Timing and tempo of organic carbon burial in the Monterey Formation of the Santa Barbara Basin and relationships with Miocene climate: SUPPLEMENTARY MATERIAL

**Eliel S. C. Anttila1, Francis A. Macdonald1, Dawid Szymanowski2,3, Blair Schoene2, Andrew Kylander-Clark1, Clara Danhof4, David S. Jones4**

*1Department of Earth Science, University of California Santa Barbara, 1006 Webb Hall, Santa Barbara, CA, 93117*

*2Department of Geosciences, Princeton University, Guyot Hall, Princeton ,NJ, 08544*

*3Institute of Geochemistry and Petrology, ETH Zurich, Clausiusstrasse 25, 8092 Zurich, Switzerland*

*4Department of Geology, Amherst College, Beneski Building, Amherst, MA, 01002*

This document includes detailed methods for all geochronological and geochemical analyses, geochronological sample locations (Table SM1), LA-ICPMS (Table SM2), CA-ID-TIMS (Table SM3), TOC (Table SM4), and XRF (Table SM5) data tables, and visualizations of geochronological data used to generate U-Pb ages (Figs. SM1-SM4). Calculations of organic carbon mass accumulation rates for the Monterey Formation in the Belridge Field of the San Joaquin Basin, referenced in Section 5.4 of the main text, are outlined in Supplementary Section 2. Finally, a sensitivity test of the carbon isotope box model discussed in Section 5.5 of the main text is outlined in Supplementary Section 3, and summarized in Fig. SM5.

**All figure components, code, and data are compiled in the following GitHub repository:** https://github.com/eliel-anttila/Anttila\_Monterey\_2023

**1. METHODS**

***1.1 Field work***

Between 2018 and 2022, stratigraphic sections were measured at multiple Monterey Formation outcrop localities on the northern edge of the Santa Barbara Channel. While all localities have been studied by previous workers, we measured our own sections to develop a consistent stratigraphic height framework across all localities, and to have reference points which we could use to compare our datasets to earlier measured sections. Within each section, bentonite samples were collected for U-Pb geochronology, with ~0.5kg of sample collected from each horizon. The locations and stratigraphic heights of all collected samples are compiled in Table SM1.

At the El Capitan section, fist-sized shale samples were collected for TOC and XRF analysis at ~1m intervals in the lower 100m of the section, and ~30cm intervals in the upper portion of the section. TOC data from El Capitan is compiled in Table SM4, while XRF data are tabulated in Table SM5.

***1.2 Geochronology sample preparation***

*1.2.1 Mineral Separation*

Bentonite samples were combined with water and disaggregated with ~30s of pulsed blending in a standard kitchen blender. To remove mineral grains from the clay matrix and remove fine sediment, the resulting slurry was then poured into sonication/agitation equipment, following the methods of Hoke et al. (2014). 30 second pulses of sonication and constant circulation with magnetic stirrers were continued with constant water flow through a series of two beakers, until the water in both beakers ran completely clear. The remaining sediment that accumulated on the bottom of both beakers was panned, and a heavy mineral fraction separated and placed in a large petri dish. We found that lightly scoring the bottom surface of the petri dish (effectively creating shallow riffles) and swirling/running the heavy mineral separate over the scored zone resulted in the concentration of zircon in and around the scored markings. Using these methods, we were able to recover ~100s of zircon crystals from most ~.5kg bentonite samples.

*1.2.2 Picking, annealing, mounting, and imaging*

Zircon crystals were picked (typically 150-200 in samples with abundant zircon) from the mineral separate of each sample with tweezers, transferred to crucibles, and annealed at 900ºC for 48 hours in a muffle furnace. Annealed zircon were then mounted in epoxy and polished. Polished mounts were subsequently imaged with cathodoluminescence (CL) using a Centaurus CL detector mounted on an FEI Quanta 400f field emission SEM at the University of California Santa Barbara (UCSB).

