Magnetic Reversal Stratigraphy in the Ebro Basin, near Horta de Sant Joan, Spain

Nicholas Swanson-Hysell
Senior Integrative Exercise
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Advisors:
Cameron Davidson, Carleton College
David Barbeau, University of South Carolina
David Bice, Pennsylvania State University
Joseph Kirschvink, California Institute of Technology

Abstract

The Upper Cornudella and Montsant formations were deposited as proximal foredeep and wedge-top sediments in the Ebro foreland basin. The paucity of fossils in these terrestrial sediments requires the use of magnetostratigraphy to determine the timing of deposition and to establish sedimentation rates. Siltstone to fine sandstone beds and lenses from within fluvial and alluvial-fan deposits were sampled from 950 meters of section for paleomagnetic analysis. Correlation of the resulting 640 m composite local magnetic polarity stratigraphy (LMPS) with the geomagnetic polarity time scale (GPTS) indicates that this section spans from 30.4 to 26.3 Ma (early to late Oligocene). This implies a sedimentation rate of ~0.11 m/ky for the fluvial facies and increasing rates from 0.13 m/ky to 0.20 m/ky through the alluvial-fan facies. This does not date the onset of tectonic activity, although it does provide an upper limit for its onset. Rather, it provides dates (28.0 to 26.5 Ma) for when deposition was taking place in the region above active thrusts—the wedge-top depozone. A possible climatic boundary in paleosols, interpreted to have been deposited in the forebulge depozone, exists 380 m below the composite LMPS. The dates for the deposition of the proximal foredeep, and inferred sedimentation rates for the distal foredeep, suggest that this boundary is the Eocene-Oligocene climate transition. This dates the forebulge, which can be attributed to either the Pyrenees or the Catalan Coastal Range.

Keywords: magnetostratigraphy, Ebro Basin, Catalan Coastal Ranges, foreland basins, Oligocene, alluvial fans
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Introduction

Quantitative analysis of basin-fill that seeks to put constraints on depositional history requires age control. This is particularly important for foreland basin-fill as the dating of deposition and determination of sedimentation rates can provide valuable information useful for the evaluation of tectonic models and for determining the timing of deformational events. For example, with such geochronology the rate of flexural wave migration, and thus the rate of thrust belt propagation, can be determined (DeCelles and DeCelles, 2001).

Terrestrial sediments, such as those deposited in the southeast part of the Ebro Basin, are fossil-poor, severely limiting the use of biostratigraphy as a dating tool. Also lacking in the Ebro Basin are volcanic ash deposits that could be used to establish an absolute age through radiometric dating. Dating of magnetic reversals through a stratigraphic section can resolve this problem and can provide a chronologic framework. Furthermore, magnetostratigraphic constraints can improve the accuracy of basin-wide lithostratigraphic correlations.

Foreland basin systems can be thought of having four discrete depozones. From proximal to distal these depozones are: wedge-top, foredeep, forebulge and back-bulge (Fig. 1; (DeCelles and Giles, 1996). These depozones become superimposed on each other due to wedge propagation and flexural wave migration. Wedgetop, foredeep and backbulge facies have been identified for a section near Horta de Sant Joan in the southwestern part of the Catalán Coastal Range (CCR). While the expected depozones associated with the migration of a foreland basin system are found in the Horta section,
Figure 1: Schematic diagram of a foreland basin system modified from (DeCelles and Giles, 1996). In this system loading of the lithosphere due to shortening causes vertical deflection. The resulting flexure results in predictable depozones which are, from proximal to distal: wedge-top, foredeep, forebulge and back-bulge. All of the depozones except the backbulge are found in the Horta section. The magnetostratigraphy of this study is through the proximal foredeep and wedge-top depozones.
their proportions are unusual when compared to other foreland basin systems (Barbeau et al., 2005). The foredeep depozone is a much smaller percentage of the total Horta section than in other well-recognized foreland basins such as in the Andes and the Himalaya (DeCelles et al., 1998; DeCelles and Horton, 2003). This suggests that there is a complicated subsidence history where, in addition to the crustal loading of the Pyrenees, the CCR may have played a role.

This study uses paleomagnetic analysis of samples collected through the proximal foredeep and wedgetop parts of the section (the upper 640 meters) to construct a magnetic polarity stratigraphy. Samples were also collected and analyzed for some of the paleosols that are thought to have been deposited in the forebulge depozone. Much of the more distal foredeep section consists of very weathered poorly consolidated mudstone and was not sampled in this study.

