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Time–temperature relations for the remagnetization of pyrrhotite (Fe_7S_8) and their use in estimating paleotemperatures

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Abstract

Paleotemperature controls the maturation of coal and hydrocarbons in sedimentary basins and is also important in determining paleogeothermal gradient and hence tectonic style in exhumed metamorphic terrains. One method of estimating paleotemperature analyses the partial remagnetization of a rock due to heating in thick volcanic or sedimentary sequences, over subcrustal heat sources such as plumes, or at convergent plate margins. The overprinted natural remanent magnetization (NRM) of a rock records both the age and the paleotemperature of remagnetization, but a temperature correction from laboratory to geological time scales is required, using theoretical time–temperature relations. Time–temperature relations are well known for magnetite (Fe_3O_4) but are reported here for the first time for pyrrhotite (Fe_7S_8), another common NRM carrier. Data for each mineral separately yield independent estimates of paleotemperature if geologically reasonable estimates of heating time can be made. Paleotemperature can be estimated without geological input if data for both minerals are combined. Together with the age of remagnetization, determined from the paleomagnetic pole of the NRM overprint, these paleotemperature estimates can be used to infer the history of heating and uplift following burial. As a test case, we examine thermally acquired NRM overprints carried by pyrrhotite (Fe_7S_8) and magnetite (Fe_3O_4) in the Milton Monzonite of southeastern Australia. These overprints record a heating event about 100 Ma ago, probably thermal doming prior to rifting of the Tasman Sea, that upgraded coal rank in the Sydney Basin. Extrapolating from laboratory to geological times, using the new time–temperature contours for pyrrhotite, we estimate that the presently exposed Sydney Basin in the vicinity of the Milton Monzonite was remagnetized by heating to 165 ± 30°C for ≈ 100 ka. Assuming a paleogeothermal gradient of 70°C/km appropriate for young or incipient rifts, the depth of burial at the time of remagnetization is estimated to have been 2.3 ± 0.4 km. This figure is in excellent agreement with independent estimates based on reflectance data for the coal accessory mineral vitrinite.

Keywords: paleotemperature; remagnetization; pyrrhotite; magnetite; P-T-t paths

1. Introduction

Knowledge of paleotemperatures in sedimentary basins is important because thermal history

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controls the maturation of coal and hydrocarbons [1]. Paleotemperatures are also needed to infer paleogeothermal gradient and tectonic history of uplifted terrains, such as those originally formed at convergent plate margins (low thermal gradient) or in rifted margin environments (high thermal gradient). Many geothermometers are in common use, among them coal rank and reflectance of vitrinite, a coal accessory mineral [1], clay mineral diagenesis [2], conodont and acritarch color alteration [3], metamorphic mineral assemblages [4], and reset $^{40}Ar/^{39}Ar$ and fission track ages [5,6]. Most of these apply to specific rock types over limited temperature ranges.

Another relatively unrestricted geothermometer uses remagnetization temperatures of natural remanent magnetization or NRM [1,7–15]. Thermal overprinting of NRM can occur in any rock type over broad temperature ranges: 20–580°C for magnetite (Fe_3O_4) or 20–320°C for monoclinic pyrrhotite (Fe_7S_8). Most continental rocks contain one or other of these minerals. A further advantage of the thermal remagnetization method is that the time of heating can also be found, by comparing the paleomagnetic pole of the NRM overprint to a known apparent polar wander path.

2. The thermal remagnetization geothermometer

The remagnetization geothermometer [7] is based on the Néel [16] theory of magnetic single-domain grains. At absolute temperature T, the relaxation time τ for thermally activated magnetization changes in weak fields is:

$$1/\tau = C \exp[-\mu_0 V H_{K0} M_{s0} \beta^n(T)/2kT] \qquad (1)$$

where $C = 10^9$–10^{10} s^{-1}, $\mu_0 = 4\pi \times 10^{-7}$ H/m, V and H_K are grain volume and coercive force, $\beta(T)$ is the normalized T dependence of spontaneous magnetization M_s, n depends on the mechanism of anisotropy, and k is Boltzmann's constant. H_{K0} and M_{s0} are the values of H_K and M_s at room temperature T_0. Because τ depends exponentially on T and $\beta(T)$, grains with a particular V and H_{K0} will acquire a permanent NRM when they are cooled a few °C through their blocking temperature T_B, at which $\tau \approx t$, a typical experimental time. The NRM of the same grains will be thermally demagnetized when reheated through the unblocking temperature T_{UB}, at which τ is again $\approx t$.

