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CRA-LICOM: A global high-frequency atmospheric and oceanic temporal gravity field product (2002-2024)

Fan Yang[2,3], Jiahui Bai[1,9], Hailong Liu[1,4], Weihang Zhang[3], Yi Wu[3], Shuhao Liu[3], Chunxiang Shi[5], Tao Zhang[5], Min Zhong[6], Zitong Zhu[7,8], Changqing Wang[7], Ehsan Forootan[2], Jiangfeng Yu[1,9], Zipeng Yu[1], and Yun Xiao[10]

[1]Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing 100029, China
[2]Geodesy Group, Department of Sustainability and Planning, Aalborg University, Aalborg 9000, Denmark
[3]School of Physics, Huazhong University of Science and Technology, Wuhan 430074, China
[4]Laoshan Laboratory, Qingdao 266237, China
[5]National Meteorological Information Center, China Meteorological Administration (CMA), Beijing 100081, China
[6]School of Geospatial Engineering and Science, Sun Yat-sen University, Zhuhai 519082, China
[7]State Key Laboratory of Precision Geodesy, Innovation Academy for Precision Measurement Science and Technology, CAS, Wuhan 430077, China
[8]Max-Planck-Institut für Gravitationsphysik (Albert-Einstein-Institut) and Institut für Gravitationsphysik, Leibniz Universität Hannover, Hannover 30167, Germany
[9]College of Earth and Planetary Sciences, University of Chinese Academy of Sciences, Beijing 100049, China
[10]Xi'an Research Institute of Surveying and Mapping, Xi'an 710054, China

Correspondence: Jiahui Bai (baijh5@mail2.sysu.edu.cn)

Abstract. Modeling sub-daily mass changes, dominated by the atmosphere and the oceans, is not only essential for understanding weather and climate change but also serves as a fundamental requirement for nearly all existing terrestrial or space-borne geodetic observations to perform signal separation. Removing these high-frequency mass changes, through the usage of so-called de-aliasing products, is of particular interest for satellite gravity missions such as GRACE and GRACE-FO to prevent the aliasing of short-term mass changes into seasonal and long-term mass variability. However, establishing a global observation network to monitor high-frequency gravity signals is impractical. Thus, ongoing efforts focus on simulating this high-frequency signal by driving atmospheric/oceanic numerical models with specific climate-forcing fields and assimilating observational data. Its realization relies on a complicated system and the uncertainty of obtained results is non-negligible for its dependency on selected forcing field and ocean model.

To explore the signal and uncertainty of de-aliasing products, we establish China's first de-aliasing computation platform, independently. This is achieved by using the recently released CRA-40 (China's first generation of atmospheric reanalysis) as forcing fields to drive our in-house 3-D atmospheric integration model and the LASG/IAP (State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics/Institute of Atmospheric Physics) Climate System Ocean Model 3.0 (LICOM3.0). With this new platform, we reproduce an alternative high-frequency atmospheric and oceanic gravity de-aliasing product, called CRA-LICOM, at 6 hourly and 50 km resolution, covering 2002-2024 at a global scale. The product is freely available at https://doi.org/10.11888/SolidEar.tpdc.302016. Inter-comparisons with the products of GFZ (Deutsches GeoForschungsZentrum) and validations against independent observations have revealed: (i) the current version

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of CRA-LICOM has well satisfied the requirement of the state-of-the-art satellite gravity missions, as well as other geodetic measurements, and (ii) despite agreement across most areas, considerable uncertainty is found at marginal seas near continental shelves, particularly at high-latitude regions. Therefore, scientific applications that aim to understand the fast-changing global water cycle, as well as mission design of future satellite gravity that seeks accurate gravity de-aliasing, can use our product as a reliable source. The current platform has the potential to be improved in terms of modeling and data assimilation capacity, which will be outlined in this study.

1 Introduction

Earth's temporal gravity field reflects equivalent water changes from various sources such as the atmosphere (A), ocean (O), hydrology (H), ice sheets (I), and solid Earth (S). Accurate disaggregation of the temporal gravity field into these sources is crucial to understanding the natural evolution of each process on and beneath the Earth (Wahr et al., 1998; Tapley et al., 2019). For example, Terrestrial Water Storage (TWS, associated with the H component) is considered an essential climate variable to diagnose the internal variability of the global water cycle and climate change (Rodell et al., 2018; Rodell and Reager, 2023).

In particular, state-of-the-art geodetic observations from, e.g., terrestrial/space-borne gravity (see Güntner et al., 2017) and GNSS (Global Navigation Satellite System, see White et al., 2022; Klos et al., 2023), often represent a mixture of these sources, that is, AOHIS (A + O + H + I + S), where separation is required to obtain desired components, such as TWS or HIS (H+I+S). Generally speaking, a reduction of AO (A+O) from the total signal is feasible because it is dominated by high-frequency changes, while TWS or HIS often associate with relatively slower gravity changes (Bai et al., 2024). Therefore, precise AO modeling is not only essential for understanding rapid climate changes but is also relevant as an a priori model to separate TWS or HIS from other signals.

In addition, the AO model is vital for the Gravity Recovery and Climate Experiment mission (GRACE, 2002-2017; Tapley et al., 2004) and its follow-on mission (GRACE-FO, 2018-present; Landerer et al., 2020), which provides monthly snapshots of HIS changes globally with unprecedented precision (Velicogna and Wahr, 2006; Scanlon et al., 2018; Chao and Liau, 2019). However, accurate acquisition of HIS components depends on reliable AO prior models to reduce aliasing errors (Wahr et al., 1998). Such errors significantly degrade HIS estimations because sub-daily AO variability is much below the feasible temporal resolution of GRACE (e.g., Han et al., 2004; Forootan et al., 2014). These aliasing errors are among the largest error sources in current space-borne gravity missions and may restrict next-generation missions (Han et al., 2007; Seo et al., 2008; Liu and Sneeuw, 2021; Chen et al., 2022), despite improved onboard instruments (Flechtner et al., 2016; Zhou et al., 2021), unless faster sampling strategies or co-estimations of AO parameters are applied (Kurtenbach et al., 2009; Wiese et al., 2011; Mayer-Gürr et al., 2012; Daras and Pail, 2017; Hauk and Pail, 2018; Mayer-Gürr et al., 2018; Purkhauser and Pail, 2019). In addition to GRACE(-FO), AO modeling is relevant to other geodetic techniques. For example, it improves the determination of the satellite altimetry orbit (Cerri et al., 2010; Rudenko et al., 2016; Bonin and Save, 2020) and is a mandatory post-processing step for terrestrial gravity measurements (Boy et al., 2002, 2009) and GNSS station displacement measurements (Dill and

Dobslaw, 2013; Han and Razeghi, 2017; Swarr et al., 2024). Consequently, efforts to achieve precise AO modeling remain ongoing within the geodesy community.