This magnetostratigraphic dataset collected from these Paleogene fluvial and alluvial-fan deposits puts constraints on their depositional history; timing deposition and providing insight into sedimentation rates. It also provides tools with which to evaluate tectonic models for the basin and the larger Iberian-European system. Because samples collected for this project are from the proximal foredeep and wedgetop facies, they were deposited after the onset of flexure, and thus mountain building. Therefore, the dates provide an upper limit for the onset of such activity in the Ebro Basin. Also by dating the onset of synorogenic sedimentation the arrival of major thrust faults in the Horta area can be determined. Sediments that are part of this syntectonic growth strata must have been deposited on the front part of the orogenic wedge on top of active thrust faults. When
compiled with existing and future research, this study should refine the kinematic history of frontal structures surrounding the Ebro basin.

**Geologic Setting**

*Basin Formation*

The Ebro basin is a foreland basin that formed in response to crustal loading of the Pyrenees, Iberian Range and the Catalán Coastal Range (CCR) during the Paleogene collision of Iberia and Europe (Anadon et al., 1986; Lopez-Blanco, 2002). The development of the Ebro Basin is mainly attributed to the subsidence caused by the Pyrenean thrust belt (Anadon et al., 1986; Barbera et al., 2001; Zoetemeijer et al., 1990). Significant deformation and shortening in the Pyrenees began in the late Cretaceous (Simo, 1986), and continued until the early Oligocene (Burbank et al., 1992), or early Miocene (Arenas et al., 2001).

The Horta section is located in the southwestern part of the CCR near the linking zone that forms the transition between the CCR and the Iberian Range (Fig. 2). This part of the range is characterized by thin-skinned deformation with thrust sheets of Permian-Cretaceous carbonate (Guimerá, 1984). This contrasts to the more northeasternly part of the range where the deformation is thick-skinned, transpressive and dominated by en echelon left-lateral strike slip faults (Anadon et al., 1986; Roca et al., 1999).

*Stratigraphy*

The Paleogene Cornudella and Montsant Formations of the southern Ebro basin were deposited as soils, and as lacustrine, fluvial and alluvial sediments in a closed basin
Figure 2: Geographic setting of the study area—near the linking zone of the southern Ebro Basin. A) Map of the Iberian Peninsula showing the location of the Pyrenees, Iberian Range and Catalán Coastal Range. CCR- Catalán Coastal Range, LZ-Linking Zone. B) False color satellite image showing the location of Horta in the southwestern CCR. Location of field area (Fig. 3) is shown with box slightly to the east of Horta.
system (Fig. 3). The lower Cornudella formation consists of paleosols, lacustrine marls and fluvial overbank deposits. The upper Cornudella consists of channel sandstones, fluvial overbank deposits and conglomerate (Fig. 4). The Montsant formation is composed of alluvial-fan conglomerate with subordinate sandstone and siltstone (Fig. 5). The boundary between these formations is slightly ambiguous as there is not an unconformity separating the units, and since Spaniards tend not to use formation names when mapping. In the sequence of coarsening-upward fluvial and alluvial-fan strata, the formation boundary is placed where continuous sheets of conglomerate begin to dominate the section, and where the system transitions from an orogen parallel fluvial system to an alluvial-fan system.

The basin was completely enclosed by the surrounding mountain ranges in the late Eocene-Oligocene causing such terrestrial strata to be deposited throughout the basin (Coney et al., 1996). In the center of the basin, limestones and gypsum were deposited, while around the margin of the basin there was considerable clastic input, such as the Upper Cornudella and Montsant formations. The basin was internally drained until the middle Miocene (Turner et al., 1984).

Samples were collected from beds and discontinuous lenses of fine-grained sediment through 640 meters of the upper Cornudella and Montsant formations. These beds and lenses were deposited during over-bank flood events of fluvial sedimentation and during suprafan dispersion episodes of alluvial-fan sedimentation (Barbeau, 2003). These rocks are red, pink or tan, due to variable levels of pedogenesis.

Paleosols at the bottom of the section are interpreted to have been deposited on the forebulge (Barbeau et al., 2005). From stratigraphically high to low these paleosols
Figure 3: A) Simplified geologic map of the study area modified from (Olmedo and Domingo, 1984). The formations are described in the text. Locations are shown for the measured sections where samples were collected. PS-Paleosols CNL-Canaletes, TT-TT-330, RDM-Rodamunts, s5-s5 canyon, SB-Santa Bárbara, CG-Cedar Grove. B) Simplified stratigraphic column shows the vertical proportions of the formations and the stratigraphic location of the sections used for magnetostratigraphy in this study.
Figure 4: Upper Cornudella formation photos. A) This road-cut is part of the Canaletes (CNL) section through the Upper Cornudella. Fluvial mudstones are interbedded with fluvial conglomerates. B) Drill holes in a siltstone bed left from sample collection.
Figure 5: Photographs of the Horta section, Upper Cornudella and Montsant formations. A) Rodamunts growth structure; the Upper Cornudella is exposed to the right of the thick dashed line. The alluvial-fan conglomerates of the Montsant formation are to the left of the line. Strong evidence for syntectonic growth exist to the left of the thinner dashed line. The carbonate thrust sheets of the Catalán Coastal Range are visible in the upper right corner. The general locations of the Rodamunts (RDM) and s5 canyon (s5) magnetostratigraphy sections are shown. The s5 section continues through the mountain even though only the ends of the section are shown. B) Muntanya de Santa Bárbara; the top of the basin-fill exposed near Horta. The location of the cedar grove (CG) section through the Montsant formation is shown.
can be divided into three categories: vertisol, calcisol, and gypsisol. These paleosols are topped by a carbonate that transitions into fluvial clastics. The change from vertisol to calcisol likely represents a transition from a humid to an arid climate (Stalker, 2005). This is supported by geochemical evidence that shows that the parent material of the paleosols is similar and therefore climate is the factor that is responsible for the different soil types. This transition either represents the Paleocene-Eocene boundary, or the Eocene-Oligocene boundary (Stalker, 2005).