In nature, a thermal NRM overprint is acquired at a remagnetization temperature $T_B = T_r$ over a time interval t_r of geological length, whereas in laboratory heatings for short time intervals t_L, thermal demagnetization occurs at a higher temperature $T_{UB} = T_L$. The last grains to be magnetized in nature and demagnetized in the laboratory are those with the largest value of VH_{K0}. For these grains, Eq. 1 gives the remagnetization relation [7]:

$$T_r \beta_r^{-n} \ln(Ct_r) = T_L \beta_L^{-n} \ln(Ct_L) = \mu_0 V H_{K0} M_{s0}/2k \qquad (2)$$

in which subscripts r, L and 0 refer to values at T_r, T_L and T_0, respectively. (For an equivalent relation in continuous heating/cooling, see [8].)

Fig. 1 is a contour plot of $T\beta^{-n}(T)\ln(Ct)$ for various values of the constant $\mu_0 V H_{K0} M_{s0}/2k$. Possible combinations (T_r, t_r) of geological remagnetization temperature and time can be found by following the contour passing through the point (T_L, t_L) determined from laboratory thermal demagnetization of a thermal overprint. A unique solution (T_r, t_r) is given by intersecting contours for two minerals with different $M_s(T)$.

Contours for single-domain magnetite and hematite (αFe_2O_3) are known [7], but the T–t contours for pyrrhotite are reported here for the first time (see also [17,18]). For Fe_7S_8, we used a molecular field approximation $\beta(T) = 1.48(1 - T/T_C)^{0.52}$ with a measured Curie temperature $T_C = 593$ K (320°C) and $n = 2.6$ [19,20]. This β function and n value give good fits to M_s and coercive force data for single-domain and small pseudo-single-domain Fe_7S_8 over most of the range (T_0, T_C). We also recalculated contours for magnetite, using $\beta(T) = 1.193(1 - T/T_C)^{0.42}$ with $T_C = 853$ K (580°C), which gives a reasonable fit to $M_s(T)$ measurements [21], and $n = 2$, for single-domain grains with shape anisotropy.

The T–t contours for both minerals are steep

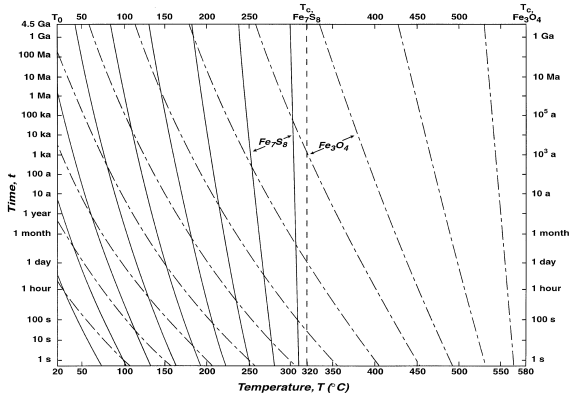

Fig. 1. Time vs. temperature contours calculated for pyrrhotite (Fe_7S_8) and magnetite (Fe_3O_4). Blocking or unblocking temperatures of magnetization for a particular ensemble of single-domain grains can be determined as a function of time by following one of the contours. Below the pyrrhotite Curie point, $T_C = 320°C$, magnetite and pyrrhotite contours defined by laboratory thermal demagnetization data intersect to give a unique determination of the remagnetization time and temperature in nature.

near the Curie temperature T_C, where M_s changes rapidly, and more gently sloping at lower temperatures. Because the Curie points of pyrrhotite and magnetite differ by more than 250°C, sloping magnetite contours and steeper pyrrhotite contours intersect at reasonably large angles in the range 20–300°C (Fig. 1). We can therefore determine remagnetization temperatures up to 300°C fairly accurately if we have thermal demagnetization data for thermal overprints from both these minerals, or data from pyrrhotite alone plus an estimate of geological heating time.

