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Paleogene events, evolution and Stratigraphy

Simonetta Monechi[1], Noël Vandenberghe[2] & Laia Alegret[3]

[1] Dipartimento di Scienze della Terra, University of Florence, Via La Pira 4, I-50121 Firenze, Italy
Simonetta.monechi@unifi.it

[2] Geology, Celestijnenlaan 200E, B-3001 Hverlee-Leuven, Belgium
Noel.vandenberghe@ees.kuleuven.be

[3] Departamento de Ciencias de la Tierra & Instituto Universitário de Cencias ambientales,Universidad de Zaragoza, E-50009 Zaragoza, Spain. laia@unizar.es

Summary

The warming trend that started in the latest Paleocene and culminated in the Early Eocene Climatic Optimum (EECO) between 52 and 50 Ma is punctuated by short-lived hyperthermals, like the Paleocene Eocene Thermal Maximum (PETM) that marks the base of the Eocene. Such short-lived hyperthermals are linked to global carbon cycle perturbations and associated with major floral and faunal changes. Orbital pacing is probably triggering these hyperthermal perturbations. After the EECO a cooling trend sets in only interrupted by a short Middle Eocene Climatic Optimum (MECO) in the beginning of the Bartonian; at the beginning of the Oligocene the cooling trend is leading to the start of the present ice-house state with the growth of the first large Antarctic ice sheets.

The Paleogene timescale is based on the seafloor magnetic field reversals in combination with radio-isotopically or astronomically dated control points. The GTS2012 used astronomically derived ages in the intervals 66-53 and 37-23Ma and a radio-isotopical age model in the interval between 53 and 37 Ma. It is expected that it will become possible in the future to extend the astrochronological scale to the entire Paleogene.

The stage GSSP's of Danian (El Kef, Tunesia), Selandian and Thanetian (Itzurun Beach, Zumaia, Basque County, Spain), Ypresian (Dababiya section, Luxor, Egypt) , Lutetian (Gorrondatxe cliff, Basque County, Spain) and Rupelian (Massignano section, Ancona, Italy) have already been defined. For defining the Priabonian and the Chattian GSSP's, suitable sections are to be discussed respectively the Alano di Piave section in the Venetian southern Alps and the Monte Cagnero section in the northeastern Apennines of Umbria-Marche in Italy. The Bartonian, historically referring to marine clays and sands in coastal sections of the Hampshire basin in England, is still lacking a suitable GSSP section; *in* GTS2004 and 2012 its base is provisionally kept at the C19n/C18r magnetic reversal. Potential Bartonian GSSP sections will be discussed.

Keywords: Paleogene Stratigraphy, Evolution, Paleoclimate, GSSP, Chronostratigraphy, Hyperthermals, Cyclostratigpraphy

The Paleogene is a period of important revolution in the climatic and biotic history of the Earth, characterized by elevated concentration of greenhouse gases and warm temperature conditions. After an initial slight cooling during the early Paleocene, a gradual warming trend started at ~57 Millions of years ago (Ma), culminating at the Early Eocene Climatic Optimum (EECO), a broad interval recording the highest temperature of the entire Cenozoic between 52-50 Ma (Zachos *et al*., 2001). Following the EECO, a long-term cooling eventually led the climate system into the ice-house state with the formation of the first large-scale Antarctic ice-sheet at the Eocene/Oligocene boundary (~34 Ma). It is now well documented that a series of short-lived (<200 kyr) events of extreme warming, known as hyperthermals, superimposed on the pre-EECO long-term warming trend (Thomas & Zachos, *2000*; Quillévéré *et al*., 2008; Westerhold *et al*., 2008). These transient events are closely linked to perturbations of the global carbon cycle as evidenced by concomitant large negative Carbon Isotope Excursions (CIEs), which have been related to the input of isotopically light carbon into the coupled ocean-atmosphere system. The best known and largest hyperthermal, the Paleocene-Eocene Thermal Maximum (PETM or Eocene Thermal Maximum 1 –ETM1), marks the end of the Paleocene at ~56 Ma, and is followed by at least two other similar events of decreasing magnitude, the ETM2 at ~54 Ma and the ETM3 (52.5 Ma) (Zachos *et al*., 2004, 2005; Stap *et al*., 2010a, b). These warming events are associated with major floral and faunal changes, large perturbations in the carbon cycle, ocean acidification as inferred from dissolution of deep-sea carbonates, and perturbations of the hydrological cycle that are consistent with the massive injection of isotopically depleted carbon into the atmosphere-ocean system (Kennett & Stott, 1991; Zachos *et al*., 2008, 2010; Westerhold *et al*., 2008).