**Previous Magnetostratigraphic Work in the southeastern Ebro Basin**

Barberà et al. (2001) constructed a magnetobiostratigraphy from sections in the southeastern Ebro Basin. Using the magnetostratigraphy of alluvial and lacustrine sediments, and fossil mammal sites, they dated the intermediate to marginal facies of the Montsant alluvial system to have been deposited from 34.8 to 23.7 Ma (late Eocene to early Miocene).

Jones et al. (2004) interpret the entire foreland basin succession to have been deposited from ~40-37 Ma to ~27 Ma (late Eocene to late Oligocene) based on magnetostratigraphy from a composite section ~10 km from the Horta section. Jones et al. (2004) correlate their section lithostratigraphically with a rodent biozone determined for a site near the town of Gandesa. These faunal remains, belonging to the Eomys zitteli zone as defined by Barberà et al. (2001), are from ~28 Ma and should correlate with chron 9r of the Cande and Kent (1995) global polarity timescale. However, Jones et al. (2004) place this site at ~27 Ma and correlate to 9n/8r. Jones et al. (2004) also use, albeit with little confidence, unpublished biostratigraphy based on a charophyte (a type of green
algae) fossil assemblage found in the upper beds of the Cornudella lacustrine-fluvial system. This gives them a lower limit of early-middle Eocene for the bottom of their section. Jones (1997) places much more weight on the mammal fossil site than the Charophyte assemblage. This site is located 950 meters above where the conglomerate begins and will be used to help establish the position of the local magnetic polarity stratigraphy in this study.

Agustí et al. (2001) use Barberà et al.’s (1994) magnetostratigraphy of alluvial terrigenous strata and lacustrine strata from the Torrente de Cinca Section (~50 km from Horta) as the anchor for the Oligocene/Miocene boundary for their mammal scale for the Neogene of Western Europe. These sediments are a more distal facies and are not directly correletable to the Horta section.

Methods

Field Procedures

Samples for magnetostratigraphy were collected along six sections that can be easily correlated into one composite section that encompasses ~640 m of strata (Fig. 3). The Canaletes (CNL) section is a road-cut on TT-330 near the bridge over the Canaletes River (Fig. 3). It is 68 meters thick and was sampled at 3.0-meter resolution. The TT-330 (TT) section is another road-cut along the same road. It is to the west from the CNL section closer to Horta de Sant Joan, was 112 meters thick and was sampled at 8.0 meter resolution. The bottom of the TT section is a marker bed of fluvial sandstones that is the top of the CNL section.
The Rodamunts (RDM) section is along a dirt road that heads up the southeast side of the Rodamunts mountain from TT-330. It is 112 meters thick and was sampled at 10.1-meter resolution. Its base correlates to the 42-meter level of the TT section resulting in significant overlap. The s5 canyon (s5) section is in a canyon that runs through the Rodamunts growth structure. It is 437 meters thick and was sampled at 6.1-meter resolution. The s5 section has 35 meters of overlap with the top of the RDM section. The Santa Bárbara (SB) section and the Cedar Grove (CG) section are located on the southeast face of muntanya de Santa Bárbara (Fig. 5). The SB section is 73 meters thick and was sampled at 5.6-meter resolution. It overlaps completely with the top of the s5 section and the bottom of the CG section. The CG section is 159 meters thick and was sampled at 7.9-meter resolution. Its top marks the top of the composite section and is within 100 meters of the highest basin-fill exposed in the Horta area.