3. An example of the pyrrhotite remagnetization method: the Milton Monzonite

To illustrate both approaches, we will use paleomagnetic data from the Milton Monzonite [22,23], an early Triassic intrusion into flat-lying sediments of the Sydney Basin (Fig. 2). This example provides an ideal test of our calculations of paleotemperature for several reasons. First, the paleomagnetic results indicate thermal overprinting of both pyrrhotite and magnetite, as described below. Second, the Sydney Basin has a simple history: subsidence from Permian to early Jurassic, with deposition of several km of marine sediments and coal measures, followed by quiescence until thermal doming and initial rifting of the Tasman Sea caused rapid uplift and erosion 100–70 Ma ago. Finally, vitrinite reflectance measurements on coal deposits [1] provide independent estimates of paleotemperature and burial depth to which we can compare our results.

Thermal demagnetization was carried out using three non-inductive resistance furnaces in a field-free space, each furnace containing ≈20 cylindrical cores 2.5 cm in diameter. While one furnace was heating, another was cooling, and the specimens from the third were being measured on a SQUID magnetometer. Heatings were programmed using a controller that gave negligible

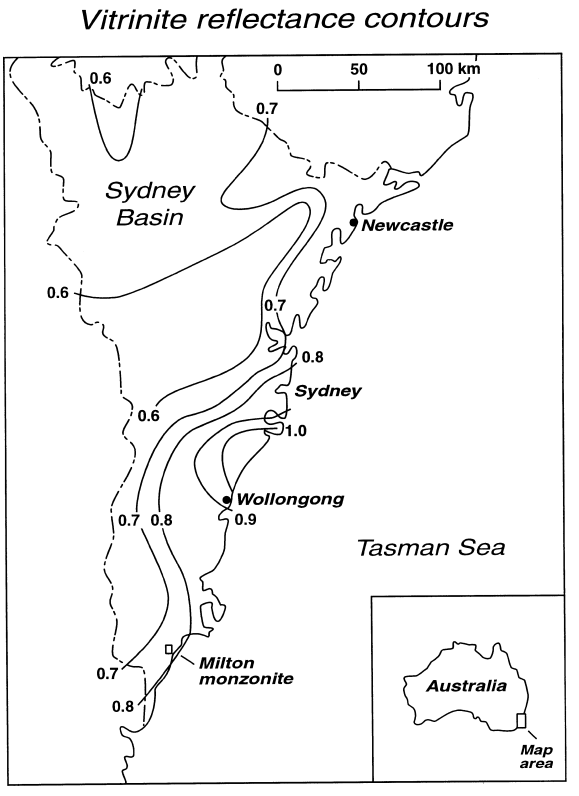

Fig. 2. Map of the Sydney Basin, Australia, showing contours of vitrinite reflectance of surface rocks in % (after [1]) and the location of the Milton Monzonite. Vitrinite reflectance and coal rank increase towards the Tasman Sea coast, indicating deeper levels of burial or a higher paleogeothermal gradient prior to rifting.

temperature overshoot. The time required for specimens to reach a uniform temperature throughout their volumes was determined to be 5–10 min by comparing surface and interior temperatures read by two thermocouples, one cemented to the surface of a dummy sample and the other embedded in its center. This equilibration time was used for the hold time at temperature T and is used later as the value of t_L. The value of T used in calculations is the average of the surface and interior thermocouple readings. Within each furnace, temperatures were quite uniform. Nevertheless, specimens were returned to the same positions for all heatings.