Cyclostratigraphic evidence suggests that orbital pacing controlled the timing of major fluctuations in the carbon-climate system and consequently of the occurrence of hyperthermals (Lunt *et al*., 2011). Similar events have now been found in the early Danian, although showing smaller amplitude than the PETM. Being linked to the rapid input of a massive amount of carbon into the exogenic pool, these events potentially represent past analogues of future climate conditions. Their study, therefore, could provide important insights to predict the future responses of Earth’s climate and biota to anthropogenic emissions of CO_2.

The Paleogene was also a period marked by significant evolutionary turnover in both marine and terrestrial biota. High-resolution stratigraphic (biostratigraphy, stable isotope stratigraphy, magnetostratigraphy and environmental magnetism) and geochemical data reveal significant changes in the climate, paleoceanography and in the biota, pointing at different causes and consequences of the Paleogene events (Galeotti *et al*., 2010; Coccioni *et al*., 2012; McInerney & Wing, 2012 with references therein).

So far, in fact, the reversal history of the Earth’s magnetic field has been the backbone of the Cenozoic time scale (GPTS) ever since it was constructed. The GPTS is constructed using seafloor anomaly profiles in combination with radio-isotopically dated age control points. An alternative method consists of astronomically dating magnetic reversals and bio-events by tuning sedimentary cyclic sedimentary successions to astronomical target curves. A critical examination of the quality of the available radiometric data and of the published astronomical numerical ages, has led *in* GTS2012 to a combination of both approaches as neither of the two methodologies is better over the entire range of the Paleogene. Astronomically derived ages are used for the time intervals between 66 and 53 Ma and 37 and 23 Ma. The ages for the chron boundaries in between come from the radio-isotopic age model. This led to a hybrid Paleogene time scale GTS2012 which in the future will probably become a fully astronomically tuned time scale.

Spectral analyses of well-resolved Paleocene and Eocene oceanic and land-based sedimentary successions has lately highlighted the importance of orbital forcing as a triggering mechanism for hyperthermal events, on the other hand allowing to develop a floating cyclochronology for a large part of the Paleogene. Covering the cyclochronological gap between ~40 Ma and the early Eocene, and integration with radiometric dating will hopefully provide soon the conditions to extend the astrochronological scale to the entire Paleogene.

A brief history and the definitions of the subdivisions of the Paleogene period have been described in the GTS2012, volume 2. Here we report the state of art of the Paleogene stages.

The Danian stage GSSP coincides with the Cretaceous/Paleogene (K/Pg) boundary. It is defined at the base of the boundary clay in El Kef (Tunisia), which contains a high concentration of impact evidence that can be used for correlation purposes (Molina *et al*., 2006). The mass extinction of planktonic foraminifera and calcareous nannofossils and the biotic turnover of other marine groups are recorded at the base of the boundary clay, and are excellent tools for global correlation in marine successions. Several auxiliary sections exist (Molina *et al*., 2009). The age reported *in* GTS2012 for the K/Pg boundary is 66 Ma.

The Selandian stage GSSP is defined at the base of the red marls of the Itzurun Formation in the coastal cliff along the Itzurun Beach in Zumaia (Basque Country, Spain) (Schmitz *et al*., 2011). At the base of the Selandian, prevalently marly lithologies replace the succession of Danian red limestone and limestone-marl couplets. Such a lithological change recoded in the deep-water setting of Zumaia is time-equivalent with the transition from Danian limestones to Selandian detrital sediments in the shallow water setting of the type area in Denmark, which suggests a link with a major sea-level regression, the Se-1 sequence boundary of Hardenbol *et al*. (1998), dated at 61.6 Ma *in* GTS2012. The second radiation of fasciculiths (calcareous nannofossils) is considered as the best criterion for global correlation of the base of the Selandian in the marine realm (Bernaola *et al*., 2009; Monechi *et al*., 2013). The location of the base of the Selandian 30 precession cycles (~630 kyr) above the top of magnetochron C27n in Zumaia suggests further correlation potential by means of cyclostratigraphy and magnetostratigraphy (Dinarès-Turrell *et al*., 2010).

The Thanetian is the youngest stage of the Paleocene. The Thanetian stage GSSP is located in the same beach section at Zumaia as the Selandian GSSP and it coincides with the C26r/C26n reversal (Schmitz *et al*., 2011), dated 59.2 Ma *in* GTS2012. Dinoflagellate events are useful for regional and inter-regional correlation (Heilmann-Clausen, 2007). While the base of the Thanetian is not associated with any global biotic turnover, marked changes in calcareous nannofossils and planktonic and benthic foraminifera occur ~160 kyr before the GSSP level (age based on cyclochronology), during a time interval unfavorable for the preservation of $CaCO_3$ known as the Mid-Paleocene Biotic Event. The latter was probably linked to a warming event (Bernaola *et al*., 2007).