In total 458 oriented cylindrical cores of ~2.5 cm in diameter were collected from 154 sites using a Pomeroy gas-powered drill. Cores were taken from siltstone and sandstone beds and lenses. Generally three cores were taken per site (Fig. 4). Site spacing was limited by lithology, as a conglomerate of carbonate cobbles dominates the section. Oriented blocks of poorly consolidated mudstone were taken in the TT section and blocks of paleosols and marl were collected in the lower section for preliminary study (Fig. 3). In the laboratory, cores were taken from oriented block samples with a drill press equipped with a Pomeroy drill bit.
Laboratory Procedures

Measurements were made at the J.L. Kirschvick Paleomagnetism laboratory at the California Institute of Technology during December, 2004. The lab is equipped with two 2G Enterprises DC SQuID superconducting magnetometers. Both magnetometers are housed in magnetically shielded rooms. The procedures for the paleomagnetic measurements included the following:

1. Measurement of Natural Remnant Magnetism (NRM).
2. Low-temperature cleaning (77 K) in a zero field for some samples. This low-temperature step causes passage through the Verwey (110 K) and Morin (258 K) transitions preferentially demagnetizing multidomain magnetite and hematite. This step was done for sample sets CG and CNL.
3. Alternating field (AF) demagnetization using 25 G incremental steps up to 100 G.
4. Thermal demagnetization using 50°C -10°C steps up to 680°C. Above 200°C nitrogen gas was blown through the furnace creating a 99% nitrogen atmosphere in order to prevent oxidation of magnetic minerals.

For the CG, RDM and CNL sections 2-4 cores were analyzed for each site. One sample per site was analyzed for the s5 section. The TT, SB and paleosol samples were not fully analyzed in the laboratory due to time constraints. The results for those samples are not discussed in this paper because the thermal demagnetization steps for these samples only reached 500°C and thus they cannot be used for principal component analysis. In addition to the paleomagnetic measurements, isothermal remnant
magnetization (IRM), anhysteretic remnant magnetization (ARM) and rotational (RRM) rock magnetic experiments were performed on a sample from the s5 section.

**Analytical Techniques**

Craig Jones’ PaleoMag software was used for principal component analysis. The directions of the characteristic remnant magnetism (ChRM) and lower stability components were determined using least squares analysis after visual inspection of Zijderveld plots (Kirschvink, 1980; Zijderveld, 1967). The declination and inclination of the ChRM directions were used to calculate the latitude of the virtual geomagnetic pole (VGP). These VGP latitudes were the basis for construction of the local magnetic polarity stratigraphy (LMPS) with positive VGP latitudes signifying normal polarity zones and negative VGP latitudes signifying reversed polarity zones. The established polarity zones in the studied sections were then correlated to the geomagnetic polarity time scale (GPTS) of Cande and Kent (1995).

The data were divided into three classes: class I, class II and class III. Class I data showed a clear line or trend to the origin, was stable to high temperatures, and had easily interpretable polarity (Fig. 6). Class II data had a more ambiguous trend towards the origin and more questionable polarity (Fig. 7). Class III samples became unstable before a ChRM could be fit to the data (Fig. 7). VGP latitudes are calculated and displayed in the LMPS figures for both Class I and II data. Only the Class I points are used to develop polarity zones for the LMPS.
Figure 6: Zijderveld (vector component) demagnetization plots of representative specimens for which the characteristic remnant magnetism (ChRM) could be conclusively determined (Type I data). These data are tilt-corrected for bedding. Open squares are inclinations in a vertical plane, while filled squares are declinations in a horizontal plane. When not corrected for bedding, the slight movement during the AF steps and the large movement from AF 10 mT to the first thermal step corresponds with a present local field overprint. The vectors that can be fit from mid-to-high temperature points through the origin are the ChRMs.
Figure 7: Zijderveld (vector component) demagnetization plots for representative specimens for which ChRM could not be conclusively determined. These data are tilt-corrected for bedding. Open squares are inclinations in a vertical plane while filled squares are declinations in a horizontal plane. A, B, C are class II data points while D, E are class III. Class II data plots have more ambiguous trend towards the origin and more questionable polarity than class I plots. Class III plots are ones where the sample became unstable before a ChRM could be fit to the data.
Results

Components of Remnant Magnetism

The intensity of the natural remnant magnetization (NRM) was weak, generally between $5 \times 10^{-5}$ and $1 \times 10^{-6}$ emu/cm$^3$ in all sections. Alternating field (AF) demagnetization and low level thermal demagnetization removed a viscous present local field (PLF) overprint (Fig. 8). As can be seen in Figure 8, this PLF overprint clusters tighter and corresponds better with a normal magnetic field for the area before corrections are made for bedding tilt. This indicates that the overprint formed after the rocks were at their present orientation and thus after tectonic activity. There is strong sedimentary evidence for growth and progressive unconformities in these deposits, indicating that they are syntectonic (Fig. 5; (Anadon et al., 1986; Barbeau, 2003; Lopez-Blanco, 2002). Thus, any component of the magnetization that is not primary should cluster better before corrections are made for bedding tilt. This overprint must have formed after tectonic activity during a time of relative quiescence.

