Paleomagnetism in tectonics: A user's guide

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This is a non-peer reviewed manuscript. It has been submitted as book chapter as:
Fu, R.R., van Hinsbergen, D.J.J., submitted, Paleomagnetism in tectonics: a user's guide. In: Webb, A.A.R., Briggs, S.M., Ganbat, A., Genge, M.C., and Moore, W.B. (eds), Methods in Tectonics. Geological Society, London, Geoscience in Practice.

Abstract

Paleomagnetism can reveal the direction of the ancient geomagnetic field recorded in rocks and provides an invaluable quantitative reference for the study of tectonics. This chapter describes the basic principles of how the magnetic field is stored in rocks, how it may be measured, and how common pitfalls in paleomagnetic analysis may be recognized and corrected for. We then explain how paleomagnetic data of coherent tectonic units, such as continents, plates, or plate circuits, may be combined into apparent polar wander paths that tracked the 'absolute' motions of tectonic plates or continents relative to the Earth's spin axis. Finally, we provide a practical guide how relative paleomagnetic displacements may be computed from such absolute positional or motion constraints, and how these provide quantitative constraints on (plate) tectonic motions of deformed crust.

Introduction

A key challenge in the study of tectonic evolution is quantifying the motions that resulted in today's deformed state of the Earth's crust, and to reconstruct paleogeographic evolution. Geological mapping and structural geological analysis may allow restoring rocks to their undeformed state, providing essential constraints on the timing, duration, direction, and nature of plate tectonic motion (Cawood et al., 2009; van Hinsbergen and Schouten, 2021), but field geological observations are essentially blind to the vast tracts of lithosphere that were lost to subduction and to relative motions between crustal blocks where the zone of direct contact is not preserved for study. The only method that we have available to quantify relative and absolute tectonic motions in the geological past is provided by paleomagnetism - the study of the Earth's magnetic field recorded in rocks (Butler, 1992; Tauxe, 2010). However, despite the unique possibilities provided by paleomagnetism, there are many interpretative steps between paleomagnetic observation and tectonic interpretation, which makes the technique vulnerable to confusion and controversy.

Paleomagnetism constrains tectonic motion in two ways. First, since the 1960s, magnetic reversals recognized on the ocean floor that are mirrored in mid-ocean ridges have been used to reconstruct oceanic crustal accretion during ocean spreading (Vine and Matthews, 1963). The interpretation of magnetic reversals and fraction zones on the seafloor is relatively straightforward, and decades of detailed mapping by the marine geophysical community has led to detailed ocean floor spreading reconstructions (Seton et al., 2020). This reveals the opening history of most of the ocean basins back to early Jurassic times, which also reveal net plate convergence and lithospheric area lost to subduction (Seton et al., 2012). Even though marine magnetic anomaly research is still improving the detailed reconstruction of plate motion (DeMets and Merkouriev, 2021; Gürer et al., 2022; Dalton et al., 2022), global plate models based on magnetic anomalies are reproducible and largely not controversial. However, the second application of paleomagnetism - the reconstruction of apparent polar wander, which quantifies motion of crustal blocks relative to the Earth's pole - is not devoid of controversy.

If the Earth's magnetic field is, on geological timescales, a dipole that aligns with the spin axis – the geocentric axial dipole (GAD) hypothesis – paleomagnetism can quantify two types of motion. First, the deviation of the paleomagnetic north pole stored in the rock from true north, called the declination, reveals net vertical axis rotation since the rock's acquisition of the magnetic signal. Second, the plunge of the paleomagnetic field, called the inclination, depends on latitude: it gradually changes from vertical on the poles to horizontal at the equator. The inclination stored in a rock thus quantifies the paleolatitude at which its magnetization was acquired. This way, paleomagnetic data place global plate reconstructions in a reference frame against the Earth's spin axis (e.g., Besse and Courtillot, 2002; Torsvik et al., 2012), which provide an indispensable, quantitative basis for paleoclimate and paleobiology (Saupe, 2023; Heath et al., 2025).

On a more regional scale of the tectonic evolution of orogens, discrepancies between paleomagnetically predicted net motion and geologically recorded strain can underscore the incompleteness of the geological record and offer opportunities to learn fundamental lessons about how past tectonics are encoded in rocks (or not). However, perhaps confusingly, paleomagnetists do not always seem to produce mutually concordant results, and paleolatitude estimates for rocks from the same time interval sometimes differ by >10°, or >1000 km (e.g., Chen et al., 2010 vs. Huang et al., 2015; Kulakov et al., 2021 vs. Hou et al., 2024). In these examples, the nominal 95% uncertainties of individual paleomagnetic poles are between 5° and 7°.

We therefore build this chapter around the question, "If paleomagnetism works, how can published paleomagnetic poles of the same age and on the same landmass disagree?" We begin with the point that paleomagnetic errors stem from different sources, some of which cannot be neatly encapsulated in a classic error bar but must nonetheless be considered. Having established how paleomagnetic data can go wrong, we explain what paleomagnetists can do to mitigate these errors, in the process describing the theory of paleomagnetic recording and how it functions in an ideal scenario. Finally, we issue recommendations for the non-specialist reader in assessing the reliability of a paleomagnetic result. Throughout, we will illustrate these points with some examples from ongoing debates in the tectonic literature.

