Deep-Water Warming in the Gulf of Mexico from 2003 to 2019

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ABSTRACT: A key consequence in climate change is the warming of deep waters, away from the faster warming rates of near-surface subtropical and tropical waters. Since surface and near-surface oceanic temperatures have been measured far more frequently in time and space than deep waters (>2000 m), deep measurements become quite valuable. Semi-enclosed basins, such as the Gulf of Mexico, are of particular interest as the waters below sills that connect with the neighboring oceans have residence times much longer than upper layers. Within the western Gulf of Mexico, near-bottom measurements at ~3500-m depths at four sites show a stable linear warming trend of $\sim 16 \pm 2 \text{ m}^\circ\text{C decade}^{-1}$ for the period 2007–18, and CTD data from eight oceanographic cruises occurring from 2003 to 2019 show a trend of $\sim 18 \pm 2 \text{ m}^\circ\text{C decade}^{-1}$ from the bottom to ~2000 m below the surface. The bottom geothermal heat flux is a contributing factor to be considered in the warming and renewal of such waters, but it has not changed over millennia and is therefore unlikely to be the cause of the observed trend. The densest waters that spill into the Gulf of Mexico, over the Yucatan Channel sill, must mix substantially during their descent and in the near-bottom interior, losing their extreme values. A simple box model connects the observed warming, well within the Gulf interior, with that expected in the densest waters that spill from the North Atlantic into the Cayman Basin through Windward Passage and suggests that the source waters at the entrance to the Caribbean have been warming for at least 100 years.

KEYWORDS: Buoyancy; Climate change; Temperature; In situ oceanic observations; Time series; Decadal variability

1. Introduction

The large amount of studies aiming at measuring and understanding climate change shows the importance of having long and accurate temperature records. The outstanding Argo program provides measurements from the top 2000 m of the global ocean since 2006 (Roemmich et al. 2015; Riser et al. 2016). A global average estimate shows warming, over the last decade, of $50 \text{ m}^\circ\text{C decade}^{-1}$ for waters between the surface and 500 m and of $20 \text{ m}^\circ\text{C decade}^{-1}$ from 500 to 2000 m (Roemmich et al. 2015). Measuring programs, like Argo, together with historical and quality controlled hydrographic data are the prime sources of empirical information for ocean climate

change studies. However, a particular constraint on these studies is the limited amount of high-quality measurements in the abyssal region, in waters deeper than ~2000 m (e.g., Wunsch, 2016; Johnson et al. 2015, to mention a few). Notably, the recent study by Desbruyères et al. (2016) offered a warming estimate of $20 \pm 12 \text{ m}^\circ\text{C decade}^{-1}$ for the Atlantic Ocean at ~2000 m for the time period 2000–16 (see their Fig. 2b). Kouketsu et al. (2011, see their Fig. 3) estimated global ocean warming, for depth levels between 3000 and 4000 m of the same magnitude as the rates presented here for the deep Gulf of Mexico (GOM), but with large uncertainties. Recent estimates of global surface rate of warming derived from Argo floats is $\sim 0.15\text{--}0.2^\circ\text{C decade}^{-1}$, but decreases rapidly with depth and is typically $10\text{--}30 \text{ m}^\circ\text{C decade}^{-1}$ from 400 to 2000 m (e.g., Wijffels et al. 2016, Johnson et al. 2019). For the western North Atlantic, a 73-yr hydrographic record at Bermuda indicates that the layer between 1500 and 2500 dbar shows a secular warming trend of $0.5^\circ\text{C century}^{-1}$ (Joyce and Robbins, 1996). Also, comparing hydrographic surveys taken 43 years apart in the western North Atlantic, a latitudinal average between 20° and 35°N gives a warming trend of $30 \pm 10 \text{ m}^\circ\text{C decade}^{-1}$ at 2000-m depth (Joyce et al. 1999, see their Fig. 7). Clearly there is a wide range of warming trends in the available literature for the 2000-m layer in the North Atlantic.

In this study, we use moored measurements at four sites just ~10 m above the bottom at ~3500 m, within the western GOM

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FIG. 1. The map shows the locations of moored thermometers as filled magenta circles at the western side of the Gulf of Mexico, and CTD profiles in other symbols: black diamonds and dots for the western Gulf of Mexico (WGM), red pentagrams for the sill region in the Yucatan Channel (SILL), inverted green triangles for the Yucatan Basin (YUB), and blue squares for the Venezuela Basin (VEB). The Yucatan Channel sill depth is 2040 m. The 2000-m isobath surrounds five basins: WGM, YUB, Cayman Basin (CAY), Colombia Basin (COL), and VEB. The two deep connections between the Caribbean Sea and the Atlantic Ocean are the Windward Passage (WP) and the Anegada–Jungfern Passage complex (AJP). The blue contours between Florida and the Bahamas indicate the 725-m isobath. Black dots within the WGM differentiate the 19 shallower profiles from the 18 profiles deeper than 3700 dbar, in diamonds.

taken between 2007 and 2018, that show much smaller uncertainty than those of Kouketsu et al. (2011) and Desbruyères et al. (2016). Our CTD measurements indicate a very similar warming trend from the bottom up to ~2000-m depth. The CTD measurements also show the properties of the densest waters that enter the gulf just above the Yucatan Channel sill, which is located between the Caribbean Sea and the GOM. These waters reach the Yucatan Sill via flows from the Windward Passage (Smith 2010).

The following section presents the data and results. The third section elaborates on results from this and other studies, in particular for the Caribbean Sea (Joyce et al. 1999; MacCready et al. 1999). The fourth section summarizes the analysis and results. Two appendixes deal with instrumental drift, the warming trend estimation, and the calculation of confidence intervals in the warming trends reported here from moored observations.

2. Data and results

One source of data for this study comes from moored thermometers at ~10 m above the bottom at depths of ~3500 m at four GOM sites, with time series extending close to 10 years. The moored instruments (Sea-Bird Electronics Inc., SBE37SM) collected data during consecutive periods, each lasting ~1–2 years from deployment to recovery (Fig. 1). The moored time series gaps, which mostly correspond to the time encompassing from recovery to next deployment, are typically shorter than 3 days, except for one measurement period in one site which was

completely lost. In addition to the moored time series, CTD profiles from a variety of oceanographic cruises complement the data (Fig. 2 and Table 1). These profiles are relevant for the analysis and interpretation of the warming of GOM's deep waters.

a. CTD profiles

The selected CTD profiles allow a direct comparison of properties between the Caribbean Sea and the GOM, in particular, just north of the Yucatan Channel sill (red symbols in Fig. 1, denoted by SILL in Table 1). We use TEOS-10 for seawater properties unless otherwise noted (IOC et al. 2010; McDougall and Barker 2011). The set of 37 CTD profiles, well within the deep western side of the GOM (WGM in Table 1), also permits an estimate of the warming from the bottom to ~2000 m below the surface. The 37 WGM profiles, as labeled in Table 1 and shown in black symbols in Fig. 1, come from eight cruises that span from August 2003 to June 2019 and reach at least 3000 m.