The movement resulting from AF demagnetization steps was usually slight, but almost always unidirectional. The field used for the AF steps was from 0-10 Gauss (0-1 mT) and was likely removing a low coercivity overprint acquired during drilling and transport. Low temperature steps also removed a PLF overprint. As can be seen in some of the example Zijderveld diagrams of the stepwise demagnetization, the movement due to the low-temperature steps could be quite large (Fig. 5; Fig. 6). This low-stability component represents a goethite mineralogy that is responsible for the overprint. Goethite has a high coercivity to AF demagnetization, but a low maximum unblocking temperature (~80-120 °C; (McElhinny and McFadden, 2000). As such, the goethite
Figure 8: Equal-area stereoplots of present local field (PLF) overprints for the s5 section. Solid circles are in the lower hemisphere while open circles are in the upper hemisphere. A, B) Vectors that were removed due to alternating field (AF) demagnetization steps between 0-10 mT. C, D) Vectors that were removed during thermal demagnetization steps from 0º-100º/150ºC. The mean vectors for the plots with no bedding-tilt correction correspond well with the PLF direction (358.2, 56; NSSD, 2004). The data becomes less clustered when a bedding correction is applied. Although the difference is slight, owing to the lack of variation in bedding tilt, it indicates that the AF and low-temperature components are a post-tectonic overprint.
component of the magnetization was largely unaffected by the AF steps, but is 
responsible for movement in the low temperature steps that removed a PLF overprint.

Some specimens became directionally unstable above low or mid-range 
temperature, but many exhibited a high-coercivity, high-temperature stable remanance 
that was removed above 600° C. This component, interpreted as the characteristic 
remnant magnetism (ChRM), decayed towards the origin and least-square fit lines were 
forced through the origin to fit the data (Fig. 6). The resultant ChRM vectors showed 
both normal and reversed polarities.

*Magnetic Stratigraphy*

Local magnetic polarity scales (LMPS) were constructed for three of the sections, 
Rodamunts (RDM), s5 canyon (s5), and Cedar Grove (CG), based on the virtual 
geomagnetic pole (VGP) latitudes calculated from bedding-tilt corrected ChRMs. The 
112-meter thick Rodamunts (RDM) section contains four reversals (Fig. 9). The polarity 
zones are supported by either one or two data points. The polarity zones in the upper third 
of the section are supported through correlation to the s5 section. There are 12 reversals 
through the 437 meters thick s5 section (Fig. 10). Several of the polarity zones are 
supported by data from only one site and the resulting ambiguity due to limited data is 
indicated on the LMPS. The large reversed polarity zone from 30 m to 135 m has poor 
resolution between 33 m and 132 m. The158 meter thick Cedar Grove (CG) section 
contains one reversal (Fig. 9). Either side of this reversal shows unambiguous polarity 
supported by many Class I data points.
Figure 9: Local magnetic polarity stratigraphy (LMPS) for the RDM and CG sections. Virtual geomagnetic pole (VGP) latitudes calculated from ChRM directions are used to develop the normal and reversed polarity zones. Negative VGP latitudes correspond with reversed polarity zones (white), while positive VGP latitudes correspond with normal polarity zones (black). While both Class I and II data are presented only the Class I points were used to develop the LMPS.
Figure 10: Local magnetic polarity stratigraphy for the s5 section. Virtual geomagnetic pole (VGP) latitudes calculated from ChRM directions are used to develop the normal and reversed polarity zones. Negative VGP latitudes correspond with reversed polarity zones (white), while positive VGP latitudes correspond with normal polarity zones (black). While both Class I and II data are presented only the Class I points were used to develop the LMPS. Polarity zones resulting from a single data point are marked with a question mark to indicate that they have minimal data supporting them.
Two sections, the highway TT-130 (TT) section and Santa Barbera (SB) section, were only demagnetized to 500° C. Their ChRMs could not be determined. The Canaletes (CNL) section needs the TT section to be correlated to the composite section and will be analyzed to extend the composite LMPS in the future.

As shown in Figure 3, these sections are geographically very close, and can be correlated by using packages of tabular conglomeratic beds that are laterally continuous and easily resolved on air-photos and in outcrop. When the 705 meters of section of the RDM, s5 and SB sections are correlated using marker beds they produce a 640 meter thick composite LMPS (Fig. 11). This LMPS has 15 reversals and the overlapping polarity zones, with the overlap resulting from the lithostratigraphic correlation, show good correlation to one another.