An overview of paleomagnetic error

The fundamental data type we consider in this chapter is the paleomagnetic pole. This is the apparent position of the Earth's magnetic pole, assumed to align with the spin axis, relative to the sampled location. We first briefly describe this mathematical concept. Each paleomagnetic pole is based on an average of magnetization directions measured from a collection of oriented rock samples. Each sample-derived magnetization direction is a three-dimensional vector that is typically described by a declination (azimuth from north) and inclination (deviation from horizontal with downward plunge positive). Combined with the sampling location, the implied magnetic pole location, along with an uncertainty based on the scatter of individual specimens directions, can be computed and represented as a latitude and longitude of the mean direction accompanied by a 95% confidence ellipse, known as the A_{95} (Fisher, 1953).

If young rocks of the last few million years are sampled, the paleomagnetic pole should coincide with the modern geographic north pole. But if (plate) tectonic motions have affected the sampled rocks since they acquired their magnetization, the paleomagnetic pole will deviate from the true pole. Paleomagnetic poles representing a succession of ages from the same landmass (or rotated to the reference frame of the same landmass using independent constraints such as seafloor magnetic anomalies) may then show progressive, apparent motion of the magnetic pole. Such as series of poles is the basis for an apparent polar wander path (APWP), which describes vertical axis rotations and paleolatitudinal motion of a landmass through time (see below; Besse and Courtillot, 2002; Torsvik et al., 2012; Vaes et al., 2023).

Confidence in the validity of the APWP ultimately depends on the reliability and the quantitative uncertainties of each included pole. In this section, we detail the sources of uncertainty affecting a pole, emphasizing that the A_{95} alone may not contain the full story.

Beside the direction itself, the age of a paleomagnetic pole must be reliably known before it can be useful in any kind of tectonic reconstruction. Therefore, error on a paleomagnetic pole can be divided into directional and chronological types- the former describes the uncertainty on the position of the pole while the latter describes the validity of the nominal assigned age of the pole. Multiple specific sources, listed in Table 1, contribute to both errors.

These sources of paleomagnetic uncertainty are further classified as random zero-mean errors (i.e., scatter) and biases. It is important to realize that the classic α_{95} uncertainty circle ascribed to paleomagnetic poles accounts for exactly one of the four classifications of errors introduced here: namely, it represents the random, zero-mean component of the directional uncertainty (lines 1-3 in Table 1) while excluding potential biases to the direction and any chronological uncertainty (lines 4-7 in Table 1).

Therefore, the size of apparent error ellipses may not be a sufficient indicator of a paleomagnetic pole's reliability. In fact, large A_{95} poles would be preferable over ostensibly precise, but biased or remagnetized poles for tectonic reconstructions. As a further consequence, even for reliable poles free from wholesale remagnetization, the nominal A_{95} circle may underestimate the true uncertainty. The simple but consequential observation that far fewer than 95% of paleomagnetic data that were used to calculate widely used APWPs actually statistically coincide with the APWP at the claimed age highlights this discrepancy (Rowley, 2019). As we will see, more careful treatment of error propagation can mitigate this issue (Vaes et al., 2023). In the following sections we summarize the theory behind paleomagnetic recording and preservation, explaining how each of the seven error sources enumerated here arise from different stages of the process.

From mineral grains to paleomagnetic pole

The paleomagnetic recording process

We begin by discussing the two error types that affect all paleomagnetic samples: random errors due to magnetic recording and analytical uncertainty (Fig. 1; lines 1-2 in Table 1). The fact that any natural rock can retain memory of a magnetic field during a short time span in the ancient past is itself remarkable and depends on a subtle, statistical alignment of ferromagnetic grain magnetizations within the rock. In Earth rocks, the most common ferromagnetic minerals are magnetite (Fe_3O_4), hematite (Fe_2O_3), and several Fe-sulfide minerals (Dunlop and Ozdemir, 1997, pp. 45-82). Quantum mechanical interactions in the crystal lattices of these minerals result in a coherent pointing direction of their electron spins, which results in a net magnetic moment. This effectively transforms each mineral grain into a diminutive, magnetized compass needle.