Generally, the AO model consists of tidal and non-tidal constituents, whereas we shall use the term AO to mainly indicate the non-tidal high-frequency (sub-daily) aspect from now on since the tidal modeling is another important but less related issue, particularly for the oceanic component. Current AO models often use climate forcing fields, followed by atmospheric gravity calculation through vertical integration of air mass, and oceanic gravity simulation through ocean circulation models (Wahr et al., 1998). So far, the only publicly available AO model that is kept up-to-date is maintained by GFZ, which has long been relied upon to produce monthly gravity fields by major GRACE(-FO) data processing centers worldwide (Dobslaw et al., 2017; Shihora et al., 2022). Their product has evolved significantly over the past two decades, focusing on improving atmospheric forcing fields (Duan et al., 2012; Hardy et al., 2017; Yang et al., 2021), refining atmospheric integration (Swenson and Wahr, 2002; Boy and Chao, 2005; Zenner et al., 2010; Forootan et al., 2013; Dobslaw et al., 2017), and switching forced ocean models (Bonin and Save, 2020; Schindelegger et al., 2021; Shihora et al., 2022). Due to these efforts, their latest product, AOD1B-RL07 (called GFZ-RL07 hereinafter to avoid confusion), has reached a high-quality level. However, as addressed by Shihora et al. (2022), GFZ-RL07 is inevitably imperfect in capturing the high-frequency variability, particularly the oceanic component, since it is a purely atmospherically forced oceanic simulation without constraints from observations.

Recognizing that there is still a considerable error in the AO model, it would be beneficial to increase the diversity of the AO model to better understand its uncertainty for further improvement (Springer et al., 2024), rather than having GFZ-RL07 as the only option, and this also builds the motivation of this work. In fact, GFZ-RL07 has long relied on atmospheric operational data or reanalysis from ECMWF (European Center for Medium-Range Weather Forecasts) as forcing data, and another available AO model produced by Gegout (2020) also relies on ECMWF data, and unfortunately has stopped updating from 2017. In this context, developing another AO model independent of the GFZ-RL07 model should expect to apply a completely different atmosphere forcing and oceanic model. In November 2013, the China Meteorological Administration (CMA) launched the global reanalysis project, and after ten years of effort, China's first generation global atmospheric and land reanalysis product (named CRA-40) became publicly available (Liu et al., 2023). Intensive evaluations of CRA-40 (see Shen et al., 2022; Liu et al., 2023) have shown a better performance than the existing global reanalysis products to the latest ECMWF reanalysis, particularly in terms of surface pressure, temperature, and specific humidity, etc., which are exactly the key variables used to establish the AO model. In addition to the new forcing dataset, we also introduce LICOM3.0 in-house, an advanced and effective ocean model among the best peer models in the world (Lin et al., 2020), to simulate oceanic variables, including ocean bottom pressure (OBP) that reflects the oceanic mass/gravity change (Liu et al., 2012).

In this study, due to the release of CRA-40, in conjunction with the ocean model LICOM, it is possible for us to develop an up-to-date global high-frequency atmospheric and oceanic gravity product, named CRA-LICOM (2002-present), which is completely independent of GFZ-RL07. We anticipate that this alternative could diversify the gravity recovery options from GRACE(-FO), and provide an opportunity to access the AO full time-scale uncertainty via an inter-comparison between these two independent products (Shihora et al., 2024). It was revealed by Kvas and Mayer-Gürr (2019) that accounting for the

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AO uncertainty information in GRACE(-FO)'s gravity recovery would considerably enhance the quality, and this strategy is suggested as the standard processing chain for official producers as well.

In this paper, we first introduce the input data sets for both modeling and validation in Sect. 2. Subsequently, a brief description of the atmospheric/oceanic gravity modeling methodology is addressed in Sect. 3. Then, we demonstrate the main output of the CRA-LICOM products in Sect. 4 and evaluate their performance with independent observations in Sect. 5. Finally, we analyze the limitations of the current release of the CRA-LICOM product in Sect. 6, discuss the conclusions, and outline the way forward in Sect. 7.

2 Input dataset description

2.1 Modeling dataset

China's first-generation global atmospheric and land reanalysis, CRA-40, is chosen herein as the climate-forcing field. It applies to the National Centers for Environmental Prediction (NCEP) Global Spectral Model (GSM)/Gridpoint Statistical Interpolation (GSI) 3D-Var system at a 6-hour time interval with 64 vertical levels and a horizontal resolution of 34 km. The original model-level output is post-processed into 47 pressure-level products, and then all variables are interpolated to four horizontal resolutions in longitude–latitude projection, including 0.25°, 0.5°, 1° and 2.5°. CRA-40 can be accessed via http://data.cma.cn/CRA, where, for our study, the dataset covering 2000-2024 is extracted. To compromise accuracy and computational efficiency, the spatial/temporal resolution, i.e., 0.5°/6-h of CRA-40, is selected for all variables required in this study. A higher spatial resolution, such as 0.25°, is not considered currently since GRACE's resolution is much coarser, e.g., 3° (Landerer and Swenson, 2012). Specifically, four variables are required to facilitate the atmospheric gravity field modeling, which are the surface pressure, the surface geopotential, the multi-layered temperature (pressure level), and the multi-layered specific humidity (pressure level). On the other hand, 11 variables are required to force the LICOM3.0 model, which are air density, temperature, zonal wind speed, meridional wind speed, specific humidity at 10 m, sea surface pressure, runoff, precipitation, downward long-wave radiation flux, downward shortwave radiation flux, and upward shortwave radiation flux.

2.2 Validation dataset

2.2.1 GFZ-RL07 AO model

The GFZ-RL07 AO model is the official de-aliasing product for all existing satellite gravity missions. It provides non-tidal atmospheric and oceanic components with 3-hour temporal resolution and spherical harmonic expansion up to degree/order 180 alongside selected tidal constituents slower than 6 hours. GFZ-RL07 is accessible via https://isdc.gfz-potsdam.de/esmdata/aod1b/ and is used here for comparison with CRA-LICOM.