The base of the Eocene and of the Ypresian stage has been defined in the Dababiya section near Luxor in Egypt (Aubry *et al*., 2007). The corresponding GSSP coincides with the start of a prominent negative carbon isotope excursion (CIE), selected for recognition and correlation with marine and terrestrial deposits. The CIE is the expression of the Paleocene-Eocene Thermal Maximum (PETM), a global warming event that took about ~200 ka to return to background values. The CIE coincides with major paleontological events in the marine and terrestrial realms, including the most severe extinction of deep-sea benthic foraminifera of the last 90 Myr, a global acme of the dinoflagellate genus *Apectodinium* and the planktic foraminiferal genus *Acarinina* both migrating to high latitudes, the rapid evolutionary turnover of calcareous nannoplankton, the extinction and origination of shallow-water larger benthic foraminifera, latitudinal migration of plants and a rapid radiation of mammals on land. The GTS2012 age for the base of the Eocene and Ypresian is 56 Ma.

The Lutetian GSSP (Early/Middle Eocene boundary) is defined at the Gorrondatxe sea-cliff section in the Basque Country, northern Spain. An important sea-level drop around the Ypresian-Lutetian transition made the choice of a deep-water reference section necessary; the dark marly level now defining the GSSP corresponds to a global maximal flooding event. The GSSP level coincides with the lowest occurrence of the calcareous nannofossil *Blackites inflatus*, occurring at the base of the Subzone CP12b within Zone NP14, and was dated 47.8 Ma *in* GTS2012. A distinctive low-carbonate interval interrupts the continuous limestone-marl alternation of the Gorrondatxe section at the early Lutetian (Middle Eocene) C21r/C21n Chron transition, ~50 metres above the GSSP. This interval has been recognized in the Atlantic and Pacific Oceans, and it presents many of the hallmarks of Paleogene hyperthermal deposits (Payros *et al*., 2012)

No GSSP has been defined yet for the Bartonian. The Bartonian refers to the marine clays and sands of the Barton Beds of the Hampshire basin, which are exposed in coastal sections in central southern England. In the absence of a suitable section for defining the Bartonian GSSP, the C19n/C18r boundary is retained as the base of the Bartonian as *in* GTS2004. *In* GTS2012 the age of this magnetozone boundary is estimated to be 41.2 Ma. Calcareous microfossil events around this boundary can help correlation (Jovane *et al*., 2007, 2010). A significant turnover in planktonic foraminiferal assemblages is observed slightly above this boundary, across the so-called Middle Eocene Climatic Optimum (MECO).

No GSSP has been defined yet for the Priabonian, although the Alano di Piave section in the Venetian southern Alps has been informally proposed (Agnini *et al*., 2011). An easily recognizable volcanic crystal tuff bed in the section, correlatable with several calcareous microfossil levels could be chosen. However, in the absence of a formal Priabonian GSSP, the stage base is kept as in previous reviews at the base of C17n.1n. The age of this magnetochron boundary was determined at 37.8 Ma *in* GTS2012.

The Oligocene, historically with a threefold subdivision, has had an official twofold subdivision for a few decades; the choice of the Rupelian GSSP, however, honours the original meaning of the base of the Oligocene. This GSSP, defining the base of the Oligocene and the Rupelian stage, has been defined in open marine marls and calcareous marls in the Massignano section, southeast of Ancona on the Adriatic coast of Italy. The GSSP is denoted by the extinction of hantkeninids in planktonic foraminiferal assemblages. The base of the Oligocene and Rupelian is dated at 33.9 Ma *in* GTS2012.

No formal proposition has yet been made for a Chattian GSSP. However, suitable and well-studied sections exist in the Umbria-Marche northeastern Apennines in Italy where the Rupelian-Chattian transition is well documented by means of microfossils. As no formal proposition has been made at present, an older criterion for the definition of the boundary has been retained, namely the base of C10n.1n. near the level of the last common occurrence (LCO) of the planktic foraminifer *Chiloguembelina cubensis*. This magnetochron level is dated at 28.1 Ma *in* GTS2012. The Monte Cagnero section (pelagic Scaglia Cinerea Formation, Umbria–Marche region, central Italy) is a potential candidate for the Chattian GSSP (Coccioni *et al*., 2008). These authors suggested that the LCO of *C. cubensis* is a reliable bioevent that can be used to recognize the O4/O5 (P21a/P21b) zonal boundary.

The Paleogene-Neogene boundary coincides with the magnetic reversal from C6Cn.2r to C6Cn.2n in the section at Lemme-Carrosio in northern Italy and dated 23.03 Ma *in* GTS2012.

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