Magnetite is the usual NRM carrier in the Milton Monzonite (Fig. 3, left). However, at two sites, pyrrhotite carries both the primary NRM and a thermal overprint (Fig. 3, right; see also [22] Fig. 5). The primary NRM of sample 7g2 (labelled HT for higher temperature in Fig. 3) has T_{UB} values up to $\approx$313°C, just below 320°C, the Curie point of pyrrhotite. The pyrrhotite overprint (labelled LT for lower temperature) has lower T_{UB}'s that do not overlap those of the primary NRM, with the result that the HT and LT vector projections have a sharp intersection with no rounding indicative of overlap of T_{UB}'s. (There is also an overprint evident in Fig. 3 carried by magnetite, with a direction similar to LT but higher T_{UB} values than either LT or HT. It is

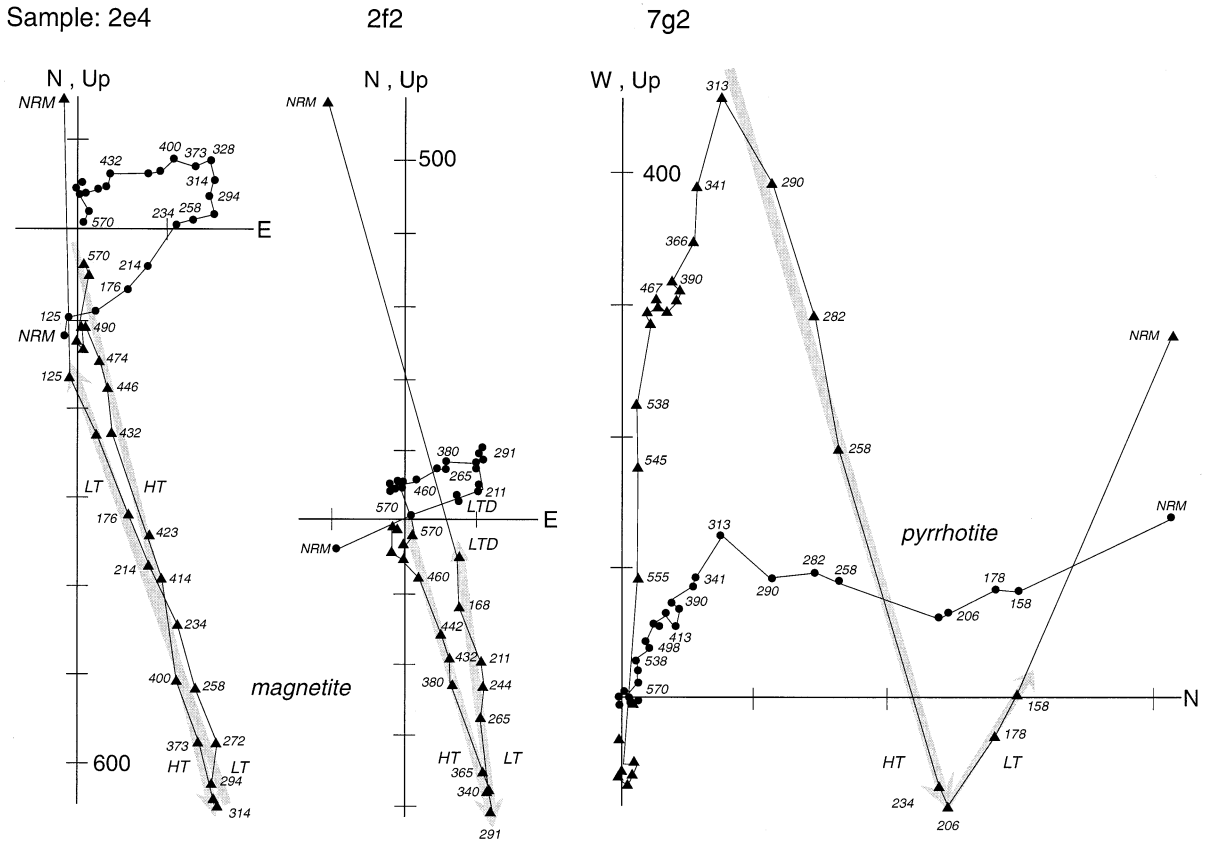

Fig. 3. Examples of vector plots of thermal demagnetization data (temperatures in °C) for samples of the Milton Monzonite. Vertical and horizontal plane projections of the NRM vector are denoted by triangles and circles, respectively. Samples 2e4 and 2f2 have both the primary HT and thermally overprinted LT vectors carried by magnetite. Prior LTD of 2f2 reduces its LT–HT junction temperature to 291°C, much below that of companion specimens. In sample 7g2, LT and HT are carried by pyrrhotite. A magnetite chemical overprint with approximately the same direction as LT is evident above 313°C but is irrelevant to the discussion in this paper.