2.2.2 GRACE Level-1b and Level-2 data

Temporal gravity field (Level-2) using GRACE Level-1b products can be used to confirm the accuracy of CRA-LICOM for current satellite gravity missions. GRACE Level-1b products, including along-track range(-rate), accelerometer, star camera attitude, and reduced dynamic orbit data, are available at https://podaac.jpl.nasa.gov/dataset/GRACE_L1B_GRAV_JPL_RL03. Additionally, the latest version (RL06) of GRACE Level-2 temporal gravity fields (in terms of spherical harmonic coefficient) from CSR (Center for Space Research from the University of Texas at Austin, Texas, USA), JPL (Jet Propulsion Laboratory, USA), and GFZ are used for further validation, accessible via https://icgem.gfz-potsdam.de/home.

2.2.3 Altimeter and Argo

Altimeter and Argo are used to obtain manometric sea level, which can validate GRACE Level-2 gravity fields as well as its underlying AO model, i.e., CRA-LICOM (Chen et al., 2018; Gregory et al., 2019). For this study, the mean monthly changes in the steric ocean are derived from Argo products BOA (Li et al., 2017), EN4 (Good et al., 2013) and SIO (Roemmich and Gilson, 2009), which cover the upper ocean above 2,000 meters. Altimeter data from AVISO at a resolution of 0.25°×0.25° and 10-day intervals are also used, calibrated for GIA (Glacier Isostatic Adjustment, see Caron et al. (2018)) effects as suggested.

2.2.4 OBP Recorders and related variables from Argo

Ocean Bottom Pressure (OBP) data from the Deep Ocean Assessment and Reporting of Tsunamis (DART) system (National Oceanic and Atmospheric Administration, 2005) are utilized for validation. DART provides high temporal resolution OBP data, down-sampled to 6-hour intervals for consistency with CRA-LICOM. OBP datasets include 68 locations spanning 2002–2023, primarily distributed over the Pacific and Atlantic Oceans. Then, an additional Argo data set is required to further confirm the temperature and salinity. Such monthly Argo data span 2005-2020 with a horizontal resolution of $1^\circ \times 1^\circ$ and a vertical resolution of 27 layers (depths of up to 2,000 m), which are available at https://apdrc.soest.hawaii.edu/projects/Argo/data/gridded/On_standard_levels/index-1.html.

2.2.5 GLDAS

The additional data set includes GLDAS (https://ldas.gsfc.nasa.gov/gldas/), which is based on advanced land surface modeling and data assimilation techniques to merge satellite- and ground-based observations into the model. GLDAS provides high-quality, global land surface fields to support the investigation of Terrestrial Water Storage (TWS) change (Li et al., 2019). In this study, we extract TWS (3 hours and $1^\circ \times 1^\circ$) from GLDAS to approximate the component of the global hydrology (H) signal to be compared with the AO component, as indicated by CRA-LICOM.

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3 Method

In summary, the process to obtain CRA-LICOM products consists of three major steps: (i) atmospheric gravity modeling, (ii) oceanic gravity modeling, and (iii) post-processing to produce the final CRA-LICOM, see Fig. 1 for a conceptual diagram of the framework. In what follows, the specific method in each step is addressed individually.

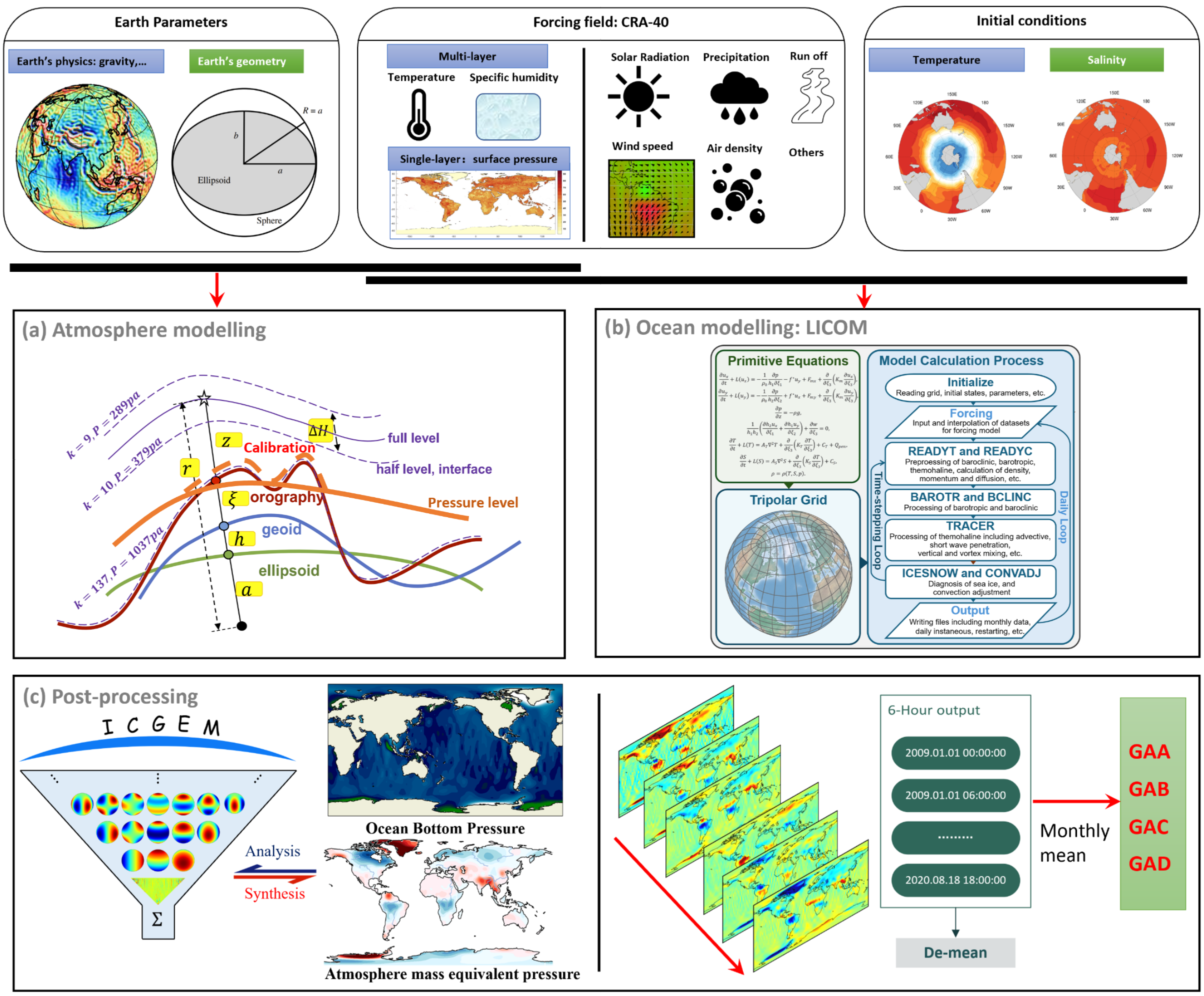

Figure 1. The diagram to illustrate the workflow of CRA-LICOM: from the input forcing field (associated with auxiliary parameters) to the output gravity products, where three major steps are addressed: (a) atmosphere modeling to calculate the surface mass and upper air mass contribution to the gravity field, using calibrated pressure level data, (b) ocean modeling with LICOM model to simulate the ocean bottom pressure forced by the atmospheric variables from CRA, (c) post-processing of the grid output to the spherical harmonic coefficients, the removal of long-term mean, and the aggregation of monthly products.