*Conglomerate Test*

A conglomerate test was performed on carbonate clasts at the 50-meter level of the Rodamunts (RDM) section (Fig. 12). Cores were collected from seven individual clasts. Reliable ChRM's were found for six of these cores. If these ChRM's have been stable since before the deposition of the conglomerate they should be randomly distributed. The test for uniform randomness of Watson (1956) requires that the length of the resultant vector (R) of the individual unit vectors in a given sample set of size N be smaller than critical value ($R_o$) that Watson calculated for various sample sizes. If $R < R_o$, there is 95% probability that these individual vectors constitute a random population. The ChRM's from the conglomerate clasts pass the Watson test for uniform randomness with
Figure 11: Composite local magnetic polarity stratigraphy (LMPS) constructed using lithostratigraphic correlations of the individual LMPS sections. The columns accurately represent stratigraphic thickness—they all have the same vertical scale. The magnetic polarity zones correlate where the sections overlap. This results in a composite LMPS with 15 reversals. For reference each polarity zone is named with a letter and a + or - indicating polarity.
Figure 12: Equal-area plot of ChRM directions from the conglomerate test. Filled squares are in the lower hemisphere while open squares are in the upper hemisphere. All directions are from carbonate cobbles sampled at the 50 meter level of the RDM section. These vectors constitute a random population according to the Watson test for uniform randomness (Watson, 1956). R-resultant vector length; R₀-critical length for given n.
N=6, R_o=3.85 and R=3.17, so that R< R_o. This is very strong evidence that the ChRM directions are the primary natural remnant magnetism (NRM).

Rock Magnetism

Rock magnetism data for a sample from the 32-meter level of the s5 section are displayed in Figure 13. The graph of anhysteretic remnant magnetization (ARM) acquisition never becomes asymptotic indicating that the sample is likely composed of hematite that did not get saturated. The Lowrie-Fuller test shows overlap between the AF demagnetization curves of the saturation isothermal remnant magnetization (sIRM) and the ARM. This suggests that there is likely a mixture of single-domain (SD), pseudo-single-domain (PSD), and multidomain (MD) grains. The IRM acquisition and demagnetization curves (Fig. 13) show that the signal is not dominated by MD magnetite.

Discussion

Magnetostratigraphy through 625 meters of the Upper Cornudella and Montsant formations reveals 15 reversals. Overlap in the polarity zones confirms lithostratigraphic correlations and reinforces the robustness of the local magnetic polarity stratigraphy (LMPS).

Correlation with the Global Polarity Time Scale

No fossils were found in the section limiting our ability to develop biostratigraphic constraints on age. Nevertheless, the E. zitelli mammal fossil locality used by Jones et al. (2004) and dated by Barberà et al. (2001) to be ~28 Mya should
Figure 13: Rock magnetism data for a sample from the 32-meter level of the s5 section. A) Anhysteretic remnant magnetization (ARM) acquisition curve. The upper dashed line is a magnetic bacteria standard that is entirely single-domain (SD) magnetite. The lower dashed line is a standard of highly interacting SD and pseudo-single-domain grains (PSD). Since the ARM for sample from this study never becomes asymptotic it is likely composed of hematite that didn’t get saturated. B) Lowrie-Fuller test. For SD grains the ARM curve lies above the corresponding saturation isothermal remanent magnetization (sIRM) curve while for MD grains the ARM curve lies below the sIRM curve (Lowrie and Fuller, 1971; Johnson et al., 1975). In this case, since the AF demagnetization curves of the sIRM and the ARM overlap one another, there is likely a mixture of SD, PSD and MD grains. C) IRM acquisition and demagnetization curves. The solid line is the step-wise acquisition of isothermal remanence while the dashed line is the AF demagnetization of the sample after the saturation of the IRM. The intersection of these two curves is the coercivity of remanence ($H_{cr}$) on the x-axis and the R (or cross-over) value on the y-axis. For SD grains the R value should be around 0.5 while it should be much less than 0.5 for strongly interacting MD grains (Cisowski, 1981). As the R value for the sample is ~0.5, the signal is not dominated by MD magnetite.
correlate to somewhere in the upper part of the Horta LMPS. This biostratigraphy, previous magnetostratigraphic work in the region (Barbera et al., 1994; Jones et al., 2004), and the postulated presence of the Paleocene-Eocene or Eocene-Oligocene boundary in the paleosols (Stalker, 2005), place the section somewhere in the Eocene or Oligocene.

The proposed correlation of the composite Horta LMPS to the global polarity time scale (GPTS) of Cande and Kent (1995) is shown in Figure 14. This correlation relates the reversal pattern seen in the Horta section to chron C11-C8 placing it in the early to late Oligocene. For this correlation to work, two reversed polarity zones, K- and M- (Fig. 11), must be ignored. Both these polarity zones are based on a single site. Other than this discrepancy, the fit between the LMPS reversal pattern and the GPTS is quite straightforward and unambiguous. As such, the correlation is robust even without solid independent age control.