***1.3 LA-ICPMS analyses***

Laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) U-Pb geochronological analyses on zircon were completed at UC Santa Barbara using a Cetac/Photon Machines Analyte Excite 193 nm excimer laser attached to a Nu Plasma 3D multicollector ICPMS, following the methods of Kylander-Clark et al. (2013). Each zircon was ablated with a 20µm laser spot. The zircon 91500 (Wiedenbeck et al., 1995) was used for age calibration. Secondary zircon reference materials included 9435, AUSZ, Mudtank, GJ1 (Jackson et al., 2004), and Plesovice (Sláma et al., 2008). *Iolite* (Paton et al., 2010) was used to correct for U-Pb mass bias and drift following the methods of Kylander-Clark et al. (2013) and Horstwood et al. (2016).

All LA-ICPMS data are compiled in Table SM2. 238U/206Pb ratios were corrected for 230Th disequilibrium using measured zircon Th/U and an assumed Th/U(magma) of 3.5. Dates for each grain were calculated by importing reduced 238U/206Pb and 207Pb/206Pb ratios into *IsoplotR* (Vermeesch 2018). For each sample, a group of zircons with young (Miocene) 206Pb/238U dates were identified. This group of analyses was used to calculate a weighted-mean age: the oldest analyses were iteratively removed until a group of zircon ages conforming to statistical standards for a single population (Wendt and Carl, 1991) was acquired. The data used to calculate these weighted mean ages are in bold text and highlighted in grey in Table SM2, while older analyses that did not go into the weighted mean (likely representing either magmatic inheritance or detritus in the bentonite bed) are italicized. A blanket systematic uncertainty of 2% (accounting for all uncertainty criteria outlined in Horstwood et al., 2016) was then applied to the calculated weighted-mean ages. The weighted-mean age, and uncertainty, is utilized in our chronostratigraphic framework as the eruptive/depositional age and uncertainty for the measured bentonite horizon. Samples with a single young grain are treated as detrital, with the 206Pb/238U age of the youngest grain serving as the maximum depositional age (MDA) for the sampled horizon.

***1.4 CA-ID-TIMS analyses***

*1.4.1 Chemical Abrasion*

Individual grains from the population of zircons that make up the LA-ICPMS weighted mean age for each sample were analyzed with single zircon U-Pb chemical abrasion isotope dilution thermal ionization mass spectrometry (CA-ID-TIMS) at Princeton University, following standard chemical abrasion methods modified from Mattinson (2005). Previously annealed single zircons were plucked from epoxy mounts, transferred to Teflon beakers, and rinsed with 3N HNO3. Grains were removed from the acid rinse and loaded into 200µl Savillex microcapsules with ca. 90 µl 29M HF. Microcapsules were assembled in a Parr bomb and zircons were leached at 180ºC for 12 hours to remove domains in the crystal lattice that may have experienced lead loss.

*1.4.2 Dissolution, spiking, and column chemistry*

Following leaching, zircon grains were moved to Teflon beakers, and repeatedly rinsed in 3N HNO3 and 6N HCl. The crystals were then transferred back to microcapsules, spiked with the EARTHTIME  205Pb-233U-235U tracer (Condon et al., 2015; McLean et al., 2015) and placed back into a Parr bomb for dissolution in ca. 90 µl 29M HF for 60 h at 210ºC. The resulting solutions were then dried down, converted to chlorides in the Parr bomb overnight, and dried down once more on the hot plate. The samples were then redissolved in 3N HCl and placed into 50 µl microcolumns filled with AG-1 X8 resin, where U-Pb and trace element solutions were separated by anion exchange following methods modified from Krogh (1973). The U-Pb solution was dried down in a Teflon beaker on the hot plate with a microdrop of 0.015M H3PO4. Each aliquot was then redissolved in a silica gel emitter (Gerstenberger and Haase, 1997), and loaded with an ultrafine pipette onto a single outgassed zoned-refined rhenium filament.