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3.1 Atmosphere

3.1.1 Atmospheric tidal constituent separation

The input surface pressure fields represent a mixture of tidal and non-tidal constituents, which need to be isolated as a first step. The logic behind the isolation is the fact that tides can often be better predicted by the model. As limited by Nyquist sampling law, the expected tidal signals extracted from CRA-40 (6 hourly) must be slower than the semi-diurnal tide, which includes P_1 (14.9589314 deg/h), S_1 (15.0000000 deg/h), K_1 (15.0410686 deg/h), N_2 (28.4397295 deg/h), M_2 (28.9841042 deg/h), L_2 (29.5284789 deg/h) and T_2 (29.9589333 deg/h). Then, a point-wise tidal pressure can be obtained from a summation of all those frequency-dependent tides ζ_s (assuming a tide s with frequency ω_s, amplitude ξ_s and phase δ_s) following

$$\zeta(\theta,\lambda,t) = \sum_s \xi_s(\theta,\lambda)\cos[\omega_s t + \chi_s - \delta_s(\theta,\lambda)] = \sum_s [A_s(\theta,\lambda)\cos(\omega_s t) + B_s(\theta,\lambda)\sin(\omega_s t)], \tag{1}$$

where (θ,λ) denotes spherical coordinate (colatitude, longitude) of the point and t denotes an arbitrary time epoch. In particular, χ_s is the Warburg phase correction documented by Petit et al. (2010). And in Eq. (1) the amplitude and phase can be translated into the coefficients $A_s(\theta,\lambda)$ and $B_s(\theta,\lambda)$ to enable the Ordinary Least Square (OLS) solution. In this study, OLS is configured by terms of trending $C(\theta,\lambda)$, tidal $\zeta(\theta,\lambda)$ and non-tidal $D(\theta,\lambda)$ signals:

$$P(\theta,\lambda,t) = \sum_s [A_s(\theta,\lambda)\cos(\omega_s t) + B_s(\theta,\lambda)\sin(\omega_s t)] + C(\theta,\lambda)t + D(\theta,\lambda), \tag{2}$$

where parameters (A_s, B_s, C, D) are fitted from the 'observations', i.e., time-series of surface pressure $P(\theta,\lambda,t)$. Subsequently, each tide ζ_s (its amplitude and phase) can be recovered from its coefficients (A_s, B_s) via Eq. (1). Be aware that a low-pass filter (with a time window of 3 days) is applied to P beforehand, to damp non-tidal signals first and stabilize the tidal estimations. Here, the tide constituents are fitted from the years 2007-2014.

3.1.2 Non-tidal air mass integration

To accurately reflect the non-tidal atmospheric gravity change, one has to exploit layered observations to account for contributions from both surface and upper air anomalies. To this end, two types of air mass integration are required: (1) a surface integration that considers the air mass as a thin layer, that is, by neglecting the vertical structure of air; (2) a 3-D vertical integration of all mass columns to obtain the upper air contribution. Regardless of either type of integration, the first step is to obtain the 'inner integral', i.e., I_n, which is often degree dependent (the degree of spherical harmonic expansion); see Forootan et al. (2013, 2014). For surface integration, I_n is treated as

$$I_n^s(r,\theta,\lambda) = (\frac{r}{a_e})^{n+2}\frac{\Delta P_0(r,\theta,\lambda)}{g(r,\theta,\lambda)} = (\frac{a(\theta,\lambda)+\zeta(\theta,\lambda)+h(\theta,\lambda)}{a_e})^{n+2}\frac{\Delta P_0(r,\theta,\lambda)}{g(r,\theta,\lambda)}, \tag{3}$$

where (r,θ,λ) is the spherical coordinate (radial distance, colatitude, and longitude) of the evaluated point. In the case of a realistic Earth, the radial distance r consists of the ellipsoidal radius a, the geoid undulation h , and the topography ζ (e.g., Yang et al., 2022). In addition, in Eq. (3) a_e denotes the Earth's mean radius, e.g., ∼6378136.6 km; $\Delta P_0(r,\theta,\lambda)$ indicates the surface pressure; and $g(r,\theta,\lambda)$ is the gravity acceleration of evaluated point.

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By contrast, the 3-D vertical integration is able to consider the vertical structure of the air by taking advantage of multi-level atmospheric input fields. Here CRA-40, in terms of pressure levels, is used that yields

175 $$I_n^v(r,\theta,\lambda) = \sum_{k=0}^{k_{max}} \left(\frac{a(\theta,\lambda)+\zeta(\theta,\lambda)+h(\theta,\lambda)+z_k(\theta,\lambda)}{a_e}\right)^{n+2} \frac{\Delta P_k(\theta,\lambda,r_k)}{g_k(\theta,\lambda,r_k)}, \tag{4}$$

where the inner integral is discretized into k-layers, and for CRA-40, it has maximal $k_{max} = 47$ layers. $\Delta P_k(\theta,\lambda,r_k)$ indicates pressure difference between adjacent layers k^{th} and $(k-1)^{th}$. Comparing Eq. (4) to Eq. (3), the z_k emerges as the geometric distance from the evaluated point to the surface, which must be solved from an accumulation of geopotential differences between adjacent pressure layers. To this end, the multi-level humidity and temperature fields are required to obtain the geopotential height difference; see Boy and Chao (2005) for more details.

It is worth mentioning that because CRA-40 is given in terms of pressure level, integration with Eq. (4) might face risks of 'outliers': there could be some cases in which the pressure of a layer is even greater than the surface pressure, in other words, the profile of the isobaric surface of this pressure layer goes through the interior of the Earth, which is obviously unreasonable in physics. It is relevant to calibrate this outlier. Otherwise, the modeling quality would be significantly degraded. To this end, we propose a calibration method, which yields

$$V_k(\theta,\lambda) = V_{k-1}(\theta,\lambda), \quad \forall \Delta P_k(\theta,\lambda,r_k) < 0, \tag{5}$$

where V_k indicates arbitrary variable in the k^{th} layer, which could be either pressure, temperature, or humidity, see our previous study (Zhang et al., 2025) for more details. In this manner, the outliers can be identified and fixed with a reasonable approximation (the neighboring value) to guarantee the robustness of the vertical integration.

