Title: **Palaeomagnetic Field Intensity**

Synonyms: Palaeointensity, archaeointensity

Definition:

The strength of the ancient geomagnetic field and it’s variability through time can be deduced from various different types of measurements. *Absolute palaeointensities* are calibrated using a known laboratory magnetic field and can be recovered from any rocks retaining a thermoremanent magnetisation yielding a palaeomagnetic measurement of the geomagnetic field intensity recorded at the time and place that the rock cooled down from above the Curie temperature of the magnetic grains. Similarly, *archaeointensity* measurements are made from heated archaeological artefacts (e.g., fired pottery). By contrast, *relative palaeointensities* provide uncalibrated records of geomagnetic intensity variations measured from the natural remanent magnetisation of sediments and *cosmogenic isotope palaeointensities* achieve the same using concentrations of cosmogenic isotopes as a proxy.

**Introduction**

Direct measurements of the intensity of the geomagnetic field only became possible in 1832 when a method of obtaining them was developed by Carl Frederich Gauss (Stern, 2002). If we wish to know how the field intensity has varied before this time then we are required to use the geological and archaeological records to make indirect measurements. Palaeomagnetic approaches are based on the principle that, for weak fields such as Earth’s, the intensity (*Manc*) of the natural remanent magnetisation (NRM) recorded by a material is linearly proportional to the strength (*Hanc­*) of the palaeofield. For absolute palaeointensity measurements, the material in question must (except in very rare cases; e.g. Games, 1980) have acquired a thermoremanent magnetisation (TRM) whereas, for relative palaeointensity, it is required to be a depositional (DRM). Unfortunately, in both cases, the constant of proportionality linking the NRM to the palaeofield intensity is a sensitive and complicated function of the magnetic properties of the recording material and extremely difficult to measure directly. Absolute palaeointensity experiments obviate this difficulty by imparting the specimen with a new TRM (*Mlab*) in the lab by heating the specimen up and cooling it in a controlled field of known intensity (*Hlab*). The linearity relationships of the two types of *M* and *H* can then be combined to eliminate the constant and solve for the ancient field intensity:

 1.

Depositional-type remanences cannot be reproduced reliably in the laboratory and therefore palaeointensity experiments performed on sediments produce records only of relative *variations* in *Hanc*.

In order for any type of palaeointensity measurement to be useful, an independent estimate for the age of the magnetisation (often assumed or demonstrated to be the same age as the material) is usually required.

The first attempts to measure the ancient field intensity were made by Folgheraiter (1899) using archaeological vases, but with no reported success. Later, Koenigsberger (1936, 1938) then Thellier (1941), made more serious attempts to recover past geomagnetic field strengths, from using old igneous rocks, and baked archaeological materials and a sedimentary contact rock, respectively. It was somewhat later that the idea of Koenigsberger (1936) to obtain palaeointensity measurements from much older igneous rocks by comparing their natural thermoremanence to a laboratory-imparted field was properly tested (Wilson, 1961). Absolute palaeointensity measurements from igneous rocks of all ages began in earnest in the mid to late 1960s (e.g., Briden, 1966; Coe, 1967) and the discipline has been subject to numerous attempts to improve the methodology since (Table 1). Palaeointensity measurements of all types are typically curiosity-driven, although the desire to understand geomagnetic polarity reversal transitions and geodynamo behaviour more generally has provided a clear focus for some studies (Valet, 1993).

**Absolute palaeointensity measurements**

Absolute palaeointensity experiments may only be performed using materials that have cooled from elevated temperatures in nature and which therefore possess a TRM imparted by the ambient (geomagnetic) field. Materials that may be used include any fired archaeological materials (ceramics, kiln lining, hearths, burnt bricks, etc.) and most types of igneous rocks.

Producing unequivocally accurate measurements of the absolute palaeointensity is extremely demanding and requires at least the following:

1. The remanence carriers must retain a pure TRM and have undergone no chemical or physical changes since cooling beneath their Curie Temperature (see e.g., Fabian, 2009).
2. Any secondary components of magnetisation (e.g., viscous or isothermal overprints, see “*Remanent Magnetization*”) must be removed so that the primary TRM is isolated.
3. The remanence carriers must undergo no chemical or physical alteration as a consequence of the laboratory heating that it is required to obtain the palaeointensity (or this must be corrected for; see e.g., Valet et al., 1996).
4. It must be shown that multidomain, vortex state, or magnetically interacting grains are not biasing the measurement by violation of Thellier’s laws of thermoremanence (see “*Rock magnetism*”).
5. Any anisotropy of TRM must be avoided or corrected for (see e.g., Veitch et al., 1984).
6. Any effect that the difference in the cooling rate between the natural and laboratory cooling cycles has on the magnitude on the calculated palaeointensity must be corrected for (see e.g., Perrin, 1998).
7. Any deviation from Thellier’s law of linearity must be detected and corrected for (Selkin et al., 2007).
8. The number (*N*) of individual sample estimates from a single rock unit or archaeological horizon must be *sufficient* to obtain a reasonable mean () and to demonstrate *adequate* internal consistency (see e.g., Paterson et al., 2010).