*1.4.3 Isotopic measurements*

Lead and U isotopic measurements were performed with one of two Isotopx Phoenix thermal ionization mass spectrometers (TIMS) at Princeton University. Pb isotopes were measured using peak-hopping mode on a Daly photomultiplier ion-counter, while U isotopes were measured as UO2 in static mode with Faraday cups connected to traditional 1012 Ω amplifiers, or to ATONA amplifiers (Szymanowski and Schoene, 2020). Instrumental mass fractionation for Pb was corrected with a factor (0.14 or 0.18 %/amu) derived from a long-term compilation of in-run 202Pb/205Pb values of double-Pb-spiked Daly analyses on each instrument. The dead time of the Daly systems was kept constant throughout the period of the study but was monitored with analyses of NIST SRM 982 Pb isotope standard over a range of intensities. U runs were corrected for fractionation using the known 233U/235U of the spike and assuming a sample 238U/235U of 137.818 ± 0.045 (2σ; Hiess et al., 2012). An 18O/16O value of 0.00205 ± 2 (1σ) was used to correct for interferences in UO2 analyses.

*1.4.4 Data reduction*

Data was compiled and reduced in *Tripoli* and *Redux* (Bowring et al., 2011; McLean et al., 2011). Initial 230Th disequilibrium in the 206Pb/238U system was corrected for each grain by estimating (Th/U)magma using a fixed (Th/U)zircon-magma partition coefficient ratio of 0.19±0.06 (1s) based on a compilation of natural zircon–melt pairs, and uncertainties for the (Th/U)magma were propagated into final date uncertainty for each grain. Isotopic ratios, compositional data, and ages for all grains is available in Table S3. Analytical uncertainty (±X) at the 95% confidence level for the U-Pb date of each grain was utilized for further weighted-mean age calculations and Bayesian eruptive age models. Weighted-mean ages were calculated in *Redux,* and are summarized in Table SM1. Bayesian eruptive age estimates were modeled using the package *Chron.jl* (Keller et al., 2018). While the ages reported in the manuscript utilize a uniform prior crystallization distribution, model ages calculated for several different prior crystallization distributions are tabulated in Table SM1.

***1.5 Total Organic Carbon analyses***

Shale samples were drilled with a 3/8” tungsten carbide bit on a drill press to generate ~15mL of unconsolidated shale powder. At Amherst College, aliquots of the powder were weighed, decarbonated with 3 N hydrochloric acid, rinsed with deionized water, dried in a 60°C oven, and reweighed. Dried insoluble residues were rehomogenized with mortar and pestle. Insoluble residue powders were loaded into tin capsules and combusted in a Costech ECS 4010 elemental analyzer interfaced with a Sercon 20-22 gas source stable isotope ratio mass spectrometer. Carbon content of samples was measured using beam area calibrated by a pure acetanilide standard. Acetanilide check standards were run after every fifth sample. TOC was calculated as the product of the percent insoluble residue of the whole rock and the percent carbon of the insoluble residue. Uncertainty in TOC measurement is +/- 1.8% (1 s.d.) based on replicate analyses of insoluble residues and replicate extractions of whole rock powders. Data is compiled in Table SM4.

TOC data from prior publications were extracted from published figures (John et al., 2002; Föllmi et al., 2005; Isaacs et al., 2001; Bohacs and Swalbach, 1994), and integrated into our measured sections by matching tie points from the measured sections in those publications to our stratigraphic framework. For each measured section, an exponential moving average (alpha of .05) was calculated with the compiled TOC data in order to characterize trends in TOC abundance, and appended with a 1σ analytical uncertainty envelope.