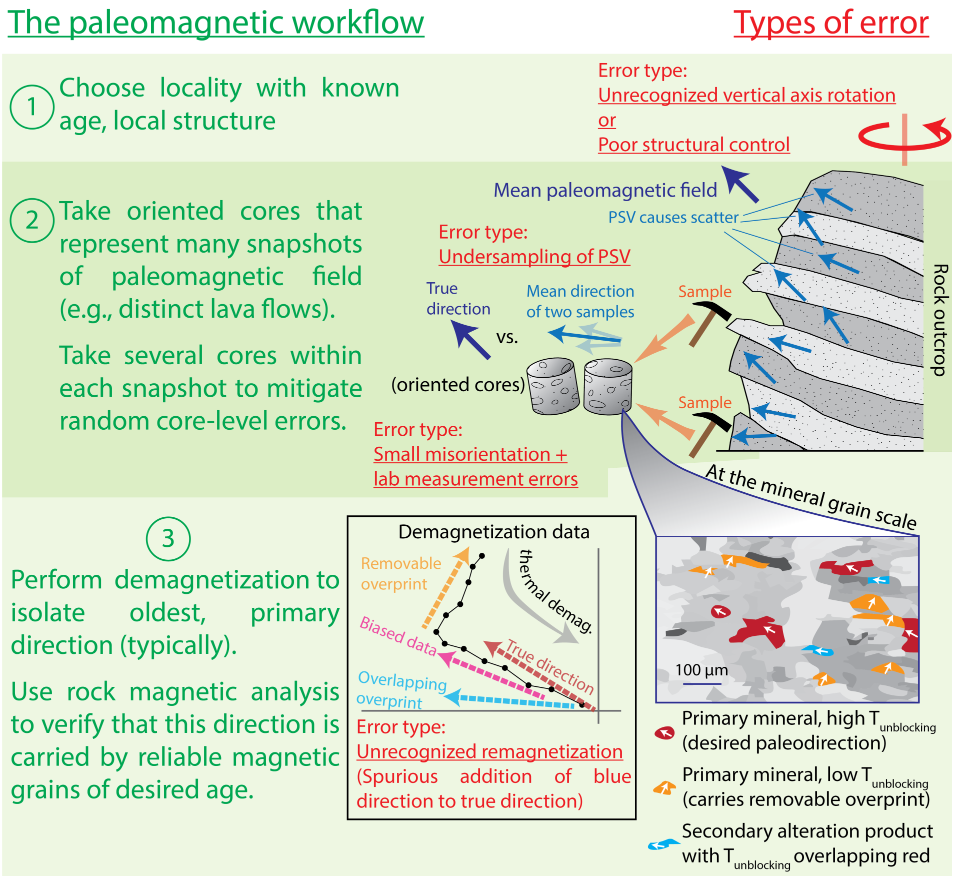

Figure 1: The basic paleomagnetic workflow (green text at left) and typical sources of error at each step (red text). The choice of sampling locality is a key first step. For reconstructing a paleopole for the plate, local rotation must be avoided or known. In any case, local structure (e.g., location of faults) should be known even for reconstructing relative block motion. Undersampling of paleosecular variation can increasing uncertainty (see Fig. 2 and text), while unrecognized overprints can lead to biased directions.

Paradoxically, recording an ancient magnetic field requires that these mineral compass needles be both unstable and extremely stable, depending on ambient condition. A rock initially becomes magnetized when some process destabilizes the magnetic pointing direction of its constituent ferromagnetic minerals, thereby allowing these directions to align to the ambient magnetic field. Three such processes are commonly observed in natural rocks. The first and best-studied phenomenon is heat. With sufficient temperature, thermal vibrations become strong enough to perturb the magnetic moment of individual ferromagnetic grains, essentially converting them to a soup of randomly fluctuating directions. When an ambient magnetic field is applied to such an ensemble of freely spinning magnetic moments, their directions will become slightly biased to be parallel to the applied field. The bias is subtle: in an ensemble of 200 grains affected by the Earth’s field, approximately 101 and 99 are expected to be aligned and anti-aligned, on average. Cooling of each grain below a critical point, called the blocking temperature ($T_{blocking}$), once again stabilizes the grain’s magnetization, thereby preserving this small overall bias in the rock. Virtually all newly formed igneous rocks on Earth are therefore born with a “thermoremanent” magnetization recording the magnetic field during cooling.

Second, chemical alteration is a common source of magnetic recording in both sedimentary and igneous rocks. Newly nucleated crystals of magnetite or other ferromagnetic minerals have inherently unstable magnetizations due to their size. As such, their magnetic directions fluctuate erratically much like minerals at high temperature until these directions, along with a bias towards the ambient magnetic field, are locked in when they grow to larger sizes (Dunlop and Ozdemir, 1997, pp. 367-387). A third source of magnetization is detrital remanence, where freely rotating ferromagnetic particles sinking through the water column, align with the ambient magnetic field and are locking into position upon burial. This process is most efficient for sedimentary rocks with silt or finer grain sizes (Dunlop and Ozdemir, 1997, pp. 425-432).

In short, cooling from high temperature, chemical alteration, and the deposition of sediment can lock in subtle biases in the magnetization directions of a ferromagnetic grain population. Once brought to ambient temperatures and protected from further reactions, the direction of ferromagnetic grains can remain incredibly stable. The recent decade has seen the development of numerical models that, for the first time, simulate realistically how ferromagnetic grains behave over million-to-billion-year timescales (Nagy et al., 2017, 2019). These studies have reproducibly concluded that, in a large range of sizes between a few 10s nm to more than 1 μm, magnetite grains at near-Earth-surface temperatures preserve stable magnetization over the lifetime of the solar system.

How, then, do errors creep into these magnetization acquisition mechanisms? First, the subtleness of the directional bias created by typical geomagnetic field strengths limits the fidelity of paleomagnetic recording. If, in an ensemble of 200 grains as described above, only a single extra grain is expected to align parallel to the ambient magnetic field on average, then any single ensemble of 200 grains can easily exhibit a stronger bias or a reversed bias. In fact, calculations show that $\geq 10^7$ grains are necessary to reproduce an ambient magnetic field with ~1˚ accuracy (Berndt et al., 2016). Fortunately, conventional paleomagnetic specimens, which are drill cores of a ~25 mm diameter and ~2 cm long, typically contain more than sufficient ferromagnetic grains to achieve <1˚ statistical errors.

However, other specimens-level effects may contribute more significant error. Anisotropy arises when the magnetization of individual grains responds to the magnetic fields of their neighbors. In rocks with significant fabric the magnetization of the whole specimen may be deflected by several degrees towards a magnetic "easy" direction. Vigilant rock magnetic work, however, can readily quantify this deflection and apply an effective correction (Jackson, 1991). Further, paleomagnetists can attempt to sample rocks that are as magnetically isotropic as possible.

Finally, small, unavoidable misorientation of drill cores both in the field and in the lab, typically on the order of a few degrees, can contribute scatter. However, because these zero-mean, random errors are applicable to the single specimen level, they tend to mildly increase the scatter, but not the ultimate average, due to typical paleomagnetic practice of taking many samples. Although the calculation of A_{95} intervals do not explicitly propagate specimens-level errors, averaging of many dozens to hundreds of specimens typical for a paleomagnetic pole is likely to bring the overall contribution from these errors to the <1° level (small purple circle in Fig. 1).