3.1.3 Non-tidal atmospheric correction

An accurate modeling of the non-tidal atmospheric gravity field requires us to account for both the direct gravitational effect and the indirect Earth's deformation effect. Therefore, a combination of the hypothetical thin-layered air (with two further corrections) and the upper air is necessary. The implementation of the combination follows the method proposed by Yang et al. (2021), i.e.,

$$I_n = (I_n^s - I_n^{tide} + I_n^{IB}) + \frac{1}{1+k_n}(I_n^v - I_n^s), \tag{6}$$

where the k_n indicates the degree-dependent loading love number, and the quantity in the first bracket indicates the non-tidal surface air that accounts for both the direct and indirect effects. By contrast, the other quantity in the second bracket indicates the upper air that accounts for only the direct effect, which makes sense since it cannot lead to Earth's deformation. In addition, corrections are made to the surface integral that includes: (i) tide removal as indicated by I_n^{tide}, which can be modeled by Eq. (1), and (ii) Inverted bBarometer (IB) correction as indicated by I_n^{IB}, which is introduced to compensate for the overall static contribution of the atmosphere to the ocean; please see Dobslaw et al. (2017) for more details.

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3.2 Ocean

3.2.1 OBP simulation with LICOM3.0

We used the low-resolution global configuration of LICOM3.0 (Lin et al., 2020; Liu et al., 2012) to obtain 3 hourly, 360×218 tri-polar (equivalent to 1° on average) OBP data spanning 2002–2024, as depicted in Fig. 1b. The ocean model adopts primitive equations, including the Navier-Stokes equations, continuity equations, conservation equations for temperature and salinity, and the equation of state for seawater. These equations are discretized on a tri-polar grid with 360×218 horizontal points and 30 vertical layers. The model workflow consists of three main blocks:

1. **Initialize**: This block manages the reading of grids, initial states, and parameters. The initial conditions use global climatology for temperature and salinity from PHC3.0 (Polar Science Center Hydrographic Climatology; Steele et al., 2001)

2. **Forcing and Output**: This block inputs the forcing fields and outputs simulated data. The atmospheric forcing fields of the model are transformed from the 6 hourly $0.5°$ resolution variables in CRA-40 by applying the standard air-sea flux calculation methods of the Ocean Model Inter-comparison Project (Griffies et al., 2016; Large and Yeager, 2004).

3. **Time-Stepping Kernels**: This block contains the kernels for solving equations within the time loop. "READYT and READYC" computes terms in the barotropic and baroclinic equations. "BAROTR and BCLINIC" solve barotropic and baroclinic equations, while "TRACER" handles temperature and salinity equations. Furthermore, "ICESNOW and CONVADJ" deals with sea ice and deep convection in high-latitude regions (Wang et al., 2021).

More information on LICOM3.0 can be found in the Appendix A. In this paper, the model was spun up for five cycles from 2002-2023, and then an integration from January 1st, 2002, to existing months in 2024 started at the end of the fifth cycle is conducted and analyzed. The OBP in LICOM3.0 is the sum of atmospheric pressure and the vertical integration of seawater density between dynamic sea level and ocean bottom, computed as

$$P_b = P_a + H_0\rho(1)g + \sum_{k=1}^{30}\rho(k)\Delta z(k)g, \tag{7}$$

where P_b is OBP, P_a is atmospheric pressure, H_0 is dynamic sea level, ρ and $\rho(1)$ are seawater density and its value at surface level, g (=9.806 $m\,s^{-2}$) is the gravity acceleration, k is vertical layer number, and Δz is vertical layer thickness. Note that Eq. (7) is applied at the 3-D domain, that is, time t and position (θ, λ), and the dimension notation is omitted for readability.

3.2.2 Oceanic tidal constituents

For the same reason as atmospheric gravity modeling, ocean tides must be estimated and removed from OBP. Be aware that LICOM3.0 does not directly simulate the lunisolar gravitational tides in the oceans. Hence, tidal fluctuations revealed by the eventual OBP fields are solely induced by periodically varying atmospheric forcing. In addition, since atmospheric pressure is set to zero in the model's momentum equations, the induced oceanic tides primarily arise from tidal variations in wind stress. Therefore, the amplitudes and fluctuations of the simulated oceanic tides are likely much weaker than the reality. For instance, the global mean amplitude of S_2 simulated by LICOM3.0 in this study is approximately 10^1 Pa, compared to $\sim 10^3$ Pa in the tidal models by Huang et al. (2024). However, the global distribution patterns of the simulated oceanic tides closely align with those of atmospheric tides, see Sect. 4.2 for more details.

In addition to the seven tidal constituents mentioned in Sect. 3.1.1, S_2 (30.0000000 deg / h), R_2 (30.0410667 deg / h) and SK_3 (45.0410760 deg/h) are also removed using the T_TIDE package (Pawlowicz et al., 2002), due to a higher sampling rate of OBP, that is 3 hours. In this study, tide fluctuations are calculated annually for 2002–2023. Furthermore, be aware that the effect of atmospheric loading must be removed beforehand. Hence, the overall formula to obtain non-tidal OBP follows

$$P_b^{nt} = (P_b - P_a) - tide(P_b - P_a), \tag{8}$$

where P_b^{nt} is the non-tidal OBP, and the second term indicates the removal of tides.

3.2.3 Non-tidal OBP correction

Because of the Boussinesq approximation in the momentum equations of LICOM, mass conservation is not preserved any longer as the density changes within this volume-conservative model. To ensure mass conservation, a global mass correction has to be implemented following the method proposed by Greatbatch (1994). This correction involves subtracting the mean OBP across the entire ocean domain at each time step, which yields

$$P_b^{dyn} = P_b^{nt} - 1/A_{oceans} \int P_b^{nt} dA, \tag{9}$$

where P_b^{dyn} is the non-tidal dynamic OBP, which shall be sent to Eq. (3) as well to obtain the inner integral I_n^O of the ocean. As indicated in Eq. (9), the role of the ocean-land mask is not negligible. Be aware that the ultimate temporal resolution of OBP is down-sampled from native 3-hour to 6-hour to be consistent with that of the atmospheric forcing field.