***1.6 X-Ray Fluorescence analyses***

*1.6.1 Sample preparation*

Shale powders were generated with a 3/8” tungsten carbide bit, following the same procedure detailed above for TOC analyses. 6-10mL aliquots of unconsolidated powder were placed into cylindrical Teflon pucks (32mm OD), the lower ends of which were covered with layer of transparent polypropylene thin film. The shale powder was firmly compacted against the thin film, using the capped top of a 20mL Falcon sample tube as a tamper. In addition to homogenizing the shale sample, powdering and compacting each sample minimized potential analytical variability due to matrix effects and/or crystal/grain orientation. Thicknesses of the resultant compacted shale powder always exceeded 2.7mm, or the maximum penetration depth of x-rays on a compacted substrate from the ThermoNiton XL5 Plus portable handheld XRF (pXRF) unit used to analyze the samples.

*1.6.2 XRF measurements*

Each compacted puck was placed into a shielded box attached to the emitting end of the pXRF unit, which was run in a benchtop configuration for all samples. Each measurement involved 30 seconds of analysis with the pXRF’s Main filter, and subsequent 60 second analysis windows using the unit’s integral Low, High, and Light filters for a total of ~210 seconds per sample. During each analysis session, a variety of shale standards, including USGS standards MAG-1, SCO-1, and SDC-1 (Gladney, 1980) and SDAR-L2 and SDAR-M2 (Wilson et al., 2021), were run multiple times at the beginning of each session (typically for ~20 minutes to allow the instrument’s internal temperature to stabilize). Standards were subsequently interspersed amongst every ~8 unknowns.

*1.6.3 Data reduction*

Elemental abundances for each standard measurement were compiled, and compared to the published elemental abundance data for that standard (generated either by ICP-MS or via XRF). For each element, the relationships between published and measured abundances for each standard were plotted, and a linear model fit was applied. The standard deviation of each group of analyses for each element of each standard was calculated, and then averaged across all standards for each element, yielding an approximation of analytical uncertainty that more appropriately encapsulates external variables than the modeled analytical uncertainty provided by the pXRF’s integral software. The standard deviation of the range of standard measurements for each element was doubled, and then added in quadrature to the elemental abundance uncertainties for each element of each unknown analysis. For elements with a model linear fit with a Pearson correlation coefficient of greater than .7, a linear calibration factor (the equivalent of correcting the model fit line to the 1:1 line) was applied to that elemental abundance in all measured unknowns. For elements with a model fit correlation coefficient of less than .7, no calibration factor was applied. Pearson correlation coefficients for linear fits for Al, Ca, Si, and Ti were calculated to be .53, .90, .96, and .97 respectively. For Al, multiple measurements from standards MAG-1, SDAR-M2, SDAR-L2, SCO-1, and SDC-1 overlapped, within uncertainty, the 1:1 line for Al. As such, no correction, other than additional uncertainty, was applied to measured Al abundances on unknowns. Calibrated elemental abundances of Al, Ca, Si, and Ti in El Capitan shales are summarized in Table SM5.

**2. ORGANIC CARBON MASS ACCUMULATION RATES IN THE BELRIDGE FIELD**

Estimated organic carbon mass accumulation rates (OCMARs) from the Belridge Field of the San Joaquin Basin, an expanded, clastic-rich Monterey Formation locality, are similar to those of the Santa Barbara Basin Monterey Formation sections and modern Californian borderland basins (Fig. 9, main text). The Macdonald Shale and Antelope Shale are members of the Monterey Formation found in the Belridge oil field (Schwartz, 1988), with the Macdonald Shale interpreted as a hemipelagic, deepwater deposit (Graham and Williams, 1985). Together, the Macdonald and Antelope shales are over 500 m thick (from Well 856C-7 in Farrell, 2020), with deposition of the Macdonald Shale occurring between ~13.5–10 Ma and the Antelope Shale between ~10–6.5 Ma (Scheirer and Magoon, 2007). Using these rough age constraints and member thicknesses from Farrell (2020), the Macdonald and Antelope shales have average sedimentation rates of 55.7 m/Myr and 90.57 m/Myr respectively. Farrell (2020) calculates average TOC contents for several different facies of an interval including the top of the Macdonald Shale and majority of the Antelope Shale, with values ranging from 4–11%. Using these values as approximate maxima and minima for average TOC content, and assuming a relatively high bulk shale density of 2.2 g/cm3, OCMAR for the Macdonald Shale ranges from ~490–1350 g/cm2/kyr, and ~800–2200 g/cm2/kyr for the Antelope Shale. While these values are generally higher than those observed in the coastal SBB, they are still within the range of values expected for OCMAR on productive margins (Fig. 9, main text).