The high failure rates and time-consuming nature of many absolute palaeointensity studies has stimulated a wealth of innovation in terms of both the types of materials that are used and the manner in which the experiments are performed. In the former respect, two noteworthy materials are submarine basaltic glass (SBG) generally taken from ocean drilling cores and ophiolites (Tauxe, 2006), and single silicate crystals isolated from both extrusive and intrusive rocks (Tarduno et al., 2006). Both of these have been argued to be higher fidelity recorders (less prone to problems numbered 1, 4, and/or 6 in the list above) than more conventional palaeointensity recorders (e.g., whole-rock subaerial lava samples). As an alternative approach to favouring any one specific recorder, the QPI criteria (Biggin and Paterson, 2014) were developed as a universal qualitative measure of palaeointensity reliability. These attempt to emulate the success of a similar tool (Van der Voo, 1990) long-applied to palaeomagnetic poles in deriving a set of qualitative “check boxes” for minimising the risk of bias associated with any of the problems listed above.

 In terms of the techniques developed to measure the absolute palaeointensity, Table 1 summarises some of those which are more widely used and provides the relevant references. The Thellier method and its derivatives remains the most widely used and trusted technique; for more information on this and other techniques listed in Table 1, the reader is directed to comprehensive reviews by Tauxe and Yamazaki (2015) and Valet (2003). Accompanying these techniques is a broad range of statistics that are used as selection criteria to screen less reliable results. For a comprehensive description of these statistics and comparisons of the effectiveness of various sets of selection criteria see Paterson et al., (2015; 2014).

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| **Family** | **Variants** | **Benefits** | **Criticisms** |
| Thellier (Thellier and Thellier, 1959) | Coe (1967); Aitken *et al.* (1988); Perpendicular (Kono and Ueno, 1977); IZZI (Tauxe and Staudigel, 2004) | Sound theoretical basis for single domain grains; potential to recover estimate from a sample that contains both secondary magnetisations at low temperatures and that alters at high temperatures; additional checks for alteration (Prévot et al., 1981) and multidomain behaviour (Krasa et al., 2003; Riisager et al., 2001) can easily be added. | Multiple heatings are time consuming and increase likelihood of sample alteration; bias due to multidomain effects may be prevalent in lava samples (Biggin, 2010). |
| Microwave (Hill and Shaw, 2000; Shaw et al., 1996) | All Thellier-type variants may be used. | In principle, all benefits of Thellier, but with reduced sample heating in terms of both temperature and time; faster experiments that can be individually tailored to samples. | Requires highly specialist equipment (microwave palaeointensity system); theoretical basis not yet firmly established; TRMs may not thermally unblock as sharply and as completely in microwave as in thermal experiments. |
| Shaw (1974) | ARM-alteration corrections (Kono, 1978; Rolph and Shaw, 1985); Double heating (Tsunakawa and Shaw, 1994); low temperature demagnetisation (Yamamoto et al., 2003) | Original method is fast; arguably domain-state independent as uses full TRMs. | The use of ARMs to check and correct for alteration in terms of the TRM properties is not always valid (Tanaka and Komuro, 2009). |
| Continuous high temperature measurement | Wilson (1961);Le Goff and Gallet (2004) | Very fast methods; Wilson method is arguably domain-state independent as it uses full TRMs; Le Goff – Gallet method incorporates cooling rate and anisotropy corrections. | Requires specialist equipment (high temperature magnetometer); samples must survive heating to Curie temperature without altering in Wilson method; unknown effect of multidomain grains in Le Goff – Gallet method. |
| Multispecimen(Hoffman et al., 1989) | Hoffman and Biggin (2005); Dekkers and Böhnel (2006); Fabian and Leonhardt (2010) | Fast; *inter*-specimen consistency is implicitly considered; domain state effects can be minimised in Dekkers-Böhnel experiments; fewer heating steps implies less alteration. | Fewer checks for *intra*-specimen consistency; no internal checks for alteration; domain-state benefits of Dekkers-Böhnel method may be lost if secondary magnetisations require removing (Michalk et al., 2010); original Dekkers-Böhnel method may not be domain state independent, but can be corrected (Fabian and Leonhardt, 2010). |
| Calibrated pseudo-Thellier | Stephenson and Collinson (1974); de Groot et al. (2013); Paterson et al., (2016) | No heating required; fast, especially on automated systems | ARM/TRM ratio is grain size dependent, which manifests as a ~25% uncertainty on the calibration factor (Paterson et al., 2016) |
| IRM Normalisation (Cisowski and Fuller, 1986) | REM (Kletetschka et al., 2000); REM′  (Gattacceca and Rochette, 2004) | No heating required | Precision limited to orders of magnitude  |
| First order reversal curve (FORC) | Muxworthy & Heslop (2011); Muxworthy et al. (2011); Di Chiara et al. (2017) | No heating required; moderately fast | Limited testing; considered accurate to a factor of 2-3 |