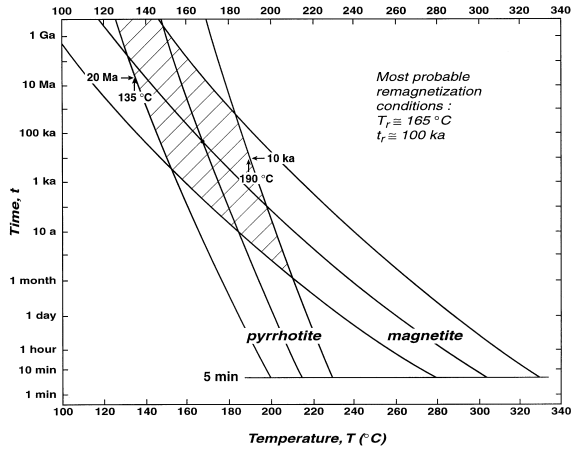

Fig. 4. Determining the remagnetization conditions (T_r, t_r) under which the Milton LT overprint was acquired in nature by extrapolating from laboratory (T_L, t_L) data for pyrrhotite- and magnetite-bearing samples, using the (T, t) contours of Fig. 1. When pyrrhotite carries the overprint, the observed range of T_L values in thermal demagnetization ($t_L = 5$ min) is 200–230°C. On geological timescales, this translates into upper and lower bounds on T_r of 135–190°C. The most probable remagnetization conditions are given by the intersection of the average pyrrhotite and magnetite contours: $T_r \approx 165$°C and $t_r \approx 100$ ka. The shaded region between the maximum and minimum pyrrhotite and magnetite contours contains all theoretically permitted (T_r, t_r) combinations, but lengths of geomagnetic polarity epochs and realistic uplift rates impose upper and lower limits on t_r. Our paleomagnetic estimates of T_r are in good agreement with independent estimates from vitrinite reflectance [1].

not relevant to our discussion because it is of chemical rather than thermal origin [22].) The non-overlapping T_{UB} ranges of the overprinting (LT) and overprinted (HT) components are diagnostic of thermal overprinting in single-domain grains [21,24,25].

The HT vector direction defines a paleomagnetic pole older than 200 Ma, whereas the overprint paleopole is about 100 Ma in age [22]. The overprint is ubiquitous throughout the Sydney Basin and was probably acquired during uplift associated with opening of the Tasman Sea. Judging by the junction temperature between the LT and HT vectors (Fig. 3, right), grains with laboratory T_{UB} values up to 206°C have been overprinted. Thus T_L for this sample is 206°C. By way of comparison, the junction temperature for sample 7d2 ([22] fig. 5) was 222°C.

Values of maximum unblocking temperature T_L for all the pyrrhotite-bearing samples are in the range 200–230°C ([23] fig. 2, 17 determinations). For the equilibration time at T_L during thermal demagnetization, we used the furnace hold time: $t_L \approx 5$ min (see above). The pyrrhotite contours through the extremes (5 min, 200°C) and (5 min, 230°C) outline the region of permitted (t, T) combinations in remagnetization (Fig. 4).

Geomagnetic and geological constraints narrow the range of possible t_r values, and therefore also T_r values. To be measurable as a single NRM vector, any overprint must be acquired within a single geomagnetic polarity epoch. The Milton overprint was acquired during a particularly long epoch, the Cretaceous normal superchron (119–84 Ma). The maximum allowable t_r is about half the length of the superchron or 20 Ma, giving a lower limit of 135°C for the maximum paleotemperature T_r during remagnetization (Fig. 4).

A minimum value for t_r is dictated by rates of uplift from depth. The high coal ranks observed in the Sydney Basin imply prolonged burial to 1–3 km depths [1]. Erosional unroofing and isostatic uplift from depth require a time t_r of at least 10 ka, which from Fig. 4 implies an upper limit of 190°C for T_r.