**3. SENSITIVITY TEST OF 𝛿13C BOX MODEL**

In order to assess the impact of different input variables on the results of the steady-state carbon isotopic box model outlined in Section 5.5 of the main text, we input ranges of 𝛿13Cin and 𝛿13Corg values (between -12 and 0‰, and -27 and -19‰, respectively) into Eq. 3. Results of this sensitivity test are shown as contour plots in Fig S5, with a red triangle denoting the position of the results provided by the assumed compositions for 𝛿13Cin (-6‰) and 𝛿13Corg (-23‰) that are currently utilized in the main text. Fig. SM5A depicts calculated forg values for a pre-Monterey Event scenario, the above 𝛿13Cin and 𝛿13Corg compositional ranges, given a steady-state total carbon burial flux of 0.6GtC/yr and a 𝛿13Ccarb composition of 0.5‰. Calculated forg values for Monterey Event conditions, with a 𝛿13Ccarb composition of 1.5‰ and all other input variables identical to those in the previous panel, are shown in Fig. SM5B. Fig. SM5C depicts the change in forg required to shift from pre- to syn-Monterey Event 𝛿13Ccarb compositions for a given combination of 𝛿13Cin and 𝛿13Corg values. Note that while the combination of enriched input carbon and organic carbon isotopic compositions results in the lowest forg (Fig. SM5A,B), this region of the parameter space also results in the largest Δ forg required to drive a ~1‰ shift in marine 𝛿13C (Fig. SM5C).

In Fig. SM5D, we examine the impact of various 𝛿13Cin and 𝛿13Corg values on the sustained OCMAR, distributed across the entire area of Neogene circum-Pacific basins (~600,000km2, Vincent and Berger, 1985) that would be required to drive the ~1‰ 𝛿13C shift of the Monterey Event. In general, more-enriched input carbon and organic carbon compositions result in higher required organic burial fluxes, with only the most depleted putative 𝛿13Cin values (<~-11‰) generating required OCMARs below 2000 mg/cm2/kyr.

Changes in the magnitude of total carbon burial flux (conservatively estimated to be ~0.6GtC/yr, following Li et al., 2023), would have a linear impact on the magnitude of the required burial fluxes described above: a smaller total carbon burial flux would necessitate a proportionally smaller OC burial flux. However, for the range of 𝛿13Cin and 𝛿13Corg compositions considered in this sensitivity test, halving the total estimated C burial flux (to ~0.3GtC/yr) results in sustained OCMARS required across Neogene circum-Pacific basins that are still higher than most OCMARs observed anywhere in the SBB, and that are between ~5-10 times larger than the OCMARS observed in syn-MCO SBB strata. As such, we suggest that even for a range of potential 𝛿13Cin, 𝛿13Corg, and global carbon burial fluxes, organic carbon burial in the Monterey Formation and similar circum-Pacific basins along cannot explain the ~1‰ marine 𝛿13C shift of the Monterey Event.

\* \* \*

**References Cited**

Bohacs, K.M. and Schwalbach, J.R., 1994, Natural gamma-ray spectrometry of the Monterey Formation at Naples Beach, California: insights into lithology, stratigraphy, and source-rock quality, *in* Field Guide to the Monterey Formation between Santa Barbara and Gaviota, California: AAPG Pacific Section, p. 85-94

Bowring, J.F., McLean, N.M. and Bowring, S.A., 2011, Engineering cyber infrastructure for U‐Pb geochronology: Tripoli and U‐Pb\_Redux: Geochemistry, Geophysics, Geosystems, 12(6).