Distributions of Conservative Temperature, salinity and density against pressure, and of Conservative Temperature against salinity highlight the hydrographic differences among the basins (Fig. 2). The very deep (pressures exceeding 4500 dbar) fractions of the profiles in the Venezuela Basin, identified as VEB in Table 1, are out of range in the distributions against pressure of Fig. 2 (blue traces).

The temperature profiles in the western GOM (Fig. 3) show clearly a progressive increase of deep temperatures. The temperature increase is quite uniform from ~2000 m down to

FIG. 2. Following the same color distinction as in Fig. 1, the distribution of depth vs (a) σ_0 , (b) Absolute Salinity S_A , and (c) Conservative Temperature Θ ; (d) the corresponding Θ – S_A diagram with σ_0 contours. The same coloring in profiles is used as the symbol for its corresponding location in Fig. 1: black diamonds (WGM), red pentagrams (SILL), inverted green triangles (YUB), and blue squares (VEB). These abbreviations, of geographical region, are used in Table 1. The rectangle in (c) is shown expanded in Fig. 3.

the deepest common level (i.e., 3640 m at 3700 dbar). The trend is shown in Fig. 4, where a conventional least squares fit provides an estimate of the warming rate and its confidence interval. A warming of $18 \text{ m}^\circ\text{C decade}^{-1}$, at such depths, translates in an increase of energy of $2.5 \times 10^5 \text{ W km}^{-3}$. The volume below 2000 m within the GOM is $\sim 7 \times 10^5 \text{ km}^3$, hence the rate of energy increase in such volume is $\sim 1.7 \times 10^{11} \text{ W}$ (or $\sim 5.4 \times 10^{18} \text{ J yr}^{-1}$).

b. Moored thermometer data

Further observations for this study come from moored thermometers at four locations, called from north to south PER, LMP, ARE and LNK (magenta circles in Fig. 1, Table 2). Table S1 (in the online supplemental material) shows the locations, dates, and instrument's serial number. The raw data have sampling intervals of 2, 5, or 15 min. These data are filtered to avoid aliasing when resampling to 1-h interval. The set of in situ (ITS-90) temperature time series show sharp discontinuities, expected due to different instrument deployment depths and offsets from thermometer calibration (Fig. 5). Instrumental calibration drift magnitudes are relatively small compared to the observed temperature trends (appendix A).

There is an extensive processing of meteorological data to account for the effects and to eliminate “mean shifts” that are produced by changes in instrument location (Venema et al. 2018). In our case, the knowledge of precisely when such discontinuities occur facilitates a great deal the data processing.

To concatenate continuous time series consistently, the mean temperature of each fraction of the time series, as defined from deployment to the recovery of the mooring, is modified for an updated mean, without changing the full series mean. The set of individual means plus the two parameters of the standard linear fit become the free parameters of the least squares fit. Details of the fitting algorithm are given in appendix B. The individual temperature records as well as the reconstructed time series for each mooring indicate that typical deviations of such means are on the order of $1 \text{ m}^\circ\text{C}$ (Fig. 5), which in terms of the general in situ temperature gradient (see Fig. 3), represents a vertical displacement of 8 m. The largest mean shift of individual means is for the fourth record in PER with an anomaly of $3.3 \text{ m}^\circ\text{C}$, which corresponds to a vertical displacement of 26 m. To compute confidence intervals, of the slope (or temperature trend), and in general of the free

TABLE 1. This table presents the locations, maximum pressure, and dates of CTD profiles plotted or used in the analysis. [Figures 2 and 3](#) do not include the last 19 profiles listed, to avoid crowded figures. All 37 WGM profiles listed are used in calculations of warming trends.

Region	Profile ID	Date	Lat (°N)	Lon (°W)	Max pressure (dbar)
VEB	A22_12_063	10 Apr 2012	17°21.70'	65°59.77'	4648
VEB	A22_12_064	10 Apr 2012	17°04.22'	66°28.23'	4848
VEB	A22_12_065	10 Apr 2012	16°41.98'	67°04.64'	4808
VEB	A22_12_066	11 Apr 2012	16°19.40'	67°41.26'	4732
YUB	CNK_05_104	27 Jun 2000	21°29.32'	84°59.67'	3182
YUB	CNK_08_021	20 Aug 2003	18°11.30'	87°00.61'	3588
YUB	XIX_06_056	6 Sep 2017	20°55.03'	85°37.23'	3146
YUB	CNK_42_030	21 Jul 2018	21°27.34'	84°55.38'	3605
SILL	CNK_05_029	19 Jun 2000	21°38.61'	85°35.93'	2001
SILL	CNK_05_030	19 Jun 2000	21°36.84'	85°42.39'	2030
SILL	CNK_05_031	19 Jun 2000	21°35.85'	85°48.09'	1979
SILL	CNK_05_032	19 Jun 2000	21°36.02'	85°54.34'	1909
SILL	CNK_42_036	21 Jul 2018	21°43.00'	85°38.12'	2074
SILL	CNK_42_037	21 Jul 2018	21°42.57'	85°46.05'	2030
SILL	CNK_42_039	22 Jul 2018	21°42.41'	85°55.37'	1911
SILL	CNK_42_049	23 Jul 2018	21°39.93'	85°43.88'	2044
SILL	CNK_42_050	23 Jul 2018	21°39.61'	85°51.38'	1974
WGM	CNK_08_057	26 Aug 2003	23°04.34'	91°06.50'	3787
WGM	CNK_09_049	8 Sep 2004	24°40.26'	90°22.71'	3700
WGM	CNK_09_050	8 Sep 2004	24°35.38'	90°33.85'	3722
WGM	CNK_09_051	8 Sep 2004	24°29.69'	90°44.37'	3737
WGM	XIX_02_023	11 Jul 2010	22°59.99'	94°31.14'	3780
WGM	XIX_02_024	11 Jul 2010	22°59.99'	92°59.94'	3767
WGM	XIX_03_027	4 Mar 2013	23°00.02'	94°00.50'	3759
WGM	XIX_04_035	8 Sep 2015	23°00.02'	94°30.00'	3767
WGM	XIX_04_037	9 Sep 2015	23°00.15'	92°59.99'	3771
WGM	XIX_05_041	22 Jun 2016	22°59.11'	94°32.91'	3719
WGM	XIX_05_042	23 Jun 2016	23°00.05'	94°31.84'	3760
WGM	XIX_05_043	23 Jun 2016	23°00.68'	92°59.84'	3775
WGM	XIX_06_006	20 Aug 2017	23°00.09'	94°30.15'	3779
WGM	XIX_06_007	20 Aug 2017	23°59.56'	93°42.02'	3794
WGM	XIX_06_043	2 Sep 2017	22°59.64'	93°00.48'	3789
WGM	XIX_06_047	3 Sep 2017	24°00.48'	92°19.62'	3787
WGM	XIX_07_032	30 May 2019	25°00.15'	94°00.44'	3708
WGM	XIX_07_61	7 Jun 2019	23°00.00'	93°00.38'	3758
WGM	CNK_09_001	26 Aug 2004	25°16.20'	94°50.17'	3558
WGM	CNK_09_039	7 Sep 2004	24°35.89'	88°55.57'	3563
WGM	CNK_09_040	7 Sep 2004	24°39.60'	89°06.90'	3554
WGM	CNK_09_046	8 Sep 2004	24°55.09'	89°49.40'	3569
WGM	CNK_09_047	8 Sep 2004	24°50.05'	90°00.33'	3588
WGM	CNK_09_048	8 Sep 2004	24°45.31'	90°11.66'	3596
WGM	XIX_02_027	13 Jul 2010	22°00.34'	93°01.08'	3349
WGM	XIX_03_028	5 Mar 2013	23°00.02'	95°00.52'	3140
WGM	XIX_04_036	9 Sep 2015	23°00.17'	93°00.09'	3607
WGM	XIX_04_041	10 Sep 2015	22°00.13'	93°00.06'	3328
WGM	XIX_04_048	12 Sep 2015	21°30.12'	93°30.04'	3041
WGM	XIX_05_048	24 Jun 2016	22°00.03'	95°02.02'	3365
WGM	XIX_06_008	21 Aug 2017	23°59.92'	95°04.03'	3531
WGM	XIX_06_014	22 Aug 2017	24°51.74'	94°56.67'	3678
WGM	XIX_06_046	3 Sep 2017	24°00.07'	92°18.18'	3661
WGM	XIX_06_049	4 Sep 2017	24°00.70'	89°59.79'	3677
WGM	XIX_07_006	12 May 2019	24°51.80'	94°58.34'	3627
WGM	XIX_07_036	31 May 2019	25°00.08'	91°59.18'	3556
WGM	XIX_07_040	1 Jun 2019	25°00.37'	90°00.02'	3554