The range of possible paleotemperatures could be narrowed further if the constraints were tighter. For example, overprinting during a polarity chron of typical length would reduce the maximum t_r value to ≈ 1 Ma. For the same (t_L, T_L) values used in Fig. 4, this would raise the lower limit on T_r to 140°C. Likewise, one could argue that maximum T_r is experienced during prolonged residence at maximum depth, which for coalification requires t_r to be much longer than 10 ka. This would lower the upper limit on T_r.

4. Combining remagnetization data for pyrrhotite and magnetite

In order to determine the most probable combination of T_r and t_r without imposing any external constraints, we next combine remagnetization data for both pyrrhotite and magnetite. We use only T_L values for single-domain grains because

Néel's theory [16] applies only to such grains. This requirement is easier to satisfy for pyrrhotite, whose critical grain size d_0 for single-domain behavior is 1–2 μm [26], than for magnetite, with $d_0 = 0.05$–0.1 μm [21].

In the Milton Monzonite, there are four main groupings of T_L values ([23] fig. 2): 200–230°C; 280–330°C; 400–440°C; and 510–550°C. Group 1 samples contain pyrrhotite, as described earlier. Group 2 samples contain single-domain magnetite (an example is sample 2e4 in Fig. 3). Group 3 and 4 samples contain multidomain grains or mixed single-domain and multidomain grains of magnetite [23].

T_L values for multidomain grains are anomalously high [23] compared to the predictions of Eq. 2. However, T_L values for groups 3 and 4 are reduced to the same range as group 2 by prior low temperature demagnetization (LTD), consisting of zero-field cooling through the magnetite isotropic point at 130 K. For example, sample 2f2, treated with prior LTD, has $T_L = 291$°C (Fig. 3), whereas its companion specimens 2f1 and 2f5 which were not treated with LTD have T_L values above 350°C. LTD has the effect of unpinning domain walls, thereby demagnetizing multidomain NRM, while leaving single-domain NRM pinned by shape anisotropy largely intact [27–30]. For paleotemperature determination, we used only T_L values from groups 1 and 2, plus data for LTD-treated samples.

In Fig. 4, the minimum, average, and maximum T_L values for pyrrhotite- and magnetite-bearing samples, (200, 215, 230°C) and (280, 305, 330°C), respectively, are anchored by the horizontal line $t_L \approx 5$ min. The most probable combination (T_r, t_r) is the intersection point of the average pyrrhotite and magnetite contours: $T_r \approx 165$°C and $t_r \approx 100$ ka. Incorporating our previous upper and lower bounds for T_r, we estimate that the Milton region was heated to a peak temperature of 165 ± 30°C during remagnetization.

If a value of 10 min was used instead for t_L, the upper and lower bounds on T_r would increase by ≈ 2°C (determined by the shift in pyrrhotite contours) but the most probable T_r would actually decrease by ≈ 3°C (because the intersecting magnetite contour shifts upward more than the average pyrrhotite contour). Thus the T_r estimates are insensitive to small uncertainties in t_L.

We avoided long hold times in order to minimize chemical alteration. However, had we used larger values of t_L, lower T_L values would have been measured. These (t_L, T_L) data, in combination with the shifted contours, would lead to T_r estimates similar to those reported here. In other words, the method can be expected to work equally well for long or short hold times.

Notice that the upper and lower bounds on T_r are determined by the pyrrhotite contours alone. The magnetite contours are more oblique and would set broader limits on T_r. For this reason, pyrrhotite data and the newly calculated pyrrhotite T–t contours are particularly valuable in setting bounds on ranges of plausible paleotemperatures.

5. Comparison with earlier studies

In an earlier remagnetization study, Middleton and Schmidt [1] compared magnetite T_L data for the Milton Monzonite with hematite T_L data for the Patonga Claystone (near the coast south of Newcastle in Fig. 2). Both units fall between the 0.7% and 0.8% vitrinite reflectance contours, but they are geographically widely separated. The estimated T_r was 250–300°C for a duration of heating t_r of several million years [1]. Shorter durations of heating would drive the paleotemperature estimate still higher. The contours used a different remagnetization theory [31]; however, using the Néel [7,16] contours gave even higher T_r estimates. One problem could be combining data from widely separated parts of the Sydney Basin but the most probable reason is the use of uncorrected magnetite T_L values, most of which are anomalously high because of the presence of multidomain grains [23].