Paleosecular variation

The preceding section described the mechanisms and the accuracy to which rock can record ambient magnetic fields. The next question is whether the recorded direction, and therefore the inferred paleomagnetic pole, reflects the true location of the Earth's spin axis. In historical times, the geomagnetic poles, defined as the points where the axis of the dipole component of the geomagnetic field intersects the Earth's surface, are constantly wandering with speeds up to 0.5° per year. This steady drift of the geomagnetic pole, which results from time-varying non-dipolar contributions to the geomagnetic field, is known as paleosecular variation (PSV). Therefore, a fast-cooling lava flow does not record the position of the Earth's spin axis but only approximates it within some tens of degrees, which is barely useful for tectonic interpretation.

A tectonically useful paleomagnetic dataset, therefore, must contain sufficient independent measurements of the geomagnetic field, over enough time, such that their average is reasonably close to the true geographic pole (Fig. 1). Fortunately, the detailed study of the recent magnetic field (e.g., Cromwell et al., 2018) and, on longer timescales, the comparison of paleomagnetic data to independent paleolatitude proxies such as evaporite deposits (Evans, 2006) appear to confirm that, on time scales of hundreds of thousands of years or more, the geodynamo aligns with the Earth's spin axis. PSV is therefore expected to contribute random scatter, not a bias, to paleomagnetic poles.

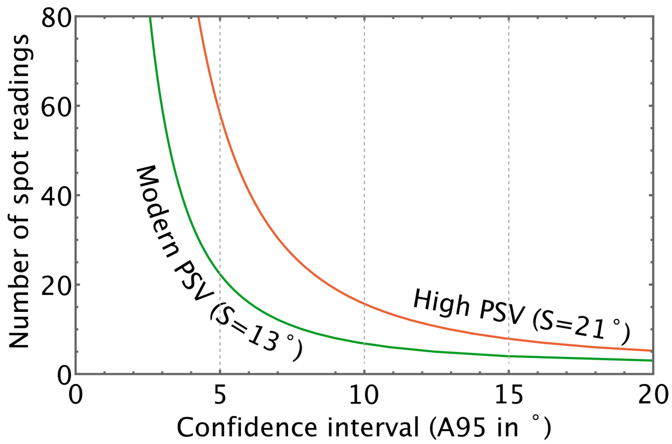

Figure 2: Number of spot readings of the geomagnetic field (e.g., igneous cooling units or single rapidly deposited sedimentary horizons) to achieve a specified paleomagnetic pole accuracy for a given paleosecular variation (PSV) regime.

What is the magnitude of this scatter? Because PSV causes magnetic field drift over only decadal to millennial timescales, each single lava in a stack emplaced over many thousands to millions of years is essentially taking a random, independent sample (a 'spot reading') of the instantaneous magnetic field direction. Compilations of large numbers of individual lava flows suggest that the amplitude of PSV, defined as the angular standard deviation of the probability density function describing the instantaneous location of the geomagnetic pole, has remained in the 10˚-30˚ range (depending on latitude) over the past 3 Gy (Biggin et al., 2020; Engbers et al., 2024). Under such typical geodynamo conditions, a sample size of 20 spot readings is sufficient to decrease PSV-induced error on the paleopole to 5˚ (95% confidence interval; Fig. 2), although the precise number of spot readings needed also depends on the latitude (see Fig. 4 in Deenen et al., 2011). Sedimentary rocks, due to their gradual accumulation, may already average some PSV. For example, a 2.5 cm core of a deep-sea sediment with deposition rate of 1 cm per thousand years could average a time of ~2.5 ky, reducing the expected scatter. However, approximating the true geographic pole requires averaging over 10^5 y timescales (Davies and Constable, 2014), and conservatively assuming that a sediment specimen represents a spot reading of the paleomagnetic field, has shown to be a good approximation (Tauxe and Kent, 2004; Vaes et al., 2021; Tauxe et al., 2024).

Regardless of magnitude, error can be tolerable if its magnitude is well-quantified and communicated. A fundamental problem across the (Earth) sciences is that datasets are traditionally poorly documented in databases, and paleomagnetism is no exception. Most legacy data has been lost, and only statistical descriptions of these data remain in papers. Even today, measurements contributing to paleomagnetic results are often not shared in the available community databases. Fortunately, of all the sources described here, PSV-induced error is the most explicitly represented by the traditional A_{95} uncertainty interval (Fig. 3). This is because A_{95} is computed from the scatter of directions from individual lava flows or sedimentary horizons, much of which likely stems from PSV. Therefore, although uncertainty in the location of the paleomagnetic pole PSV is unavoidable, its effect is well-accounted for in standard