FIG. 3. The two frames show the deep end of 18 available profiles within the GOM that reach 3700 dbar: (left) the in situ temperature (ITS-90) and (right) the Conservative Temperature. The set of profiles in the right frame are the same as the ones shown within the black rectangle of Fig. 2c. The corresponding year of cruises is listed.

parameters, a Monte Carlo computation was performed, as shown in appendix B. The temperature rates of change range between 14.6 and 17.5 $\text{m}^{\circ}\text{C decade}^{-1}$ with confidence intervals around $1\text{--}2$ $\text{m}^{\circ}\text{C decade}^{-1}$ (Fig. 5).

3. Discussions

The bathymetry in use is from Smith and Sandwell (1997). For this study the Cayman Basin and Yucatan Basin can be considered as a single basin, since the deepest waters that have direct connection between them are at ~ 3900 m, and the deepest waters that enter the GOM are at ~ 2000 m. A similar situation holds for the Colombia Basin and the Venezuela Basin: the deepest waters that freely connect them are at ~ 4030 m (through the Aruba gap), while the sill between the Colombia Basin and Cayman Basin is found in the Jamaica Rise (i.e., JRS in Fig. 1) at ~ 1520 m.

The densest waters entering the GOM at the Yucatan Channel (see flow fields shown in Ochoa et al. 2001; Sheinbaum et al. 2002; Candela et al. 2019) are denser (see Fig. 2a and left frame of Fig. 6) than the waters in the deepest regions within the GOM. The densest waters at the deepest CTD casts within the western GOM (Fig. 2a) have σ_0 that appears at ~ 1800 m in the Yucatan Sill vicinity.

Surface cooling events within the GOM cannot produce waters as cold and dense as those found below 2000 m; the only known source of such waters are the deep inflows through the Yucatan Channel, as proposed by Sturges (2005). The moored temperature series shown are in the GOM's interior bottom

boundary layer (BBL), a layer that is supplied by the deep inflow at Yucatan Channel, likely in a cyclonic circulation (DeHaan and Sturges 2005; Pérez-Brunius et al. 2018). There is no evidence of significant outflow of deep GOM's interior BBL waters through the Yucatan Channel, as it was described by Banyte et al. (2018a) for the Panama Basin.

The depth extension of the warming, as evident from CTD profiles, is from the bottom up to ~ 2000 -m depth (Figs. 2–4) and is nearly uniform. It is remarkable that the vertical temperature gradient over such a thick layer shows no significant time dependence.

The geothermal bottom heat flux in the GOM, which is comparable to Mesozoic ocean basins in general, is $\sim 40\text{--}47$ mW m^{-2} (Nagihara et al. 1996). In contrast, there is an average geothermal heating 4 times larger in the Panama Basin in the Pacific Ocean, which is considered the primary driver of its abyssal circulation (Banyte et al. 2018b). The geothermal heat flux in the GOM, if distributed uniformly in a 200-m-thick stagnant layer, implies a warming of $\sim 15\text{--}18$ $\text{m}^{\circ}\text{C decade}^{-1}$. The observed warming is similar but consistent over a much thicker layer. Considering the volume below 2000 m and the bottom, and the observed 18 $\text{m}^{\circ}\text{C decade}^{-1}$ warming (see Fig. 4) results in a heat gain that is ~ 6 times larger than that expected by the geothermal flux over a stagnant layer.

Geothermal heating can be a substantial contributor to abyssal heat budgets, and Munk (1966) canonical model has become a key element for comparisons. Even though Munk (1966) canonical model is for interior layers away from the BBL, it is instructive to show that the temperature trend we

FIG. 4. (left) In situ temperatures anomalies at four different depths, from the 37 profiles listed as WGM in Table 1, are plotted along a conventional least squares fit. The individual slopes ($\text{m}^\circ\text{C/decade}^{-1}$), with corresponding confidence intervals at 95% confidence level, are depicted in the text as well as the corresponding depth. Offsets in the anomalies, for 3500, 3000, and 2000 m are -0.10° , -0.05° , and 0.05°C , respectively. The 2500-m depth anomalies are without offset. (right) The profile of slopes as a function of depth, and the band associated with the confidence interval.

report in this study is far too high to be reconciled with such model. Munk's model assumes a steady balance of advection via upwelling against turbulent diffusion specified by

$$w \frac{\partial \Theta}{\partial z} \approx K \frac{\partial^2 \Theta}{\partial z^2}, \quad (1)$$

where w is the vertical velocity, Θ is the Conservative Temperature, K is the vertical diffusivity of heat, and z is the vertical coordinate. In this model, the vertical velocity is the consequence of a continuous supply of dense (cold) waters in a distant BBL. The analytical solution to this equation, which accepts Θ as the lateral average including the sloping BBL (Munk and Wunsch 1998), is

$$\Theta(z) = \Delta \Theta \exp(wz/K) + \Theta_o, \quad (2)$$

and a straightforward least squares fit, via a numerical optimization, determines the three free parameters ($\Delta \Theta = 4.005 \times 10^{-3}^\circ\text{C}$, $\Theta_o = 4.053^\circ\text{C}$, and $w/K = 1.475 \times 10^{-3} \text{m}^{-1}$). The mean vertical profile of Conservative Temperature within the western GOM and the fitted profile in the 2000–3700-dbar depth range agree very well (Fig. 6). We disregard the difference between decibars and meters in these depth calculations.