**Magnetic Mineralogy**

The ability to correlate the Horta LMPS with the GPTS of Cande and Kent (1995) argues for early acquisition of the ChRM by detrital hematite, and/or immediate post-depositional authigenic hematite that would have formed during pedogenic reddening. The development of a post-depositional pigment could not have taken place over a very long period of time because the sediments were deposited in an actively aggrading alluvial fan system (Anadon et al., 1986; Lopez-Blanco, 2002). Clast supported conglomerates of carbonate pebbles to boulders, which were deposited rapidly as sheetflood strata, comprise ~75% of the strata through the section (Barbeau, 2003). These
Figure 14: Correlation of the local magnetic polarity stratigraphy (LMPS) to the global polarity time scale (GPTS) of Cande and Kent (1995). A) Correlation to the Eocene-Oligocene GPTS. This fit is unique. For it to work two polarity zones based on one data point each, K- and M-, must be ignored (Fig. 12). B) Closeup of fit to chron C11-C8. LMPS ‘x’ has true stratigraphic thickness. This results in a constant sedimentation rate of 0.17 m/ky. LMPS ‘y’ has adjusted thicknesses to improve the fit with the GPTS resulting in a variable sedimentation rate. C) Sedimentation rate for the adjusted LMPS. These rates are averaged for each chunk of the LMPS that was stretched differentially. Note that the rates increase up section from 0.11 m/ky to 0.20 m/ky.
deposits are typically 1-6 m thick. With an average deposition rate of 0.17 m/k.y. one of these sheets would need to be deposited every 6,000-40,000 years. Thus periods of non-deposition and pedogenesis could not have been sustained for very long. After deposition of one of these conglomerate sheets, pedogenesis of the underlying sediment would stop, and the post-depositional processes that were forming pigmentary hematite would cease.

The possibility of a hematite dominated magnetic mineralogy is supported by the isothermal remnant magnetization (IRM) acquisition and demagnetization curves that show that the magnetic signal is not dominated by multidomain magnetite, and by the anhysteretic remnant magnetization (ARM) acquisition data that suggests that hematite is the dominant magnetic mineralogy (Fig. 13). It is clear from the Lowrie-Fuller test that there is a mix of single-domain and multidomain magnetic grains.

*Sedimentation Rate*

This original correlation to the GPTS assumed a constant sedimentation rate for the section and thus was stratigraphically proportional—no parts of the LMPS were stretched. The average sedimentation rate from the correlation of the stratigraphically proportional LMPS to the GPTS is 0.17 m/k.y. before decompaction. Once this fit was established, the LMPS was adjusted to fit more accurately to the GPTS (Fig. 14). While the chron of the original LMPS aligned fairly well with the GPTS some of the tie-lines between chron were not straight in order to make the correlation work. The adjusted fit improves the correlation by dividing the LMPS into 4 sections and stretching them differentially. This results in a variable sedimentation rate with rates increasing from 0.11 m/ky in the lower part of the section to 0.20 m/ky in the upper part (Fig. 14).
The transition from the Upper Cornudella formation to the Montsant formation occurs between meter level 80 and 180 of the composite LMPS. The adjusted LMPS shows a marked increase in sedimentation rate around meter level 110. This increase in sedimentation rate corresponds with the change from the orogen parallel fluvial system of the Upper Cornudella to the alluvial fan deposition of the Montsant. Using 110 m as the formation boundary an average rate can be found for the fluvial system, 0.11 m/ky, and the alluvial system, 0.16 m/ky.

However, the adjusted LMPS indicates that there is a continued increase in sedimentation rate through the Montsant (Fig. 14). The sedimentation rate can be better modeled with a second order polynomial, \( y = 1 \times 10^{-0.5} \times x^2 + 0.1315x \), than with a linear fit (Fig. 15). This polynomial fit shows an increase in rates from 0.13 m/ky at the bottom of the Montsant formation to 0.20 m/ky at the top of the section. This increase in rate corresponds with the sedimentary evidence that shows the alluvial-fan deposits coarsening and becoming more proximal. Solid evidence of syntectonic deformation begins in the section at 329 m (Barbeau, 2003). This provides further evidence that the deposits are becoming more proximal as the syntectonic sediments were deposited above the active thrusts. Rates of accumulation in foreland basin systems generally increase as the fold-thrust belt approaches (Ojha et al., 2000).