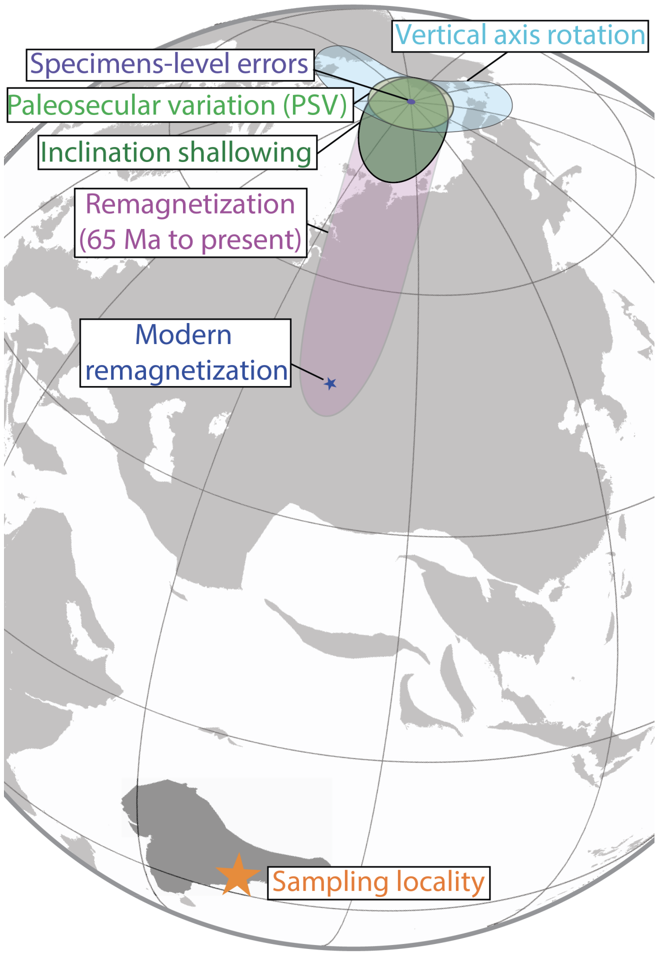

Figure 3: Visual summary of different sources of error affecting paleomagnetic poles. Map shows reconstruction of Indian subcontinent at 65 Ma (Merdith et al. 2021) and the approximate 95% confidence regions for a paleomagnetic pole at that age contributed by each error source. Estimates assume 20 paleomagnetic sites (e.g., lava flows), each of which consists of 5 specimens recording an instantaneous magnetic field. We assume modern-like PSV amplitude ($S = 13^{\circ}$), sedimentary flattening factors between 1.0 and 0.6, and vertical axis rotation of up to 10˚. The elongated "Remagnetization" ellipse is based on hypothetical overprinting between 65 Ma and the present, potentially resulting in any pole position between the true pole and the one recorded if the rock were remagnetized at the locality's modern latitude of 22˚N. If full remagnetized occurred, the resulting pole would be found somewhere in this purple-shaded region, depending on the age of remagnetization. See text and Table 1 for more information.

paleomagnetic data analysis and can even be used to evaluate whether a dataset reliably sampled the paleomagnetic field (Tauxe et al., 2024).

Inclination shallowing in sedimentary rocks

From the above, it follows that the tectonic paleomagnetist aims to collect many spot readings of the ancient field from a succession of rocks spanning $\geq 10^5$ y to average PSV, but not so long as to smooth out tectonic motions (≥5 My). Fast-cooled lavas are often preferred by paleomagnetists because they are typically better magnetic recorders than sediments and may be less susceptible to alteration due to lower porosity. However, the discontinuous nature of igneous rock emplacement means that purely volcanic successions with the necessary time coverage are not available for most time intervals in Earth history. Further, episodic lava emplacement complicates magnetostratigraphic dating and interpretation of any interbedded or bounding radiometric ages. Lastly, while mafic lavas may have formed sub-horizontally, andesitic lavas from stratovolcanoes may have primary dips of 20° or more, adding another source of potential bias.

Sedimentary rocks, particularly fine-grained, isotropic rock such as fluvial redbeds or clay-rich marine sediments, are widespread in both space and time, cover their time intervals more continuously, and allow confident reconstruction of the paleohorizontal. In many ways, sediments, if they escaped remagnetization, are better targets for tectonic paleomagnetism. But they have one systematic and influential bias that lavas do not have: they get compacted.

As explained above, during sedimentation the magnetic field is recorded in sedimentary rocks by magnetic particles. Those may be magnetic minerals eroded from elsewhere or magnetotactic bacteria from the water column (Mann et al., 1990; Vasiliev et al., 2008). Once sediment forms and locks in magnetization, it compacts by undergoing vertical shortening, which also deforms the original paleomagnetic inclination, making it flatter (Fig. 3). Compaction does not significantly influence the declination, and therefore, sediments are widely used for tectonic rotation studies (e.g., across the Mediterranean region; van Hinsbergen et al., 2020). However, when unaccounted for, this "inclination shallowing" systematically leads to underestimation of the paleolatitude at which the sediment was deposited. Flattening factors of 0.6 (i.e., the modern thickness is 60% of the original) are common in fine-grained clastic sediments and are typically less severe in carbonates (~0.7-1.0; Vaes et al., 2021). Some rocks have documented flattening factors as low as 0.4 (Arason and Levi, 1990). Such a bias may underestimate paleolatitudes by 10s of degrees, or 1000s of km, in the mid-latitudes.

Although inclination shallowing underpinned many tectonic paleomagnetic controversies in the past, several modern methods exist to correct for inclination shallowing. These include studies of the rock fabric using magnetic anisotropy measurements (Jackson et al., 1991; Tan et al., 2003), but the most widely used is the Elongation/Inclination, or E/I method, of Tauxe and Kent (2004) (also see Tauxe et al., 2008).

The E/I method uses the predictability of PSV-induced scatter. As explained above, a large collection of spot readings of the paleomagnetic field will form a point cloud that is circularly distributed around the paleomagnetic pole. The associated scatter of local paleomagnetic directions in terms of inclination and declination forms an ellipse, the shape of which varies with latitude. In compacted sedimentary rocks, the inclination, and therefore the scatter ellipse, becomes deformed, causing mismatch between the observed mean inclination and the expected shape, or elongation, of the scatter ellipse. The Tauxe and Kent (2004) E/I method bootstraps the dataset to estimate the compaction factors required to bring the two observations into agreement. This method assumes that each sedimentary specimen captures a spot reading of the field and that all scatter is the result of PSV. Several test cases have shown that the E/I and anisotropy methods reproduce the same inclination shallowing in a wide variety of sediments (Huang et al., 2015b). Recently, Vaes et al. (2021) showed that, when the scatter of a sedimentary dataset falls within the range expected from PSV and datasets are large enough (>80 samples), inclinations reliably restore to independently constrained values.