The fitted ratio K/w or vertical scale is $\sim 700 \text{m}$, and to derive w or K there is a need for an external choice. This vertical scale is much the same as in Munk (1966) and Munk and Wunsch (1998) for the Pacific Ocean, although in those references, the depth range is 1000–4000 m. The use of $w = 1.6 \text{cm day}^{-1}$, as explained below, implies $K = 1.3 \text{cm}^2 \text{s}^{-1}$, much the same as Munk (1966) estimates. A significant discrepancy arises from the fact that the observed $\partial \Theta / \partial t$ (see units in right panel of Fig. 6) is one order of magnitude larger than $w \partial \Theta / \partial z$ [or, given Eq. (1), than $K \partial^2 \Theta / \partial z^2$]. Therefore, the term $\partial \Theta / \partial t$, where t is time, cannot be neglected. A balance like $\partial \Theta / \partial t + w \partial \Theta / \partial z \approx 0$ rather than Eq. (1) requires interior downwelling velocities on

TABLE 2. Locations of four moorings, with number of deployments and records lost.

Moorings	No. of successful deployments	Records total/lost	Lat ($^\circ\text{N}$)	Lon ($^\circ\text{W}$)
PER	8	8/0	25°12.3'	95°02.2'
LMP	9	9/0	23°22.6'	95°07.3'
ARE	9	9/0	22°29.6'	95°00.7'
LNK	9	10/1	21°51.0'	94°02.3'

FIG. 5. The four frames show the temperature time series for locations shown in magenta circles in Fig. 1. From north to south these are PER, LMP, ARE, and LNK. Subplot titles show estimated temperature rate of change and its 90% confidence interval. The superimposed green “reconstruction” that avoids the discontinuities, with an offset of 5 m°C, shows the trend in black. Deviations of the individual record means from the reconstruction are shown as the black horizontal traces centered at the ordinate of 4.42°C (see also Fig. B1).

the order of $w \sim 8 \times 10^{-5} \text{ m s}^{-1}$ ($\sim 700 \text{ cm day}^{-1}$). Regarding the other limit, considering a balance like $\partial\Theta/\partial t \approx K\partial^2\Theta/\partial z^2$ implies $K = 30 \text{ cm}^2 \text{ s}^{-1}$, a value much too large to be realistic.

Besides the geothermal heating, the role of the sloping BBL, given the differential intensity of mixing relative to the same isopycnal levels in the interior, is of paramount importance (Kunze et al. 2012; Ferrari et al. 2016; Holmes et al. 2018). In Munk and Wunsch (1998), the role of lateral advection was introduced. This allows for nonuniform distribution of upwelling velocities as well as eddy coefficients. This perspective is consistent with the study of Holmes et al. (2018), in which interleaving exchanges occur (mainly along isopycnals) between interior and BBL waters (along sloping bottoms). Within this scheme or perspective, an expected feature for the GOM is the interleaving of the spilling waters that cross the sill and mix during their downhill descent.

The Windward Passage has a sill depth of 1680 m, whereas the Anegada–Jungfern Passage complex has a maximum depth of 1815 m (Smith 2010). The Venezuela Basin, for waters below the Anegada–Jungfern passage sill, show no temperature trend before 1970, followed by a $\sim 10 \text{ m}^\circ\text{C decade}^{-1}$ warming from 1970 to 2003 (Johnson and Purkey 2009). Hydrographic data (Fig. 2b) suggest that the deep waters that flow into the GOM through the Yucatan Channel come from the Windward Passage, transiting the Cayman Sea to the Yucatan Basin. Notice that the deep waters near the Anegada–Jungfern

Passage complex have lower salinities relative to the deepest waters at the entrance to the GOM. Bulgakov et al. (2003) and Smith (2010) have measured the flow of deep waters into the Cayman Basin through the Windward Passage. Although snapshots of the flow in the Windward Passage show inflows and outflows below $\sim 1200 \text{ m}$, the persistent flow at its sill is from the Atlantic Ocean into the Caribbean Sea (see Fig. 14 of Smith 2010).

A warming trend of $30 \text{ m}^\circ\text{C decade}^{-1}$ at the levels around 2000 m has been reported from CTD sections in the western subtropical North Atlantic Ocean (1950s–1990s; Joyce et al. 1999) and $50 \text{ m}^\circ\text{C decade}^{-1}$ in the Bermuda station “S” for the layer 1500–2500 m (1922–95; Joyce and Robbins 1996). Such a warming trend of inflowing waters through the Anegada–Jungfern Passage contributes to the warming inside the Venezuela Basin (Joyce et al. 1999). For the GOM, the incoming deep flows are not so intermittent as in the Venezuela Basin, and there is evidence of vigorous mixing during their downslope transit (see Fig. 5a of Pérez-Brunius et al. 2018). Extensive mixing under similar conditions is also suggested by the conceptual model of Bryden and Nurser (2003).

An approximate volume of the BBL over the GOM’s interior, limited by the 2000-m isobath is the horizontal area comprised within such isobath, which is $6.2 \times 10^5 \text{ km}^2$, times a thickness of 100 m above bottom, i.e., $6.2 \times 10^4 \text{ km}^3$. If we consider a certain inflow below $\sim 1800 \text{ m}$ through the Yucatan

FIG. 6. Mean vertical profiles within GOM (black traces) and fit of the analytical solution [Eq. (2)] below 2000 dbar in magenta trace. See text for the choice of $w \sim 1.6 \text{ cm day}^{-1}$ used in the right panel. The red traces correspond to the mean profile at the Yucatan Sill.