3.3 Post-processing to CRA-LICOM

Having obtained the degree-dependent inner integral of the atmospheric (I_n^A) and oceanic (I_n^O) components, a harmonic analysis that maps the global multi-level gridded pressure fields ($I_n = I_n^A + I_n^O$) to the gravity field is necessary, which yields

$$C_{nm} + iS_{nm} = \frac{3}{4\pi\rho_e} \frac{1+k_n}{2n+1} \int_0^{2\pi} \int_0^{\pi} I_n \cdot P_{nm}(\cos\theta) e^{im\lambda} \sin\theta d\theta d\lambda, \tag{10}$$

where $P_{nm}(\cos\theta)e^{im\lambda}$ denotes the normalized surface spherical harmonics, k_n is the loading love number, and ρ_e is the Earth mean density, 5517 kg/m^3; $[C_{nm}, S_{nm}]$ are corresponding coefficients of spherical harmonic expansion at degree n and order m, which shall constitute the so-called gravity field. Solving $[C_{nm}, S_{nm}]$ from Eq. (10) relies on a two-step procedure following Sneeuw (1994), but implemented in practice degree by degree due to the nature (degree dependent) of I_n. In addition,

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a long-term mean from 2003-2014 is subtracted from the time series, and the derived anomaly ΔI_n instead of I_n is applied to Eq. (10) eventually.

To be consistent with the convention of satellite gravity (Dobslaw et al., 2017), CRA-LICOM is further classified into four 6-hour products: ATM (indicating only the non-tidal high-frequency atmospheric gravity field), OCN (indicating only the non-tidal high-frequency oceanic gravity field), GLO (the sum of ATM and OCN) and OBP (indicating only the ocean bottom pressure, and thus it excludes the upper air contribution). Eventually, these four 6-hour products are correspondingly post-processed into monthly mean products (ATM→GAA, OCN→GAB, GLO→GAC, OBP→GAD) for scientific users who are only concerned with the large time scale. All of these constitute our ultimate CRA-LICOM product, but 'CRA-LICOM' hereafter always indicates the 'GLO' product for simplicity, unless otherwise a special statement.

4 Results

4.1 Standard product

CRA-LICOM produces a high-frequency (6 hourly) global gravity field product with a spectral resolution of degree/order 180, which exceeds the capabilities of GRACE and GRACE-FO missions (∼300 km or degree/order 60, see Landerer and Swenson, 2012). The product spans 2002 to 2024, and future updates will extend its coverage beyond 2024.

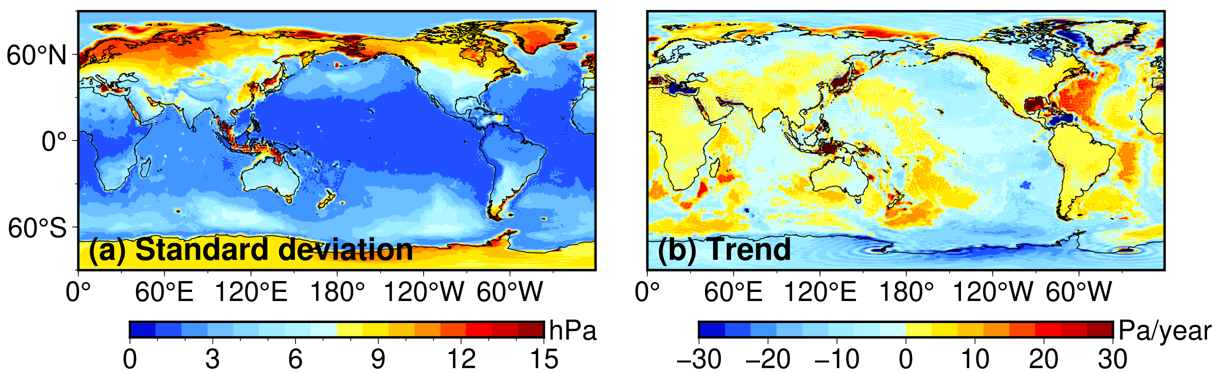

Figure 2. Equivalent pressure fields synthesized from CRA-LICOM during 2002-present: (a) the standard deviation (hPa), (b) the secular trend (Pa/year).

The standard deviation (STD) and secular trend of the product are illustrated in Fig. 2. The global mean STD is 4.83 hPa, comparable to GFZ-RL07 (4.98 hPa, see Fig. B2 of Appendix B). Variations are higher over continents (mean STD: 7.40 hPa) than over oceans (mean STD: 3.10 hPa), suggesting a dominant contribution from atmospheric changes over land. This is reasonable because the atmospheric active change and the oceanic passive reaction often cancel out over the open oceans. The secular trend shows a maximum of 245.05 Pa/year and a minimum of -193.93 Pa/year at a confidence level 95%. Obviously, these trends contribute minimally to the overall signal, comparing Fig. 2b to a. However, areas with significant trends in Fig. 2b warrant cautious interpretation, particularly for GRACE(-FO) gravity fields obtained with CRA-LICOM. However, a majority

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(80.2%) of the trend map is still within the range of ±10 Pa/year, equivalent to ∼1 mm of change in water per year. For these areas, the CRA-LICOM trend could be considered as uncertainty due to the fact that the GRACE (-FO) error can reach up to a few centimeters (Yang et al., 2024a). However, this might be worth considering for next-generation gravity missions that target an accuracy of a few millimeters.