Paleomagnetic datasets from thick sedimentary sections, accurately dated through integrated stratigraphy, may thus be ideal for constraining apparent polar wander at very high resolution on par with records of Cenozoic paleoclimate. However, much work needs to be done: the correction of paleomagnetic data from sedimentary rocks for inclination shallowing requires access to the original measurements. Those measurements, or even the directions interpreted from those, have almost never been published and community databases are highly incomplete. The vast majority of legacy paleomagnetic data from sedimentary rocks, which comprise for instance almost the entire Phanerozoic of the China Blocks, can therefore not be upgraded to modern standards and need to be resampled.

Remagnetization

Now we come to the most pernicious, least predictable form of paleomagnetic error. Metamorphism or fluid alteration that affect a rock millions of years after its formation routinely overprint the initial magnetization. As explained above, rocks typically contain tens of millions of diminutive "magnets" some of which are easier to realign than others. As a result, almost every rock that carries an original, primary magnetization also has magnetic minerals that carry a younger, "overprint" direction. To account for this, standard paleomagnetic demagnetization procedures involve stepwise randomization of the magnetic field in a specimen by heating them at increasing temperatures or applying increasing alternating fields. In general, magnetic minerals that are easier to remagnetize are also easier to demagnetize; therefore, this stepwise demagnetization usually succeeds in separating distinct "components" of the magnetization, each of which represents the magnetic field at a point in time (Kirschvink, 1980). Therefore, although remagnetizations are ubiquitous, their mere presence does not preclude the recovery of stable, more ancient paleomagnetic directions.

Sometimes, however, remagnetization can realign even the most stable grains, thereby completely replacing the original paleomagnetic record. A paleomagnetic pole computed from such an overprinting magnetization would correspond to the location of the geographic pole at the time of the remagnetization event, which can be any time between rock formation and the present-day (elongated purple region in Fig. 3). Therefore, unlike the sources of uncertainty described above, remagnetization can result in error of an essentially arbitrary magnitude, with any point along the apparent polar wander path of the landmass post-dating the initial age of rock being a potential location for the remagnetized pole.

Therefore, although remagnetization resets the direction of a paleomagnetic pole, its consequence can be thought of as an incorrect assignment of the pole's age. The possibility of remagnetization forces us to consider whether the age of the magnetization is decoupled from that of the rock. An additional complication is that if a rock underwent a tectonic tilt, it needs to be established whether the magnetization was acquired before, during, or after tilting, which adds even more potential bias to the paleomagnetic pole.

A foundational responsibility of any paleomagnetic study, therefore, is to establish the age of magnetization above and beyond the simple assumption that it is equal to the formation age of the rock. The most powerful tool used to achieve this is the paleomagnetic field test. Although subtle variations exist, the basic types are the conglomerate, fold, baked contact, and reversal tests; these have formed the backbone of paleomagnetic studies seeking to demonstrate the old age of their magnetic signal since the founding of the field in the 1950s. Other more detailed reviews of these tests exist (Meert et al., 2020); therefore, we provide only a brief description. In the conglomerate test, the magnetizations of clasts in single unit are measured and compared. Randomly oriented magnetizations represent a passing test while uniformity in their directions points to post-depositional remagnetization. The fold test compares nearly co-eval magnetizations with different bedding attitudes. Agreement in the directions after they are restored to horizontal bedding provides evidence for magnetization age older than that of folding. In the baked contact test, an intrusion or lava, along with a profile of samples in the country rock, are examined. Rocks that have escaped post-intrusion remagnetization should show a distinctive baked profile with country rock nearest the intrusion exhibiting the same magnetization direction as the intruding body. In the reversal test the observation of stratigraphically consistent samples with normal and reversed magnetization polarity is interpreted as evidence against wholesale remagnetization of the succession, although some overprinting magnetizations can show dual polarity. Finally, we may test whether the scatter of a series of paleomagnetic directions exhibit the expected amount of PSV as an additional check on the pristinity of the paleomagnetic directions (Tauxe et al., 2024).

The strength of these field tests lies in their simplicity, as they rely on simple geometric relationship akin to the principle of superposition. An important take-away is that passing each test does not imply the same constraint on the age of magnetization. In particular, the power of passing fold and baked contact tests is greatly enhanced if the age of folding or intrusion is close to that of original rock emplacement. In addition, it is rare that all tests can be applied to a paleomagnetic dataset, and passing a single test may not demonstrate primary magnetization. For instance, a fold test is largely irrelevant if folding post-dates the most likely episodes of remagnetization. Paleomagnetists therefore typically also apply detailed rock magnetic and microscopic investigations to evaluate whether the magnetic mineralogy forms part of the original rock or may have formed by alteration. For example, recent work using the quantum diamond microscope (QDM), an emerging magnetic field imaging technology, has determined the mineral carriers of altered Archean basalts and argued that their magnetizations were acquired during shallow seafloor alteration soon after igneous emplacement (Brenner et al., 2022, 2024).

Regional tectonic interpretations using an APWP

Since the 1960s, large amounts of paleomagnetic data have been published from all continents and plates spanning the Archean to the Holocene. Because rocks at any one locality are severely limited in time coverage, datasets must be combined and somehow averaged to quantify the motion of landmasses through time. Paleomagnetic data from a single, rigid plate, can all be used to compute the apparent pole of that plate throughout its existence. Paleomagnetic data from a neighboring tectonic plate are offset because of relative plate motion. However, if this relative motion can confidently reconstructed using seafloor magnetic lineations, this motion can be corrected for, and a single apparent polar wander path can be constructed from all plates tied together through ocean basin reconstructions: a global APWP (Besse and Courtillot, 2002; Kent and Irving, 2010; Torsvik et al., 2012; Vaes et al., 2023). In this section we discuss the statistical theory and recent innovations in constructing these consolidated records of motion.