Channel [see Fig. 3a of Candela et al. 2019 for a cross section of the mean flow, and Table S3] that feeds the deep GOM's BBL, then the same volume has to be removed from the deep GOM by upwelling and flows along the bottom, and eventually be exported as lighter waters. For this inflow transport to replace the estimated BBL volume, a time lapse of 20 ± 14 or 25 ± 20 years is required, depending on if the transport is calculated from below 1790 or 1810 m, respectively. The confidence limits are at 80% in these estimates, and the considerable variability relative to the mean in deep transport estimates implies large uncertainties. Longer estimates of residence time are obtained using the same transports, but assuming a thicker layer than the BBL, for example, by replacing the waters below the isopycnal at $\sim 1800 \text{ m}$ in the Yucatan Channel. The density at $\sim 1800 \text{ m}$ at the Yucatan Channel ($\sigma_0 \approx 27.7618$) is found at $\sim 2700 \text{ m}$ in the GOM interior. The area extending horizontally within the Gulf's 2700-m isobath is $4.7 \times 10^5 \text{ km}^2$ enclosing a volume of $3.1 \times 10^5 \text{ km}^3$ and the transports below $\sim 1800 \text{ m}$ at Yucatan Channel are $0.10 \pm 0.07 \text{ Sv}$ of inflow and $0.01 \pm 0.02 \text{ Sv}$ ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) of outflow. This estimate comes from Candela et al. (2019, their Fig. 3a), and was obtained by best matching the total 4-yr mean transport at the Yucatan Channel and the Florida Strait sections. The residence time, or time that it takes for the inflow to refill such volume, is on the order of 100 ± 70 years, and, accordingly, an average vertical velocity of $1.6 \pm 1.7 \text{ cm day}^{-1}$. Another valid estimate of the GOM's residence time can be calculated from Candela et al. (2019, their Fig. 3c), which was obtained by considering the mean from all the measured current series, within the observed 4-yr period, that

had at least 1-yr-long record. In this case the mean inflow below 1800 m gives 0.22 Sv with a corresponding deep water ($>2700 \text{ m}$) residence time of 45 years. Both the inflow and the residence time are within the estimated uncertainty limits mentioned before. These are crude, but databased, estimates of the residence time, which agree with radiocarbon based estimates for the GOM (Amon et al. 2020, manuscript submitted to *Nat. Commun.*).

Given the limited knowledge on the velocity and mixing within the gulf, only crude estimates of the mechanisms that cause the $\sim 18 \text{ m}^\circ\text{C decade}^{-1}$ from the bottom to $\sim 2000 \text{ m}$ below the surface are possible; some assumptions that, in our view are deemed reasonable, cannot be verified with present-day knowledge. Nonetheless, the evidence for warming is quite similar to the study of MacCready et al. (1999), which shows, for the Caribbean Sea, a warming of the deep Venezuela Basin waters consistent with dense episodic spills into the basin through the Anegada–Jungfern passage. These episodes rarely involve waters as dense as those found in the deepest bottom of the basin, but remain the only plausible source of the deep waters. In the GOM, the inflow of dense waters through Yucatan Channel is more continuous and the only plausible source of the densest waters found in the GOM's interior. In contrast, in the Venezuela Basin the interleaving of a thick intermediate layer seems to be the dominant mechanism for the warming (MacCready et al. 1999).

A simplified model proposed by Joyce et al. (1999) assumes that the change in source waters that flow into the Venezuela Basin reflect the nearly uniform warming observed at depth. The model is set by

FIG. 7. Schematic of the model. Notice the isopycnals deepening, the high intensity of mixing on the downslope transit of the waters, the upwelling along other sloping areas of the BBL, and the interleaving communicating the BBL with the interior.

$$\tau \frac{\partial \bar{\Theta}}{\partial t} + \bar{\Theta} = \Theta_s, \quad (3)$$

where τ is the residence (or flushing) time, $\bar{\Theta}$ is the mean temperature in the deep interior or control volume, and Θ_s is the temperature of the source waters, at the entrance to the basin. The residence time is set by $\tau = \text{Vol}/\text{Tr}_0$, where Vol is the volume of the deep interior and Tr_0 is the transport into the same volume. Joyce et al. (1999) find $\tau \approx 215$ years for the model to best fit the observations in the Venezuela basin. Since the control volume is time invariant, the outflow transport equals the inflow transport. This balance should be seen applicable for long-term means as the heave of isopycnals and isotherms on the time scales of months and seasons is substantial. Chang and Oey (2011) describe plausible slow motions within the GOM that are considered short-term turbulent fluctuations in the light of this equation. Equation (3) represents a single-pole autoregressive process that, in the presence of a steady trend in the temperature of incoming waters, implies the same trend is approached asymptotically in the deep interior or control volume. A conceptual sketch of the spilling model (Fig. 7) is similar to figures found in the schematics in Bryden and Nurser (2003), Speer and Tziperman (1990), or Amon et al. (2020, manuscript submitted to *Nat. Commun.*).

In the specific case of the GOM, Eq. (3) needs to be modified by adding a heat source to explain the fact that the temperatures in the deep GOM are higher than those at the Yucatan Sill source. Considering this, the model can be reformulated as

$$\frac{\partial \bar{\Theta}}{\partial t} = \frac{\Theta_s - \bar{\Theta}}{\tau} + G, \quad (4)$$

where G represents a heat source (in our case turbulent plus geothermal). In the steady state, we would have $\bar{\Theta} = \Theta_s + \tau G$. So, as long as G is positive and sufficiently large, the trend of $\bar{\Theta}$ can be positive (and $\bar{\Theta}$ be larger than Θ_s), even if $(\Theta_s - \bar{\Theta})/\tau$ is negative, as is the case here. Now that we have a model that is consistent with the system, we can investigate the behavior of

the trends. For simplicity, let us assume that G is constant, and in that case

$$\frac{\partial^2 \bar{\Theta}}{\partial t^2} + \frac{1}{\tau} \frac{\partial \bar{\Theta}}{\partial t} = \frac{1}{\tau} \frac{\partial \Theta_s}{\partial t}. \quad (5)$$

Assuming that $\partial \Theta_s / \partial t = \Delta$ is a constant and the initial conditions are $\bar{\Theta}(0) = \bar{\Theta}_0$ and $\partial \bar{\Theta} / \partial t|_{t=0} = 0$, where $\bar{\Theta}_0$ is the initial temperature before warming started and there is no linear trend at $t = 0$. The solution to (5) with the “forcing” Δ and the given initial conditions is

$$\bar{\Theta}(t) = \bar{\Theta}_0 + \Delta \tau e^{-t/\tau} + \Delta(t - \tau), \quad (6)$$

which implies that the linear trend Δt of the source waters will not be transmitted to the deep GOM until the second term on the left is almost zero. Seeing how $\partial \bar{\Theta} / \partial t$ behaves as a function of the residence time τ (Fig. 8) is illustrative because this simple equation indicates that the longer the basin’s residence time τ is, the longer the warming trend in the source waters had to have existed. An important problem that arises in applying this simple model is that we do not have a reasonable estimate of the trend of the source waters at the Yucatan sill, in addition of the large uncertainties on the deep GOM residence time. This points to the fact that we first need to estimate the warming trends in the Cayman–Yucatan Basin.