Condon, D.J., Schoene, B., McLean, N.M., Bowring, S.A. and Parrish, R.R., 2015, Metrology and traceability of U–Pb isotope dilution geochronology (EARTHTIME Tracer Calibration Part I): Geochimica et Cosmochimica Acta, 164, p. 464-480.

Farrell, J.C., 2020, Lateral Variation of Fine-Grained, Organic-Rich Sediments of the Miocene Monterey Formation, Belridge Field Area, California [Masters thesis]: California State University, Long Beach, 81p.

Föllmi, K.B., Badertscher, C., de Kaenel, E., Stille, P., John, C.M., Adatte, T. and Steinmann, P., 2005, Phosphogenesis and organic-carbon preservation in the Miocene Monterey Formation at Naples Beach, California—The Monterey hypothesis revisited:Geological Society of America Bulletin, 117(5-6), p. 589-619.

Gerstenberger, H. and Haase, G., 1997, A highly effective emitter substance for mass spectrometric Pb isotope ratio determinations: Chemical geology, 136(3-4), p. 309-312.

Gladney, E.S., 1980, Compilation of elemental concentration data for the United States Geological Survey's six geochemical exploration reference materials (No. LA-8473-MS): Los Alamos Scientific Lab., NM (USA).

Graham, S.A. and Williams, L.A., 1985, Tectonic, depositional, and diagenetic history of Monterey Formation (Miocene), central San Joaquin basin, California: AAPG bulletin, 69(3), p. 385-411. doi:<https://doi.org/10.1306/AD4624F7-16F7-11D7-8645000102C1865D>

Hiess, J., Condon, D.J., McLean, N. and Noble, S.R., 2012, 238U/235U systematics in terrestrial uranium-bearing minerals: Science, 335(6076), p.1610-1614.

Hoke, G.D., Schmitz, M.D. and Bowring, S.A., 2014, An ultrasonic method for isolating nonclay components from clay‐rich material: Geochemistry, Geophysics, Geosystems, 15(2), p. 492-498.

Horstwood, M.S., Košler, J., Gehrels, G., Jackson, S.E., McLean, N.M., Paton, C., Pearson, N.J., Sircombe, K., Sylvester, P., Vermeesch, P. and Bowring, J.F., 2016, Community‐derived standards for LA‐ICP‐MS U‐(Th‐) Pb geochronology–Uncertainty propagation, age interpretation and data reporting: Geostandards and Geoanalytical Research, 40(3), p. 311-332.

Jackson, S.E., Pearson, N.J., Griffin, W.L. and Belousova, E.A., 2004, The application of laser ablation-inductively coupled plasma-mass spectrometry to in situ U–Pb zircon geochronology: Chemical geology, 211(1-2), p. 47-69.

John, C.M., Follmi, K.B., De Kaenel, E., Adatte, T., Steinmann, P. and Badertscher, C., 2002, Carbonaceous and phosphate-rich sediments of the Miocene Monterey formation at El Capitan State Beach, California, USA: Journal of Sedimentary Research, 72(2), p. 252-267.

Keller, C.B., Schoene, B., Samperton, K.M. (2018), A stochastic sampling approach to zircon eruption age interpretation: Geochem. Persp. Let. 8, p. 31–35.

Krogh, T.E., 1973, A low-contamination method for hydrothermal decomposition of zircon and extraction of U and Pb for isotopic age determinations: Geochimica et Cosmochimica Acta, *37*(3), pp.485-494.

Kylander-Clark, A.R., Hacker, B.R. and Cottle, J.M., 2013, Laser-ablation split-stream ICP petrochronology: Chemical Geology, 345, p. 99-112.