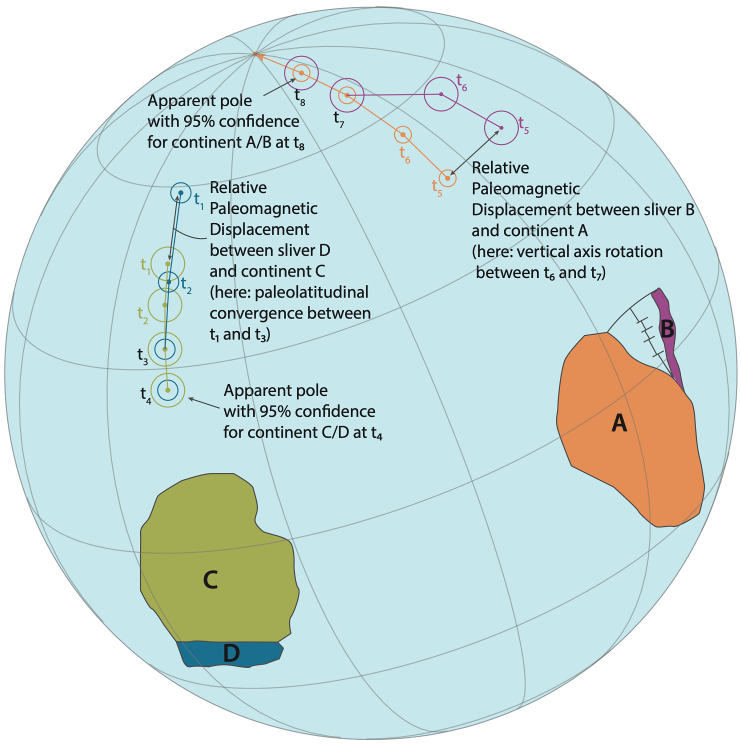

Figure 4: Schematic of how relative plate motions manifest in apparent polar wander paths (APWPs). On the left, latitudinal convergence between two landmasses between t_2 and t_3 results in coincidence of co-eval paleomagnetic poles by t_3. On the right, a relative vertical axis rotation brings poles together by t_7.

As explained above, paleomagnetic data from a single study of rocks of a particular location and narrow age range are routinely combined into a paleomagnetic pole that summarizes the finding of that study. The number of single data points feeding into that pole varies widely, and larger datasets tend to have smaller uncertainties than smaller ones if both contain a scatter that represents PSV. In addition, each paleomagnetic pole comes with an age uncertainty (Table 1). Those uncertainties need to be considered and propagated into the final APWP, but as the custom developed in combining paleomagnetic poles, those uncertainties have been ignored.

The philosophy of making APWPs started in the 1950's. Creer et al. (1954) made an apparent polar wander path for the British Isles, essentially by connecting the dots between single poles of different ages. In those cases, every paleomagnetic "pole" consisted of data from a single study and consisted of an average of many paleofield spot readings. As the paleomagnetic database grew, published poles of similar age needed to be somehow combined. The practice became to compute APWPs by averaging paleomagnetic poles instead of the underlying spot readings, and to determine a A95 cone of confidence from the scatter of those poles. The classical and widely used APWPs consisted of tie points with the average position and average age of their constituent poles, which were chosen to lie within a typically 10-20 My age range. In other words, a paleomagnetic pole with an age of 136±17 Ma would have been used as a datapoint in a 140±10 and a 130±10 Ma age window even though it could have been 119 or 153 Ma old. Age or position uncertainty of individual poles was not considered, with the implicit assumption that the scatter among poles reflects these underlying uncertainties. This also meant that poles based on a handful or hundreds of datapoints were weighted equally.

Resulting APWPs were updated over the decades, and the differences between the path's courses varied little despite growing databases and increased age control. A major problem was articulated by Rowley (2019), who noted that, using the global APWP of Torsvik et al. (2012) as an example, almost half of the paleomagnetic poles that were used to compute the APWP were inconsistent with the coeval tie point on the APWP at the $p = 0.05$ level. Theoretically, one would expect that 95% of these poles to be unresolvable from the APWP at this level. Clearly, the confidence intervals of the poles and the consolidated APWPs are mismatched.

Vaes et al. (2022) identified the sources of that error. First, the A_{95} of the classical APWPs and the A_{95} of a study pole, which are used to compute the uncertainty of the difference, represent different statistical levels: a set of poles versus a set of spot readings. Second, because the number of spot readings to define a pole is study-dependent and arbitrary, an APWP tie point A_{95} computed with only study-level pole data is missing critical information about the statistical strength of the underlying data. Vaes et al. (2022) showed that the distance of poles from an APWP is primarily a function of the number of these underlying data, and, therefore, developed a global APWP for the last 320 Ma based on site-level data. This APWP calculation method gives a unique solution for any collection of paleomagnetic data and quantitatively propagates age and position uncertainty in poles into the final APWP. Moreover, the uncertainty of this global APWP is much smaller than previous renditions, because it is based on a much larger data set.