The densest inflowing waters, with the warming trend as stated by Joyce et al. (1999) for ~ 2000 m ($30 \text{ m}^\circ\text{C decade}^{-1}$), enter the Cayman–Yucatan through deep inflow over the Windward Passage sill (Smith 2010). The Cayman–Yucatan Basin has a volume of $\sim 8.7 \times 10^5 \text{ km}^3$ below 2000 m and direct measurements suggest an upper limit estimate of the deep inflow over the Windward Passage sill is $\sim 0.4 \text{ Sv}$ (Johns et al. 2007; Smith 2010), yielding a residence time for this volume of at least 70 years. This implies that a warming signal inflowing at the Windward Passage sill would take on the order of 70 years to properly manifest itself at the Yucatan Channel sill, but more importantly, it would be a time-dependent forcing for the

FIG. 8. Time derivative of solution (6), $\partial\bar{\Theta}/\partial t = \Delta(1 - e^{-t/\tau})$ for different values of residence time τ as indicated. Here, $\Delta = 3 \text{ m}^\circ\text{C year}^{-1}$ taken from Joyce et al. (1999) for waters at $\sim 2000 \text{ m}$ in the western North Atlantic. The green line is the time derivative of solution (8), which considers a time-dependent forcing at the GOM's entrance, $\partial\bar{\Theta}/\partial t = \Delta_s[1 + (\tau_s e^{(-t/\tau_s)} - \tau e^{(-t/\tau)})/(\tau - \tau_s)]$, with $\Delta_s = 3 \text{ m}^\circ\text{C yr}^{-1}$; $\tau_s = 70$ years, the residence time of the Cayman–Yucatan Basin; and $\tau = 45$ years, a lower limit for the residence time of the deep waters ($< 2700 \text{ m}$) in the GOM. The horizontal line marks the $16 \text{ m}^\circ\text{C decade}^{-1}$ present warming trend of GOM deep waters.

GOM, as Fig. 8 indicates. So instead of solving Eq. (5) with a constant right-hand side, we need to solve

$$\frac{\partial^2 \bar{\Theta}}{\partial t^2} + \frac{1}{\tau} \frac{\partial \bar{\Theta}}{\partial t} = \frac{\Delta_s}{\tau} (1 - e^{-t/\tau_s}), \quad (7)$$

where Δ_s is the constant warming trend at the Windward Passage entrance, τ is the residence time of the GOM, and τ_s the residence time of the Cayman–Yucatan Basin. The solution to this equation subject to the same initial conditions as before, namely, $\bar{\Theta}(0) = \bar{\Theta}_0$ and $\partial\bar{\Theta}/\partial t|_{t=0} = 0$ is

$$\bar{\Theta}(t) = \bar{\Theta}_0 + \Delta_s(t - \tau_s - \tau) + \frac{\Delta_s}{\tau - \tau_s} (\tau^2 e^{-t/\tau} - \tau_s^2 e^{-t/\tau_s}). \quad (8)$$

The time derivative of this solution suggests that, for the GOM deep waters to attain their present $16 \text{ m}^\circ\text{C decade}^{-1}$ warming trend, the waters of the western North Atlantic at the entrance to Windward Passage should have started warming about 100 years ago (Fig. 8, green curve). Given the crude approximations used to obtain this number, these 100 years should be taken as the lower-limit estimate of when the Atlantic Ocean started warming. Finally, we note that the derivative of the above solution will be equal to the derivative of the solution of (4) if $G = (\bar{\Theta}_0 - \bar{\Theta}_{s0})/\tau$, which is the constant warming needed to maintain the temperature difference between the deep GOM and the source waters at $t = 0$ (i.e., before the warming started).

What this analysis implies is that the inflow of warmer deep waters and their entrainment during mixing are key ingredients for the observed warming in the deep GOM. However, it could be argued that the warming of the source waters does not necessarily have to be directly reflected in the overall warming pattern in the abyssal gulf. For example, changes in the circulation could result in changes in the depths of isotherms near the sill, or changes in the T – S relation could also impact the temperature of the source waters. Nevertheless, given the evidence of the long-term warming of the deep waters in the Atlantic (e.g., Joyce and Robbins 1996; Joyce et al. 1999), we consider it the most important contributor to observed temperature trend in the deep gulf.

The slow warming of deep waters and the faster warming above 1000 m imply that the GOM large-scale stratification is increasing, similar to observations for global oceanic conditions. Vertical diffusion of heat, from warmer waters above is not supported by our observations as a significant factor in the warming (see Fig. 6 and requirement of excessive diffusivity in Munk's model fit). In fact, the increase in stratification is more likely to reduce the vertical turbulent exchanges.

4. Conclusions

Two independent time series observations, CTD profiles taken from 2003 to 2019 and moored thermometers, deployed from 2008 to 2019 in the western GOM, show that

the overall temperature trend in the abyssal depths of the Gulf of Mexico is $\sim 18 \text{ m}^\circ\text{C decade}^{-1}$. The low uncertainty of $\pm 2 \text{ m}^\circ\text{C decade}^{-1}$, at 90% confidence level, is due to the low variability in temperature measurements at such depths and, for the moored data, the appreciable length of the available measurements. This heating is, as shown by the CTD profiles, uniform from $\sim 2000 \text{ m}$ to the bottom at $\sim 3700 \text{ m}$ in the abyssal plains. In particular, for the moored observations, the time series show a steady increase in the last decade without any sign of a hiatus or deviations from a linear trend.

The geothermal heat flux is an essential contributing factor in the observed temperature distribution; it induces convection at the base of the BBL and therefore is a factor for the renewal of waters. Nonetheless, it cannot explain by itself the warming trend, as such flux has existed for millenniums and is too weak to account for the warming trend in a layer as thick as observed in the deep GOM ($>2000 \text{ m}$).

The evidence presented in this paper emphasizes the critical role of mixing and entrainment of overflow waters that ultimately fills the deep regions of the GOM and contributes to a relatively homogeneous warming of the GOM below 2000 m .

An important consideration is the fact that the warming trend measured by the deep thermometers in the GOM is a statistically reliable and solid measurement that adds key information for understanding the warming behavior of the intermediate waters in the Atlantic Ocean. This points to the importance of long-term measurements in marginal seas that, in combination with simple models, can be used to investigate the warming history of the broad oceans connected to them.

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We would like to express our immense grief for the loss of our two great friends and colleagues José Ochoa and Vicente Ferreira-Bartrina. This work belongs to them; both were excellent scientists and their motivation will be with us forever.

FIG. A1. (a) The relationship from instrument output (i.e., counts) to temperatures as determined by the 19 May 2008 calibration. (b),(c) Considering the post-cruise calibration performed on 17 Oct 2017, the distribution of δT_0 as a function of the instrument output and temperature, respectively.