Middleton and Schmidt [1] independently estimated maximum paleotemperatures of surface rocks in the Sydney Basin using vitrinite reflectance data. Assuming residence times for coalification of 70–110 Ma, their estimates were $\approx$100–200°C using a paleogeothermal gradient of 70°C/km typical of heat flow in incipient or young rifts,

such as the Rio Grande Rift and the Red Sea [32,33]. Middleton and Schmidt recognized that these temperatures were significantly less than those estimated from the magnetic overprinting, but argued that 'coalification depends on long duration of heating, whereas overprint magnetizations reflect elevated temperatures', possibly of briefer duration.

Our work negates this view. The T_r estimate of 165 ± 30°C, using the Néel–Pullaiah [7,16] remagnetization contours for pyrrhotite and magnetite, is very comparable to values from vitrinite studies. Coalification and magnetic overprinting seem to respond in similar ways to long-term burial, although the relationship between vitrinite reflectance, time and temperature (Eq. 3 of [1]) is more complicated than the first-order Arrhenius law describing thermoviscous remagnetization (Eq. 1 of this paper). The earlier magnetic overestimates of T_r likely were biased by multidomain grains. Our approach thus reconciles previously divergent estimates of paleotemperatures during deep burial based on coal maturity and on remagnetization.

6. Advantages and limitations of the remagnetization geothermometer

Our methodology is new in two respects. First, it uses remagnetization data from a single formation, eliminating uncertainties about possible differences in paleotemperatures in different parts of a sedimentary basin. Second, it makes use of T_L data and remagnetization contours for pyrrhotite which are much less sensitive to the choice of remagnetization time t_r than are the corresponding data and contours for magnetite.

In fact, the pyrrhotite contours used by themselves set the lower and upper bounds on T_r (see Fig. 4). Pyrrhotite is of special importance because remagnetization temperatures in the 100–300°C (i.e. 373–573 K) range are close to the pyrrhotite Curie point of 320°C (593 K), so that M_s and β change rapidly with T, producing steep T–τ contours. Magnetite in the same temperature range is far from its Curie point, so that M_s and β change slowly with T, producing gently sloping T–τ contours.

The remagnetization method of estimating paleotemperature has some advantages over conventional estimates based on organic or clay diagenesis, color changes in microfossils, and petrographic observations of hydrothermal or metamorphic minerals. It requires only basic paleomagnetic instrumentation. Any igneous, sedimentary or metamorphic rock with a pre-existing NRM (not necessarily primary) and a thermal overprint can be used. Pyrrhotite is the preferred NRM carrier because its T–t contours are steep, but magnetite is usable if t_r can be estimated within one or two orders of magnitude. Finally, the vector direction of the overprint yields a paleomagnetic pole that dates the time of remagnetization.

There are two principal difficulties in using the remagnetization method. The first is locating suitably overprinted formations, particularly ones in which the thermoviscous overprint is carried by pyrrhotite. Although these are generally considered to be rare, there are a number of examples [18,34,35] from mountain belts, and pyrrhotite is increasingly being recognized as an important magnetic mineral in other settings as well. Secondly, one must ensure that the thermal overprint is carried by single-domain grains. This requirement is not too difficult to satisfy for pyrrhotite, high Ti titanomagnetite, or hematite, all of which have critical single-domain sizes $d_0 > 1$ μm, but it is a problem with magnetite, for which d_0 is 0.05–0.1 μm. Reliable T–t contours cannot be found for multidomain magnetite because a particular T_B has a spectrum of unblocking temperatures rather than a single T_{UB} as in single-domain grains [36,37]. However, LTD prior to thermal demagnetization is effective in reducing T_L to single-domain levels and is quick to perform using liquid N_2 on batches of samples.

