex. However, clear patterns can be discerned and much progress has been made in recent years in understanding the geological factors that control their magnetic properties. Although magnetic properties are not predictable with any reliability from first-order rock names, a more detailed classification of intrusive igneous rocks does correlate well with magnetic properties, because there are many links between magnetic properties of igneous intrusions and their geological, chemical and mineralogical characteristics. These patterns arise directly in some cases, e.g. the correlation between oxidation ratio and magnetite content, but in many cases they are indirect.

One example of an indirect relationship is the strong association between paramagnetic ilmenite-series granitoids and S-type granitoids. The magnetite-poor nature of most S-type granitoids arises from their reduced character, which reflects incorporation of crustal carbon. Carbon content is an incidental, rather than a defining, characteristic of an S-type granitoid.

Data on geological, geochemical and mineralogical associations with magnetite in granitoids are summarised in Table 1.

Of particular importance to exploration is the clear, albeit indirect, relationships between a number of important types of intrusive-related mineralisation and the magnetic properties of the associated intrusions and their alteration systems (Table 2). There is now sufficient knowledge to develop and test improved magnetic exploration models for intrusive-related mineralisation. Key elements that need to be incorporated into magnetic exploration models include:

1. Regional-scale links, summarised in Table 2, between intrusive-related mineralisation of a given type and the magnetic petrology of the associated intrusions.
2. The magnetic expression, at regional to deposit scale, of structural controls on location of intrusions or fluid pathways, and of lithological controls on deposition of mineralisation.

3. The magnetic expression of fractional crystallisation (zoned magnetic properties or complex patterns that suggest multiple comagmatic intrusions of varying fractionation) at a district to deposit scale.
4. Use of magnetic petrological principles and magnetic petrophysical data to predict the magnetic properties of intrusive phases, host rocks and alteration zones for each deposit type.
5. Incorporating predicted magnetic properties into conventional ore deposit models to enable calculation of theoretical magnetic signatures of deposits for a range of geological settings.

Clark et al. (1992b) showed some examples of simple magnetic exploration models. Although such models are inevitably simplistic, if they are based on petrophysical data and magnetic petrological principles they should improve the utility of magnetic surveys in exploration.

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