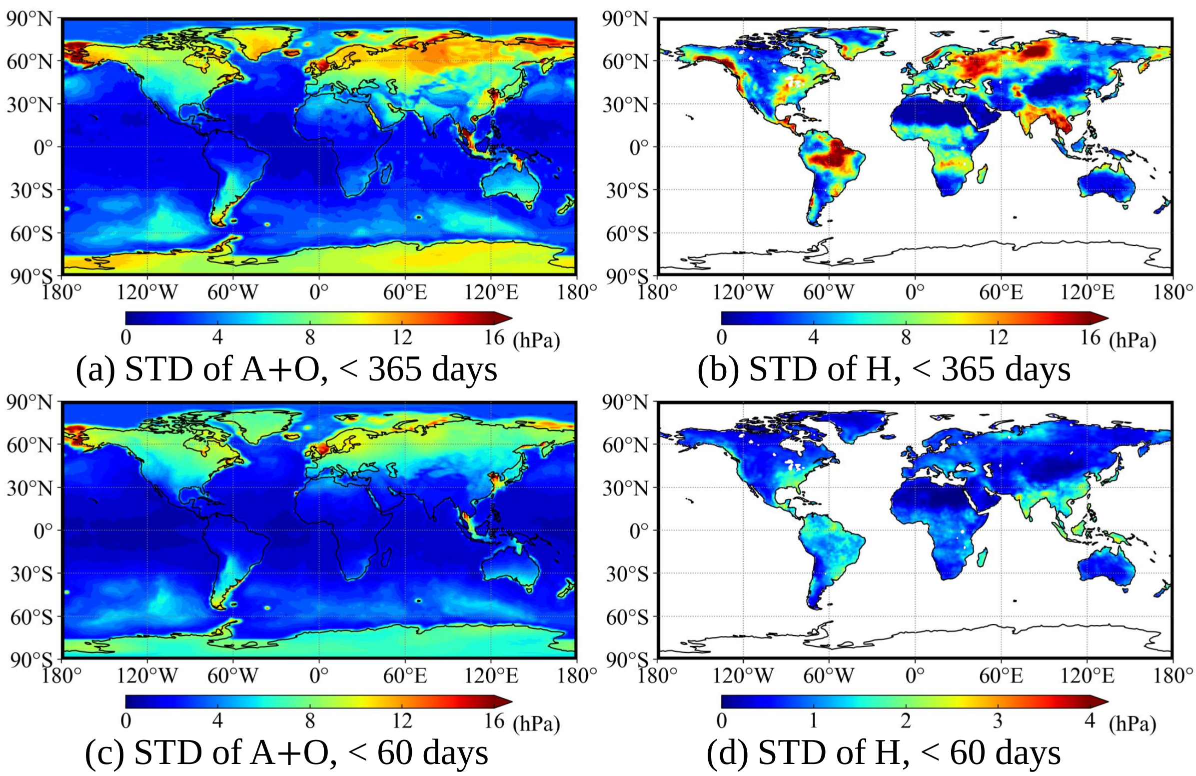

Figure 3. Comparison between AO and H variation of one year (2005). (a-b) are STD of an entire year, while (c-d) are STD after Butterworth high-pass filtering with a 60-day cutoff window.

To understand the contribution of AO (CRA-LICOM) to the Earth's gravity system, we compared it against hydrology (H) variations using GLDAS-TWS data. The hourly GLDAS-TWS is down-sampled to 6 hours to be consistent with AO. The experiment is carried out over 12 consecutive months (2005.01-2005.12, arbitrarily chosen), where the STD of one year of both AO and H are calculated, see Fig. 3a-b. It is found that A has a mean STD of 6.65 hPa (over the continents), O's mean STD is 4.27 hPa (over the oceans), and H's mean STD is 5.76 hPa (over the continents). Despite an overall comparable magnitude, much more pronounced variations can be captured from H than A over the climate zones, e.g., the Amazon and Ganges Delta. Then, we introduce high-pass filtering (Butterworth filtering) with a 60-day cut-off window to the one-year time series to retain only the high-frequency signals within the aliasing spectrum (twice the sampling rate of monthly gravity solution from GRACE(-FO)); see Fig. 3c-d. In this case, the AO component is much larger than H, that is, the mean std of AO and H are 3.71

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hPa and 0.76 hPa, respectively. And AO has higher magnitudes compared to H in 94.2% of continental areas. This confirms the necessity of incorporating AO in studies focusing on high-frequency gravity changes.

4.2 Auxiliary product

CRA-LICOM also provides auxiliary products, including tidal constituents and upper air anomalies. As primary tides, the solar diurnal tide S_1 and the lunar semi-diurnal tide M_2 in the atmosphere are shown in Fig. 4a and b, where S_1 has a global mean of 35.59 Pa with a particular spatial pattern. S_1 is more pronounced on continents (up to 120 Pa) than on oceans, with most of its energy concentrated in the Southern Ocean and mid- to low-latitudes. The major tide S_1 obtained, in terms of magnitude and spatial pattern, is fairly consistent with that reported in the official product of GFZ (Dobslaw et al., 2016). The other atmospheric tides are considerably smaller, for example, the global mean amplitude of M_2 is around 2.00 Pa. For these smaller tides, we claim that their spatial pattern is heavily influenced by the employed forcing fields, so that they might look differently; for example, our M_2 differs a lot from that of Dobslaw et al. (2016). This can be confirmed by a supplementary experiment in Fig. B1 in Appendix B, where we use the same method but a different forcing field (ECMWF reanalysis) to obtain tides that are extremely similar to the official GFZ product. Be aware that CRA-LICOM does not estimate and remove the atmospheric semi-diurnal tide of S_2 due to the coarse time resolution (6-hour) of forcing fields, which means that one must not add back S_2 to avoid double bookkeeping.

In correspondence, the oceanic tides of S_1, M_2 are demonstrated. Unlike atmospheric tides, the oceanic tide S_2 is also available due to a higher sampling rate of OBP, see Sect. 3.2.1. In general, one can see from Fig. 4c–d that ocean tides exhibit spatial patterns similar to their atmospheric counterparts but with generally smaller amplitudes due to the underlying mechanism that only the oceanic response to atmospheric forcing is simulated; see Sect. 3.2.2 for more details. For example, the global mean of S_1 and M_2 in the ocean is 17.95 Pa and 1.08 Pa for the oceanic tide, but that is 23.31 Pa and 2.09 Pa for the atmospheric tide. However, the high similarity in the spatial domain can still confirm the internal consistency between the atmospheric and oceanic tides obtained. Figure 4e further confirms the oceanic solar semi-diurnal tide S_2 as the largest tide with a global mean of 37.23 Pa. Its spatial pattern with four zonally distributed peaks (up to 90 Pa) has reproduced a majority of the features of previous work as well (Dobslaw et al., 2016), although not perfectly. The discrepancy is because of (i) the lower temporal resolution (6 hours) of the forcing field and (ii) the use of different atmospheric forcing fields.

Another auxiliary product is the upper air anomaly, which is obtained by $I_n^v - I_n^s$ in Eq. (7) and thereby can be an indicator of multi-level atmospheric data quality. Although the magnitude is small compared to the surface pressure, the upper air anomaly constitutes a non-negligible component of the atmospheric dealiasing product (Swenson and Wahr, 2002). In Fig. 5, we calculate the standard deviation from various time periods to investigate the variation. The global mean of each scenario from Fig. 5a-d is found to be 17.43 Pa, 16.88 Pa, 17.75 Pa, and 17.34 Pa, respectively. Although this magnitude is much smaller than that of the total AO signal in Fig. 2, it has a magnitude as large as the amplitude of the major tides in Fig. 4, so it is not negligible. In addition, by comparing Fig. 5a-c to Fig. 5d, one can further see that upper air anomaly does not exhibit evident annual variation, and all preserve a similar spatial pattern. This fact suggests a rather stable contribution of upper air anomaly modeling due to the nature of ocean circulation.