7. Paleogeothermal gradient and depth of burial estimates

In analyzing sedimentary basins or metamorphic terrains, one would like to know not only paleotemperature but also paleogeothermal gradient and depth of burial. Neither is given directly

by paleomagnetic methods. However, if an independent mineral geobarometer can be found [4], one can calculate the paleogeothermal gradient, which may have important implications for the tectonic mechanism of burial/uplift (e.g. pre-rifting thermal uplift of a plate margin, subduction followed by arc migration, Tibetan-style collision and crustal thickening). Conversely, if a reasonable value can be inferred for paleogeothermal gradient, T_r yields maximum burial depth.

Burial depths of presently exposed Sydney Basin rocks estimated from vitrinite reflectance data range from 0.9 to 2.9 km [1]. Our estimate of burial depth for the Milton region, combining the paleotemperature estimate of 165 ± 30°C with a paleogeothermal gradient for young or incipient rifts of 70°C/km, is 2.3 ± 0.4 km. The remagnetization estimate is in excellent agreement with the deeper estimates from vitrinite data.

8. Conclusions

The thermal remagnetization geothermometer [7,16] has been used mainly with magnetite-bearing rocks and has had two associated problems:

(1) The correction from the demagnetization temperature T_L of a thermal NRM overprint measured in the laboratory to the corresponding temperature T_r experienced in nature during remagnetization is usually large, because the (T, t) contours (Eq. 1 and Fig. 1) are gently sloping when T_r is far below the magnetite Curie point of 580°C.

(2) T_L values measured for multidomain magnetite, the typical grain size range in many rocks, are anomalously high compared to single-domain T_L values assumed in remagnetization theory. This in turn makes the paleotemperature T_r estimates anomalously high. There is unfortunately no simple means of 'correcting' multidomain T_L values [36,37].

The pyrrhotite remagnetization geothermometer presented in this paper (Figs. 1, 4) improves matters considerably. The typical range of remagnetization temperatures is close to the pyrrhotite Curie temperature of 320°C, leading to steep (T, t) contours and a smaller correction from T_L to T_r. Furthermore, single-domain pyrrhotite is much commoner in nature than single-domain magnetite because the pyrrhotite single-domain threshold size is around 1–2 μm, an order of magnitude larger than for magnetite. Therefore the problem of anomalous multidomain values of T_L is much less severe than for magnetite.

We also propose a method of circumventing the problem of anomalously high T_L values in magnetite-bearing samples. Rather than attempting to correct these values by some theoretical relation (a fruitless task, since a single T_B for a multidomain grain has a spectrum of T_{UB} values [36,37]), we have found it effective to pre-treat samples by zero-field cycling to liquid N_2 temperature (low temperature demagnetization or LTD). LTD greatly reduces the NRM of multidomain grains in a mixture of grain sizes, and usually reduces originally high T_L values to single-domain levels [23]. The thermal remagnetization geothermometer is then usable.

Pyrrhotite is not as common a remanence carrier as magnetite, but where it occurs [18,34,35], it is the mineral of choice for remagnetization geothermometry. The pyrrhotite (T, t) contours in Figs. 1 and 4 are much steeper than magnetite (or hematite) contours, and set much narrower limits on burial temperature when minimum and maximum times for remagnetization are constrained by factors like uplift time, coalification, and lengths of polarity chrons. In the ideal situation where two minerals carry the same thermal NRM overprint, a unique (T_r, t_r) can be estimated without other input, but generally geological and geomagnetic constraints are needed to set upper and lower bounds on remagnetization temperature. Because time enters Eq. 1 as a logarithmic factor, the constraints on time do not need to be particularly tight.

The test case presented in this paper demonstrates the usefulness of the technique. The Milton Monzonite and its surroundings are estimated (using mainly overprint data for pyrrhotite) to have been reheated and remagnetized (around 100 Ma BP) at $T_r = 165 \pm 30$°C, in excellent agreement with estimates from vitrinite reflectance data [1]. Former much higher magnetic estimates of T_r [1] were flawed by anomalously high multidomain T_L

values. Using a reasonable estimate for paleogeothermal gradient, the depth of burial is calculated to have been 2.3 ± 0.4 km, again in excellent agreement with vitrinite estimates for this part of the Sydney Basin.

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