Li, Z., Zhang, Y.G., Torres, M. and Mills, B.J., 2023, Neogene burial of organic carbon in the global ocean: Nature, 613(7942), p.90-95. doi:https://doi.org/10.1038/s41586-022-05413-6

Mattinson, J.M., 2005, Zircon U–Pb chemical abrasion (“CA-TIMS”) method: combined annealing and multi-step partial dissolution analysis for improved precision and accuracy of zircon ages: Chemical Geology, 220(1-2), p. 47-66.

McLean, N.M., Bowring, J.F. and Bowring, S.A., 2011, An algorithm for U‐Pb isotope dilution data reduction and uncertainty propagation: Geochemistry, Geophysics, Geosystems, 12(6).

McLean, N.M., Condon, D.J., Schoene, B. and Bowring, S.A., 2015, Evaluating uncertainties in the calibration of isotopic reference materials and multi-element isotopic tracers (EARTHTIME Tracer Calibration Part II): Geochimica et Cosmochimica Acta, 164, p. 481-501.

Paton, C., Woodhead, J.D., Hellstrom, J.C., Hergt, J.M., Greig, A. and Maas, R., 2010, Improved laser ablation U‐Pb zircon geochronology through robust downhole fractionation correction: Geochemistry, Geophysics, Geosystems, *11*(3).

Schwartz, D.E., 1988, Characterizing the lithology, petrophysical properties, and depositional setting of the Belridge diatomite, South Belridge field, Kern County, California, *in* Graham, Sephan A., (ed.), Studies of the Geology of the San Joaquin Basin: Pacific Section SEPM, 60, p. 281-301.

Sláma, J., Košler, J., Condon, D.J., Crowley, J.L., Gerdes, A., Hanchar, J.M., Horstwood, M.S., Morris, G.A., Nasdala, L., Norberg, N. and Schaltegger, U., 2008, Plešovice zircon—a new natural reference material for U–Pb and Hf isotopic microanalysis*:*Chemical Geology, 249(1-2), p. 1-35.

Szymanowski, D. and Schoene, B., 2020, U–Pb ID-TIMS geochronology using ATONA amplifiers: Journal of Analytical Atomic Spectrometry, 35(6), p. 1207-1216.

Vermeesch, P., 2018, IsoplotR: A free and open toolbox for geochronology: Geoscience Frontiers, 9(5), p. 1479-1493.

Wiedenbeck, M.A.P.C., Alle, P., Corfu, F.Y., Griffin, W.L., Meier, M., Oberli, F.V., Quadt, A.V., Roddick, J.C. and Spiegel, W., 1995, Three natural zircon standards for U‐Th‐Pb, Lu‐Hf, trace element and REE analyses: Geostandards newsletter, 19(1), p. 1-23.

Wendt, I. and Carl, C., 1991, The statistical distribution of the mean squared weighted deviation: Chemical Geology: Isotope Geoscience Section, 86(4), p. 275-285.

Wilson, S.A., Potts, P.J. and Webb, P.C., 2021, Three New Geochemical Reference Materials for Mineral Exploration and Environmental Contamination Studies: SdAR‐L2, SdAR‐M2 and SdAR‐H1: Geostandards and Geoanalytical Research, 45(2), p. 359-367.

**\* \* \***

**SUPPLEMENTARY MATERIAL TABLE AND FIGURE CAPTIONS**

**Table SM1.** Compilation of bentonite samples dated with U-Pb geochronology. Sample locations, reference sections, and stratigraphic heights are included, along with calculated weighted mean LA-ICPMS ages, TIMS weighted mean ages, and modeled Bayesian eruptive ages utilizing a variety of prior crystallization distributions. Ages used in the age-height model are in bold text.

**Table SM2.** Compiled LA-ICPMS U-Pb data. Analyses used to calculate weighted mean ages are in bold text, while analyses interpreted as detrital or inherited are italicized.

**Table SM3.** Compiled CA-ID-TIMS U-Pb data. Analyses used to calculate weighted mean and modeled eruptive ages are in bold text.