APPENDIX A

Instrumental Drift

The manufacturer of CTDs (SBE 911plus CTD) and the moored temperature sensors (SBE 37 MicroCAT), Sea-Bird Scientific, states (Sea-Bird Electronics Inc. 2016) that a drift of magnitude larger than $2 \text{ m}^\circ\text{C decade}^{-1}$, either increasing or decreasing, is unusual (Sea-Bird Electronics Inc. 2016). The straightforward assumption to correct for drift is to apply a linear correction in time from one calibration to the following so-called “post-cruise” calibration. The temperature computed with the calibration constants (i.e., the a values supplied by Sea-Bird Scientific) is

$$T = \frac{1}{[a_0 + a_1 \ln(n) + a_2 \ln^2(n) + a_3 \ln^3(n) + a_4 \ln^4(n)] - 273.15}, \quad (\text{A1})$$

where T is the ITS-90 temperature and n is the instrument output or counts which is the actual value the instrument records, say at time (date) equal t_0 can be corrected with the post-cruise calibration performed at t_1 . Let δT_0 be the additive correction for temperature T_0 computed with the post-cruise calibration constants. The instrument output that in the calibration at t_0 implied T_0 , say n_0 , produces with the post-cruise calibration constants T_1 . If immersed in a bath of temperature T_0 at t_1 , the use of the calibration pertinent of t_0 implies T_1 (i.e., $\delta T_0 \equiv T_0 - T_1$). The assumption of a uniform drift between both calibration dates is stated by

$$T_C(t) = T_U(t) + \delta T_0 \times (t - t_0)/(t_1 - t_0), \quad (\text{A2})$$

where T_U is the temperature (uncorrected) computed with the calibration constants corresponding to t_0 and T_C is the corrected (true) temperature for intermediate dates between t_0

TABLE A1. A set of seven SBE37 with calibration and post-cruise calibration done by the manufacturer (Sea-Bird Scientific) showing the instrumental drift in temperatures. The table shows the serial number (SN) the instrumental drift at the reference temperature of 4.5°C, the dates of calibrations, and the time span between such calibrations (years).

SN	Drift (m°C decade ⁻¹)	Original calibration date	Post-cruise calibration date	Years
5615	−3.2	20 Feb 2014	11 Oct 2017	3.64
5617	−0.6	21 Oct 2007	15 Dec 2017	10.15
6167	−2.0	18 May 2008	8 Dec 2017	9.56
6170	−2.6	21 May 2008	20 Feb 2014	5.75
6177	−1.1	19 May 2008	11 Oct 2017	9.40
6179	−1.3	18 May 2008	18 Oct 2017	9.42
6775	−1.5	9 Apr 2009	8 Dec 2017	8.67

and t_1 . For the instrument placed at t_1 in a bath of temperature T_0 the instrument output is, due to nonnull drift, no longer n_0 and Eq. (A2) has $T_U(t_1) = T_1$, hence producing $T_C(t_1) = T_0$. That the drift $\delta T_0/(t_1 - t_0)$ applies in the neighborhood of T_0 is the basic assumption for instrumental drift correction. Notice that Eq. (A2) has been stated for the neighborhood of T_0 and even for differing calibration constants, it is possible that $\delta T_0 = 0$ (i.e., the intersection of two fourth-degree polynomials is not ruled out). Figure A1 shows δT_0 for one SBE37 as a function of the instrument output (i.e., counts) and, since δT_0 is far too small compared with T_0 , it can be posed without loss of meaning as a function of T_0 instead of as a function of the instrument output. The polynomial fit is carried with the seven reference temperatures depicted by circles in Fig. A1 (i.e., $\sim 1.0^\circ$, 4.5° , 15.0° , 18.5° , 24.0° , 29.0° , and 32.5° C) and seven of the residuals of the fit are on the order of 0.1 m°C, hence the same precision is assumed within the temperature range proper of the instrument. Of primary interest for this study is the drift [i.e., $\delta T_0/(t_1 - t_0)$] for the reference temperature $T_0 \approx 4.5^\circ$ C, as that is close to the moored temperature series shown in Fig. 5. For the example shown in Fig. A1 the drift at 4.5° C is -1.1 m°C decade⁻¹, which is negative and quite small in amplitude. Notice from Fig. A1c that the drift at $\sim 26^\circ$ C is null and for higher temperatures is positive although still quite small in amplitude. Table A1 shows the calculation of drift [i.e., of $\delta T_0/(t_1 - t_0)$], as just described for seven SBE37s for which the post-cruise calibration is available. The low amplitude of such instruments' drift is noticeable. Note that all these drifts are negative, and the correction is not needed given that is far too small compared with the confidence interval that the environment trends show.

These estimates are larger than those of Uchida et al. (2008), which report mean drifts of ~ 0.1 m°C decade⁻¹ in same instrument model. In our measurements, the drifts remain tolerable, as without drift corrections the statistical uncertainty of the warming trend is of the same order.

APPENDIX B

Least Squares Fit and Confidence Intervals via Monte Carlo Simulations

The ad hoc least squares fit used to estimate the temperature trend is straightforward in the sense that it minimizes the sum of squares of the set of differences between an analytical

function and the data. The analytical function has some free parameters, and for any choice of such parameters the sum of squares can be computed. The common and practical procedure is to build a function that, given the set of free parameters, computes the residuals (i.e., the differences between the data and the analytical function) and their sum of squares. Hence, minimizing such a sum of squares, via iterations as the parameters vary, allows finding the least squares result sought. We use the function “fminsearch” from MATLAB to solve for this minimum, which is based in the Nelder–Mead simplex method as described for the “amoeba algorithm” in Press et al. (2007).

The analytical function of choice makes the least squares fit ad hoc; it consists of the addition of two functions, one that is the addition of rectangular functions that modifies the mean temperature of each “deployment to recovery period” (i.e., continuously measured by one instrument) plus a linear contribution. The slope of the linear contribution is the main interest of this study. The fit does not require “regularization” because the discontinuous function, defined by the set of partial means, is restricted to conserve the overall series mean. This discontinuous contribution depends on $n - 1$ free parameters where n is the number of available deployment to recovery periods. This choice does not eliminate the overall

FIG. B1. The graph shows the fitted analytical function (i.e., the discontinuous function with piecewise linear segments of identical slope) and residuals in a different perspective than Fig. 5 frame LMP. The dashed line along the fitted function is the linear contribution once the discontinuities have been removed. The lower irregular trace is the series of residuals with the zero as a dashed horizontal line surrounded by plus and minus the standard deviation of the residuals.