Modern alluvial fan aggradation rates have been measured in the Owens Valley of California to be between 0.08 m/ky and 0.15 m/ky (Beaty, 1970). Accumulation rates found for fluvial sediments in the Himalayan foreland basin range from 0.32 m/ky to 0.47 m/ky (Ojha et al., 2000). These rates are similar in magnitude to the ones calculated here.
Figure 15: Sedimentation rate based on correlation for adjusted local magnetic polarity stratigraphy. The fit between the first points is linear as it connects the two endpoints for the period of fluvial deposition. The slope of this fit gives an average sedimentation rate of 0.11 m/ky from 0 m to 105 m. The dashed line is the linear best-fit to the data from 105 m to 640 m. This results in an average depositional rate of 0.16 m/ky for alluvial-fan deposition. The alluvial deposition can be fit much better with a second order polynomial. The polynomial shown, \( y=0.00001x^2+0.1315x \), is simplified in that it uses relative time as opposed to absolute time. It takes the time of the first point in the fit, 29.4 Ma, and uses that as time 0 with the next point, whose time is 27.9 Ma, being 1,500,000. The derivative of this fit, \( dx/dy=0.00002x+0.1315 \), can be used to calculate instantaneous depositional rates for any point on the curve, again using relative time. Example instantaneous rates from the bottom, middle and top (0.13, 0.17 and 0.20 respectively) are shown on the graph.
for the Horta section through magnetostratigraphy, assuring that these interpreted rates are at least reasonable for fluvial and alluvial-fan deposition.

Deposition in an alluvial fan is episodic in nature. The architecture of the alluvial-fan deposits in the Horta section suggests that it formed by processes of fan entrenchment and trench migration (Barbeau, 2003). Where there was no trench deposition occurring, fine-grained sediment (the sediments drilled for paleomagnetic analysis in this study) could be deposited as overbank deposits during suprafan dispersion episodes (Barbeau, 2003). This means that much more time can be represented by a fine-grained bed and lense than by a bed of conglomerate. This leads to the question of whether rates are even meaningful in a system where deposition is so episodic and where there is lateral accommodation space.

Long-term rates are significant, however, particularly when averaged out over 1-4 million years as they are in this study. As discussed earlier, a conglomerate sheets of 1-6 m was likely deposited every 6,000-40,000 years with overbank deposits forming between these episodes. Given the number of depositional episodes encapsulated in a million years of strata the average depositional rates discussed here are long-term enough to be meaningful. While they have meaning, it is difficult to resolve the temporal variables that go into the depositional rate, climate for example, and the spatial variables, like trench migration (Paola, 2005). Thus, the previous discussion of relative change in sedimentation rate has much more practical implications than a discussion of the actual rates.
Timing of Basin Development

The dates found through this magnetostratigraphy suggest that the paleosols of the Lower Cornudella (Fig. 3) contain the Eocene-Oligocene boundary. The possible climatic transition in the paleosols occurs ~380 m below the base of the composite LMPS. Using the Eocene-Oligocene boundary and the date found for the bottom of the LMPS, this 380 m of sediment had to be deposited between 33.7 Ma (the Eocene-Oligocene boundary; (Cande and Kent, 1992) and 30.4 Ma. This would require an average sedimentation rate of ~0.12 m/ky. This sedimentation rate is very similar to that determined for the Upper Cornudella and, as such, is a feasible rate if depositional rates were constant through the foredeep deposits. This extrapolation, if correct, gives a general date for the forebulge succession, and perhaps the arrival of the Pyrenean foreland basin. The paleosol succession can also be modeled as being deposited on the forebulge of the CCR (Barbeau et al., 2005) and could be dating the arrival of that flexural wave.

Solid evidence of growth, syntectonic deposition, occurs 329 meters into the composite LMPS (Barbeau, 2003). Based on the correlation of the adjusted LMPS to the GPTS this growth starts at 28.0 Ma. This transition to synkinematic deposition also marks the boundary between the foredeep and wedge-top deposits, as wedge-top deposits are characterized by growth structures (DeCelles and Giles, 1996). The wedge-top facies is part of the orogenic wedge and while it was being deposited the wedge was actively deforming (Fig. 1). Therefore, the thrust faults of the Catalán Coastal Range, near Horta de Sant Joan, were active from 28.0 Ma to at least 26.3 Ma.
Conclusions

Paleomagnetic sampling and analysis of 640 m of terrestrial foreland basin-fill in the southeast Ebro Basin, near Horta de Sant Joan, reveals 15 reversals. Correlation of the resulting local magnetic polarity stratigraphy (LMPS) to the global polarity time scale (GPTS) dates deposition from 30.4 to 26.3 Ma (early to late Oligocene). From this fit, an average sedimentation rate of ~0.16 m/ky was calculated. Changing the proportions of the local magnetic polarity zones to tighten the fit to the GPTS indicates that sedimentation rates increase up section. The positioning of the LMPS suggests that a climatic shift recorded in paleosols lower in the section is the Eocene-Oligocene boundary. The onset of synkinematic deposition in the section is dated at 28.0 Ma.

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