**Table SM4.** Compiled total organic carbon and 13Corg from El Capitan shales.

**Table SM5.** Compiled XRF data (Al, Ca, Si, and Ti abundances) from El Capitan shales.

**Figure SM1.** Visualization (ranked-age and concordia diagrams) of geochronological analyses used to generate ages for the following samples from El Capitan: A) EAGC2010 (LA-ICPMS and TIMS), B) EAGC2011 (LA-ICPMS and TIMS), C) EAGC2012 (LA-ICPMS and TIMS), D) EAGC2001 (LA-ICPMS), E) EAGC2002 (LA-ICPMS), F) EAGC2003 (LA-ICPMS and TIMS MDA), G) EAGC2006 (LA-ICPMS and TIMS), H) EAGC2008 (LA-ICPMS and TIMS), I)F1917-33.7 (TIMS), J) F1917-13.4 (LA-ICPMS and TIMS MDA), K) F1917-8.2 (LA-ICPMS and TIMS), L) F1917-0.7 (LA-ICPMS and TIMS), M) M1801-0.73 (LA-ICPMS), N) EAGC2009 (LA-ICPMS MDA).

**Figure SM2.** Visualization (ranked-age and concordia diagrams) of geochronological analyses used to generate ages for the following samples from Naples Beach: A) EA1902-16.6 (LA-ICPMS and TIMS), B) EA1902-12.34 (LA-ICPMS and TIMS MDA), C) EA1902-12.0 (LA-ICPMS), D) EA1902-11.73 (LA-ICPMS and TIMS, E) EA1902-11.55 (LA-ICPMS), F) EA1902-9.2 (LA-ICPMS), G) F1916-24.9 (LA-ICPMS), H) M1806 (LA-ICPMS), I) M1805 (LA-ICPMS), J) M1804\_6.5 (LA-ICPMS), K) M1808 (LA-ICPMS).

**Figure SM3.** Visualization (ranked-age and concordia diagrams) of geochronological analyses used to generate ages for the following samples from Tajiguas Beach: A) EAGC2016 (LA-ICPMS), B) EAGC2013 (LA-ICPMS), C) EAGC2014 (LA-ICPMS), D) EAGC2107 (LA-ICPMS), E) EAGC2103 (LA-ICPMS and TIMS MDA).

**Figure SM4.** Visualization (ranked-age and concordia diagrams) of geochronological analyses used to generate ages for the following samples from Tajiguas Beach: A) F1914-3.6 (LA-ICPMS and TIMS), B) F1915-7.2 (LA-ICPMS and TIMS MDA).

**Figure SM5.** Sensitivity test of the carbon isotope box model defined in Eq. 3 and Fig. 10 of the manuscript. A) forg calculated for pre-Monterey Event conditions ( 𝛿13Ccarb of ~0.5‰), given a range of 𝛿13Cin and 𝛿13Corg values and a global carbon burial flux of ~0.6GtC/yr. B) forg calculated for syn-Monterey Event conditions ( 𝛿13Ccarb of ~1.5‰), given a range of 𝛿13Cin and 𝛿13Corg values and a global carbon burial flux of ~0.6GtC/yr. C) change in forg (Δ forg) required to drive a shift in 𝛿13Ccarb from ~0.5‰ to ~1.5‰ for a given combination of 𝛿13Cin and 𝛿13Corg values. D) Sustained organic carbon mass accumulation rates (OCMARs) in circum-Pacific basins (total area of 600,000km2) required to explain a shift in 𝛿13Ccarb from ~0.5‰ to ~1.5‰ through organic carbon burial in the Monterey Formation and equivalent strata for a given combination of 𝛿13Cin and 𝛿13Corg values. In all panels, the red triangle indicates the model output for the assumed 𝛿13Cin and 𝛿13Corg compositions (-6‰ and -23‰, respectively) discussed in the main text.

\* \* \*