FIG. B2. The graph shows the empirical residuals in the uppermost trace for the location LMP, as is shown in Fig. B1, and with a uniform offset of 5 m°C for four simulations of residual series for the Monte Carlo simulations.

series mean as one of the free parameters, which, in addition to the slope of the trend, makes the least squares problem of $n + 1$ free parameters.

The mean temperature registered in all deployments is $\langle T \rangle = (1/J) \sum_{j=1}^J T(t_j)$, where J is the total number of samples, regardless of deployments. It must be noted that the sampling at $t_1, t_2, \dots, t_j$ is not equally spaced, since there are gaps between the deployments. Consider the notation of $t_{K1}, t_{K2}, \dots, t_{KN_K}$ for the, 1-yr, equally spaced, times of successful temperature measurements, within the K th deployment. For example, for location LMP there are nine successful mooring deployments with corresponding recoveries, then, K runs from 1 to 9. Let n be the largest K (i.e., $K \in \{1, 2, \dots, n\}$, see second column in Table 2), hence $J = \sum_{K=1}^n N_K$ and $t_1, t_2, \dots, t_j = t_{11}, t_{12}, \dots, t_{1N_1}, t_{21}, t_{22}, \dots, t_{2N_2}, \dots, t_{n1}, t_{n2}, \dots, t_{nN_n}$. The lapse and number of hourly samples of each deployment to recovery time series vary, for example in LMP the shortest is the first deployment with $N_1 = 7039$, the longest is the eighth deployment with $N_8 = 13\,294$, and the last deployment has $N_n = N_9 = 10\,981$ samples. For LMP $J = 85\,757$, and the eight gaps vary from 7 to 15 h. Let $H_K(t) = 1$ if $t \in [t_{K1}, t_{KN_K}]$ and zero otherwise, a rectangular function, and $\bar{T}_K$ be the average of temperatures at times $t_{K1}, t_{K2}, \dots, t_{KN_K}$ (i.e., the temperature average for deployment K), therefore $N_K \bar{T}_K = \sum_{i=1}^{N_K} T(t_{Ki}) = \sum_{j=1}^J T(t_j) H_K(t_j)$, and $\langle T \rangle = (1/J) \sum_{K=1}^n N_K \bar{T}_K$. The equation

$$T(t) = T_O + \frac{DT}{Dt} t + \sum_{k=1}^n (\hat{T}_K - \bar{T}_K) 0 H_K(t) + \delta T(t), \quad (\text{B1})$$

where $T(t)$ is the temperature measured at time t , T_O is a constant, DT/Dt is the temperature trend, and $\{\hat{T}_K | K = 1, 2, \dots, n\}$ is a set of “updated means” (i.e., one per deployment), helps setting the least squares problem. The purpose of introducing the updated means is the correction of mean shifts due to thermometer location change, as described by Venema et al. (2018). The green traces in Fig. 5 show the temperature series corrected of this mean-shift problem [i.e., $T(t) - \sum_{k=1}^n (\hat{T}_K - \bar{T}_K) H_K(t)$]. To spread and minimize the correction in the full set of thermometers used, we add the restriction $\langle T \rangle = (1/J) \sum_{K=1}^n N_K \hat{T}_K$ on the updated means, which implies $\sum_{K=1}^n N_K (\hat{T}_K - \bar{T}_K) = 0$.

TABLE B1. The list presents, as an example, the fitting of a few parameters for the Monte Carlo simulations. The first three columns show the fitted mean temperature minus the observed mean for deployments from first to third (see text about the series discontinuities and free parameters in the ad hoc least squares fit and Fig. B1). The first row (highlighted in bold) shows the values whose confidence interval are to be found, which are shown at the 90% limit in the last row (also in bold), as deduced from 500 simulations.

$\delta T \equiv \langle T \rangle_{\text{updated}} - \langle T \rangle_{\text{original}} \text{ (m}^\circ\text{C)}$			Std (residuals) (m°C)	DT/dt (m°C decade ⁻¹)
-2.42	0.39	1.04	0.68	14.6
-2.40	0.40	1.05	0.63	14.6
-0.62	1.82	2.05	0.60	10.6
-2.48	0.35	1.02	0.67	14.7
-1.52	1.12	1.54	0.59	12.9
±1.31	±1.04	±0.74	±0.05	±2.2

The least squares problem is defined as the minimization of $F(T_O, DT/Dt, \hat{T}_1, \hat{T}_2, \dots, \hat{T}_{n-1}, T) = \langle (\delta T)^2 \rangle \equiv (1/J) \sum_{j=1}^J (\delta T)^2$, because given $\hat{T}_1, \hat{T}_2, \dots, \hat{T}_{n-1}$ sets the value of T_n , i.e., there are $n + 1$ free parameters. As previously stated the solution is sought numerically, via iterations. As the free parameters vary, the set of calculation of $\langle (\delta T)^2 \rangle$ allows finding its minimum [see, e.g., the “amoeba algorithm” in Press et al. (2007)].

The common algorithm for estimating confidence intervals for the slope of the linear contribution does not apply. In this ad hoc fitting, (i) there is the presence of other nonstandard parameters and (ii) the residuals do show significant deviation from a Gaussian distribution. Figure B1 shows the residuals for the data shown in Fig. 5 LMP frame.

For Monte Carlo simulations, it is fundamental to build series with the same statistics as the residuals, referred as synthetic residuals. It is possible to add synthetic residuals to the fitted function from the data and compute the resulting free parameters with the same algorithm. These last parameters differ from the estimated data-derived parameters, precisely because the residuals are not the same. The synthetic residuals are assumed to be a plausible outcome as implied by the data themselves. A sufficiently large set of statistically independent realizations with synthetic residuals, allows for the calculation of confidence intervals for any of the free parameters.

The two functions that are basic in the statistics of the residuals are the probability density function (pdf) and the power spectral function (psd). Alternative equivalences of pdf and psd are the cumulative probability function (cpf) and the lagged autocovariance function. Further statistical characteristics will not be pursued to be estimated from the data or reproduced by the synthetic residuals. We follow Sotiriadi (2014) procedure to build synthetic residuals series with the psd and pdf dictated by the residuals. The procedure, which is based on Price’s theorem (Price 1958), is only an approximation in this case because the residuals have considerable skewness and Sotiriadi’s procedure is formally for pdfs without skewness. The skewness of the simulations is matched completely to those implied by the data residuals, but the psd is distorted.

Figure B2 shows several examples of synthetic residuals, and Table B1 shows some of the set of fitted parameters in those examples. The relevant quantity to figure out via 500 Monte

Carlo simulations is the expected error (under the same residuals statistics) for the temperature trend, as shown in the example in Table B1 for location LMP and just a few parameters. In 500 simulations, sorting the fitted parameters, the 90% confidence interval follows from the 25 and 475 fitted values. The 90% confidence limits in the estimate of the temperature trend at each location are shown in labels in Fig. 5.

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