Table 1: Selection of absolute palaeointensity methods currently in use.

**Relative Palaeointensity measurements**

Palaeomagnetic analyses of sedimentary rock and soft sediment samples can produce continuous time series of geomagnetic intensity variations which may be calibrated using absolute palaeointensity estimates or used alone as records of geomagnetic behaviour. Relative palaeointensity studies are generally undertaken on the assumption that the samples retain a depositional remanent magnetisation (DRM) acquired as the magnetic grains in the sediment physically align themselves during or shortly after deposition from the water column. The DRM acquisition process is sensitively related to the size, shape and composition of the remanence carriers and is further complicated by flocculation effects in saline environments. Much later, the intensity of the magnetisation can be significantly affected by consolidation and compaction processes (see review of Tauxe and Yamazaki, 2015).

 Some normalisation of the measured magnetisation intensity is required to account for changes in the concentration and nature of the magnetic recorders (which should ideally be small) and produce a record of geomagnetic intensity variations. Candidates for normalisation include ARM, IRM, and low-field susceptibility. These should ideally display a strong coherence with the NRM signal (after removal of any secondary components of magnetisation), but no coherence with the normalised record (Valet, 2003). The reliability of a relative palaeointensity record can be further supported by the presence of antipodal directions (indicating no inclination flattening) and by the coherence of multiple curves from the same region covering over the same time interval (as calibrated using independently-obtained timescales).

 The sedimentation rate and the interval over which the magnetisation is “locked-in” obviously sets a limit on the time resolution of any relative palaeointensity record. In deep-marine sediments, a standard 2.5cm palaeomagnetic sample can represent a time period of 10 kyr or more, which will produce significant smoothing of secular variation (see: *Geomagnetic field, secular variation*). Roberts et al. (2013) provide a detailed discussion of DRM processes, relative paleointensity determinations, and normalization techniques.

**Cosmogenic isotope palaeointensity measurements**

This method of palaeointensity measurement, applicable to the last few hundreds of kyr, is based on the fact that production rate of certain radioisotopes (14C, 36Cl, and 10Be) is strongly influenced by the degree of shielding the geomagnetic field provides the Earth from incoming cosmic rays. Records of the concentration of these isotopes in sediment and ice cores (and tree rings and corals in the case of 14C) can therefore provide a proxy for geomagnetic dipole moment once the necessary corrections for latitude and relevant physical, chemical, and biological processes have been made. See review by Valet (2003) for more information.

**Geophysical applications**

Palaeointensity measurements provide information about the geodynamo across a variety of time scales. Since direct observations of field strength field were first made in the 1830s, the main axial dipole field has fallen by around 10%. The rate and duration of this dipole decay is a focus of many archaeointensity studies, but is still debated. Using VADM values derived directly from archeointensity estimates, Poletti et al. (2018) suggest that the dipole intensity has been decreasing since ~750 CE – over a millennium of weakening. Recent studies based on global field models (e.g., Arneitz et al., 2019; Hellio and Gillet, 2018), however, argue that the axial dipole has been weakening only for the last few centuries with variable trends prior to this.

Archaeointensity and absolute palaeointensity data are particularly important for constructing global field models, which now extend to 100 ka (Panovska et al., 2018). Each of the observable palaeomagnetic field components (declination, inclination and intensity) can be mapped, via their respective Green’s functions, to the radial field at the core mantle boundary (Johnson and Constable, 1997). Measurements of field intensity yield different information than directional data about the core mantle boundary (particularly at higher latitudes; Johnson and Mcfadden, 2015).