A practical guide to interpret tectonic motions from paleomagnetic data

APWPs provide an 'absolute' reference frame: they indicate how a stable continent, plate, or plate circuit moved relative to the geomagnetic field, which on geological timescales coincides with the Earth's spin axis. Such absolute motions provide the basis for the study of e.g. paleoclimate, but the study of tectonic motions requires the computation of relative displacements, in terms of convergence and divergence. Those may then be compared to geological records of shortening or extension, respectively, and interpreted in terms of plate tectonic history (e.g., van Hinsbergen et al., 2012). In other words, for tectonic purposes, “relative” paleomagnetic displacements need to be computed from “absolute” APWPs, or individual poles, of different tectonic units, whereby age and position uncertainty needs to be considered.

To this end, Vaes et al. (2024) developed an online tool - apwp-online.org - to compute “absolute” APWPs from sets of paleomagnetic directions with an age range (poles), and relative paleomagnetic displacements (RPD). RPDs may be computed as a difference between two APWPs, or the difference between an APWP and paleomagnetic poles. These differences are then translated into differential paleolatitudinal motion and vertical axis rotation. The apwp-online.org tool uses a bootstrap method of Vaes et al. (2022) to compute a reference tie point, using the same dataset behind the APWP, that has the same age and number of spot readings as a selected study pole or a study APWP. They showed that, with that method, ~95% of the data behind the APWP are statistically indistinguishable at $p = 0.05$, implying that a pole found to be significantly displaced using this method can be interpreted as a tectonic signal.

These final steps allow using paleomagnetic data as quantitative kinematic constraint on tectonic problems. Examples applications include demonstrating of relative tectonic motions using the difference between APWPs of different continents, e.g. during the Archean, as an argument for the existence of plate tectonics at that time (e.g., Brenner et al., 2022), or during the Proterozoic during the assembly and breakup of supercontinents (Evans, 2013). In orogenic belts, local deformation may obscure the vertical axis rotation component of past plate motion, but paleolatitudinal motion may still be helpful in reconstructing ocean opening and closure, for instance for continental blocks and intra-oceanic arcs surrounded by sutures between India and Asia (e.g., Westerweel et al., 2019; Martin et al., 2020), or in the Cordilleran orogen (e.g., Johnston, 2008; Andjić et al., 2025). On the other hand, RPDs may be used to constrain the vertical axis rotation component of deformed units within orogens relative to the stable adjacent plates or continents. Examples of the latter include the analysis of folded terranes in accretionary orogens forming oroclines, such as in Alaska (Johnston, 2001), or Paleozoic Iberia (Pastor-Galán et al., 2015) and Australia (Cawood et al., 2011), or thrust sheet rotations such as in the Sevier-Laramide orogen of the USA (Yonkee and Weil, 2015), or back-arc extension-related forearc rotation such as in the Aegean region (Kissel and Laj, 1988; van Hinsbergen et al., 2010), or Japan (Otofuji and Matsuda, 1983), or forearc sliver motion along the Caribbean plate margin (Montheil et al., 2023). As illustrated above, paleomagnetic analysis of rocks knows many pitfalls and possible sources of bias and uncertainty, but once these are carefully taken into account and uncertainties are propagated into the final answer, it provides an invaluable quantitative source of information for interpretation of (plate) tectonic reconstruction and interpretation.

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Table 1: Types of errors affecting paleomagnetic poles.

	Error name	Type	Effect	Affected rocks	Physical basis	Mitigation
1	Rock magnetic recording	Directional, zero-mean	Increases directional scatter at specimen level	All	Grain-scale magnetostatic interactions; finite number of recording grains	Take multiple specimens per cooling unit or sedimentary horizon
2	Analytical error	Directional, zero-mean	Increases directional scatter at specimen level	All	Misorientations during core extraction and measurement; instrument noise	Take multiple specimens per cooling unit or sedimentary horizon
3	Undersampling of PSV	Directional, zero-mean	Increases directional scatter at paleomagnetic pole level	All, especially volcanic rocks	Random variations in the morphology of core-generated magnetic field	Sample and average multiple cooling units (Fig. 2)
4	Inclination shallowing	Directional, bias	Shifts paleolatitude towards lower value	Most sedimentary rocks	Compaction resulting in magnetization with shallower direction than geomagnetic field	With enough samples, apply statistical correction, such as with E/I method
5	Unrecognized tectonic rotation	Directional, bias	Typically shift the paleopole along an arc of constant radius from sampling location	Rocks in tectonically active regions	Deformation of local block rotates magnetization	Prefer tectonically simple blocks and/or be cognizant of possible block rotations
6	Geochronological error	Geochronological, zero-mean	Shifts nominal pole along APW path	All	Analytical error in radiometric dating or biostratigraphic range	Prefer units with high-precision ages
7	Remagnetization	Geochronological, bias	Relocation of paleopole to different part of APW path	Metamorphosed or altered rocks	Overprinting magnetization of younger age replaces original through heating or alteration	Apply field tests and rock magnetic analysis to exclude possibility of remagnetization

Table 2: Summary of common paleomagnetic field tests and their properties.

Test name	Pass condition	Age constraint on magnetization	Additional signs of confidence	Possible pitfalls
Conglomerate	Magnetizations of clasts have random directions	Unremagnetized since deposition	Clasts show internally consistent directions	Clasts are more resistant to remagnetization than rocks of interest
Fold	Magnetization aligns after unfolding	Older than age of folding	Large angle fold increase confidence	Very low angle folds may not distinguish magnetization age
Baked contact	Rocks near intrusion carry later magnetization	Older than age of intrusion	Overprint max. unblocking T. decays away from contact	Overprint may have different effect on intrusion and country rock
Reversal	Strata-bound magnetization show dual polarity	Unremagnetized since formation	Strata share common lithology	Non-synchronous overprinting events impart dual polarities