Relative palaeointensity records from varved sediments are capable of providing good temporal resolution over the Holocene and agree well with absolute intensity records as well as records of cosmogenic nuclide production (Snowball and Sandgren, 2002). Correspondingly, absolute archaeointensities have been combined with the 14C record from tree rings and the 10Be record from ice-cores to infer variation in solar activity, with implications for understanding solar forcing of climate (Solanki et al., 2004). Marine sediments can provided relative palaeointensity records over longer timescales. Here it is usually assumed that a typical sample will represent a sufficiently long period for non-dipole features to be averaged out (Valet, 2003) allowing composite stacks of relative palaeointensity from different areas of the world to be constructed. A global stack for the past 2 Myr, Sint 2000 (Valet et al., 2005), agrees well with absolute palaeointensity from lava flows and confirms a low field strength during excursions. This has been incorporated into a continuous model of axial dipole moment variation for the past 2 Ma (Ziegler et al., 2011). Importantly, this record is long enough to include reversals and finds a correlation between polarity interval length and field strength., This same correlation is seen, albeit weakly, in the 11 Myr Oligocene record from site Deep Sea Drilling Project, site 522 (Constable et al., 1998) and in SBG absolute intensity records (Tauxe, 2006).

 It becomes increasingly difficult to obtain relative palaeointensities from sediments more than a few million years old so longer term variations in field strength are studied using absolute palaeointensity derived from igneous rocks. The emerging picture characterises the time-averaged dipole moment over the last few hundreds of millions of years as substantially lower than its current instantaneous value, but subject to dramatic variations on timescales ranging from centennial or shorter to at least hundreds of millions of years (Biggin et al., 2012; Tauxe and Yamazaki, 2015).

Prévot et al. (1990) found evidence of a long period of low field intensity which they called the *Mesozoic dipole low.* However, subsequent work (Kulakov et al., 2019; Tarduno et al., 2006; Tauxe and Staudigel, 2004) has indicated that this did not extend into the Cretaceous normal superchron (CNS; 84-121 Ma) and its age of onset remains unclear (Anwar et al., 2016). Other notable lows have recently been identified in the mid-Palaeozoic (Hawkins et al., 2019) and Ediacaran (Bono et al., 2019). The timing of these lows, on the order of 200 million years apart, together with a similar temporal variability seen in Phanerozoic polarity reversal frequency (the dipole lows being coincident with inferred intervals of reversal hyperactivity) gives rise to the hypothesis that dipole moment is quasi-periodic on these long timescales. The likely candidate for the source of such variations is changes in the total heat flowing across the core-mantle boundary and/or its spatial pattern, which might themselves be linked to substantial fluctuations on similar timescales in the global flux of slab material subducted into the mantle (Hounslow et al., 2018).

Claims of even longer, billion-year variations in dipole moment linked to the secular evolution of the core (including inner core nucleation) and mantle remain controversial (Biggin et al., 2015; Bono et al., 2019; Smirnov et al., 2016; Tarduno et al., 2015). Similarly, the magnitude and timing of the expected variations under different core thermal evolution scenarios is not yet clear (Driscoll, 2016; Landeau et al., 2017). Vigorous activity in this potentially fruitful research area is anticipated for years to come.

Palaeointensity experiments are not confined to terrestrial rocks, with numerous studies on lunar, Martian, and meteorite materials. Lunar regolith from the Apollo missions is thought to hold a thermoremanence and a variety of palaeointensity techniques suggest a strong dynamo (20-100 µT) operating between ~4.2 and ~3.6 Gyr (Cournède et al., 2012; Suavet et al., 2013), prior to a possible prolonged period of weak field (~5 µT) that extended to at least ~2.5 Ga (Tikoo et al., 2017). Martian meteorites have been used to estimate the ancient Martian field (Shaw et al., 2001; Weiss et al., 2008b) and paleointensity studies on pallasite meteorites indicate that small (~200 km) planetesimals are also capable of sustaining dynamo generated magnetic fields with field strengths of ~30-130 µT (Bryson et al., 2015; Tarduno et al., 2012). Some of the most ancient meteorites, formed at the onset of our solar system, are also capable of remembering magnetic fields in the solar nebula, yielding insight into solar system evolution (Wang et al., 2017; Weiss et al., 2008a).

From 450 year old pots to 4.5 billion year old meteorites, studies on the strength of Earth’s ancient magentic field contiune to florish, bringing with them new innovations in methods, materials and analyses. And as paleointensity studies grow we are gaining new understanding on the evolution of the magnetosphere in near-Earth space, the planet beneath our feet and the solar system beyond.

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**Cross-references**

Archeomagnetism

Curie temperature

Geomagnetic Field, Excursion

Geomagnetic Field, Polarity Reversals

Geomagnetic field, secular variation

Palaeomagnetism, Instrumentation

Remanent Magnetization

Rock magnetism