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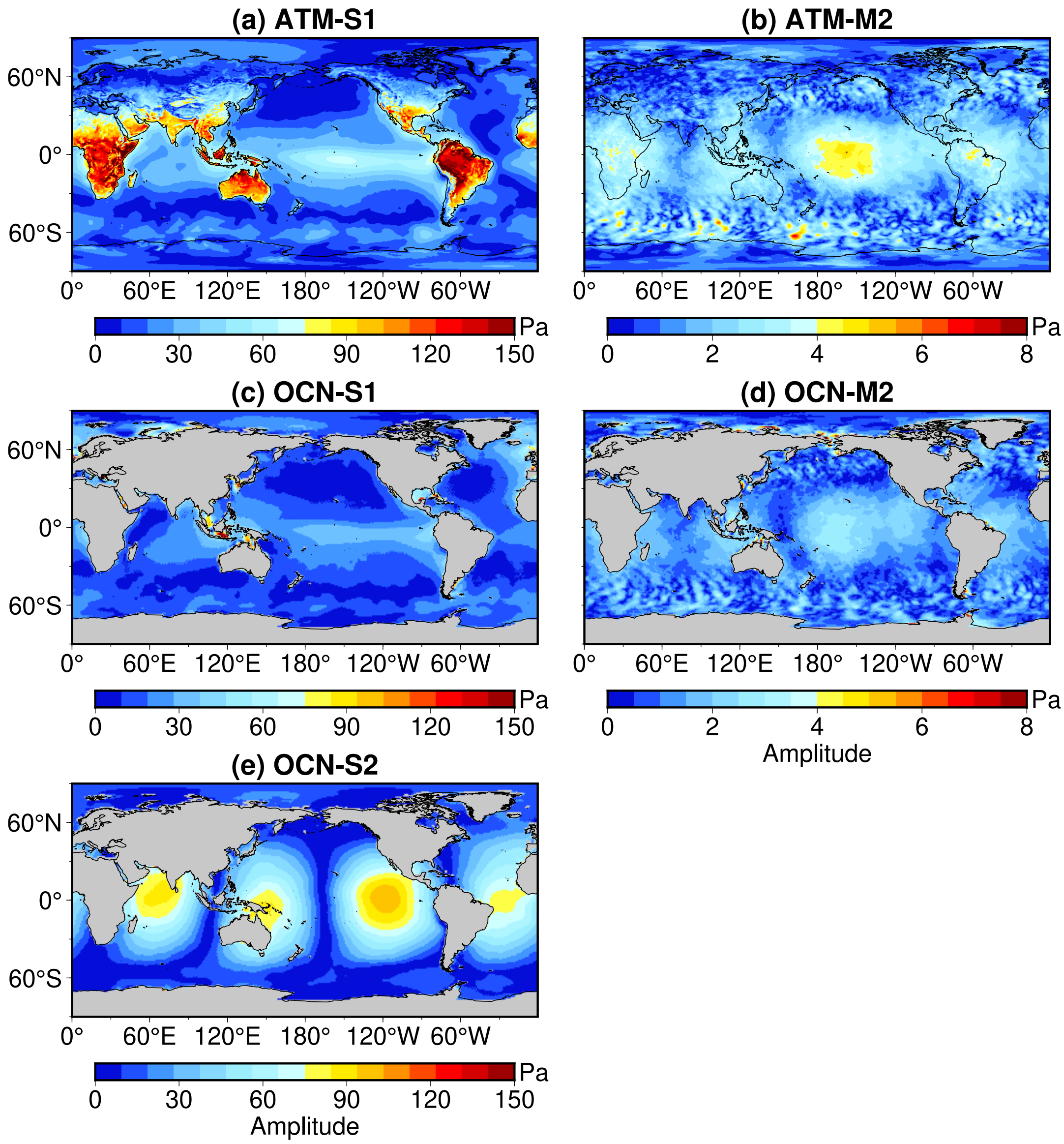

Figure 4. Amplitudes (Pa) of tidal estimated by CRA-LICOM over 2007–2014. The top panels show atmospheric tides based on 6 hourly data: (a) S_1 and (b) M_2. The middle and bottom panels demonstrate ocean tides derived from 3 hourly OBP data: (c) S_1, (d) M_2, and (e) S_2.

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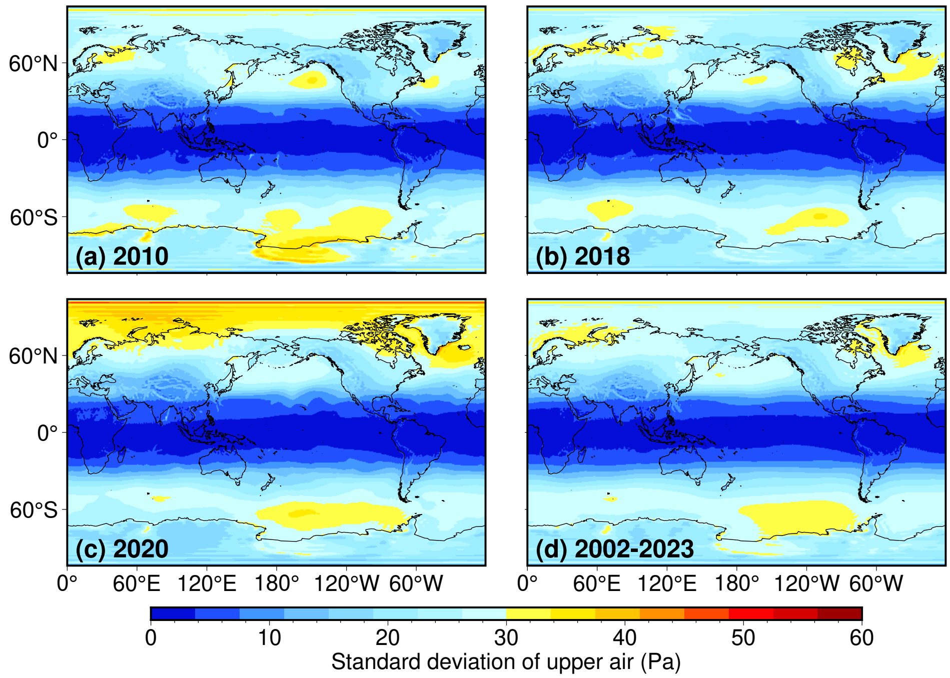

Figure 5. The standard deviation (Pa) of fields synthesized from upper air anomaly (mean field is removed) for (a) 2010, (b) 2018, (c) 2020, and (d) 2002-2023.

5 Validation and Applications

5.1 Inter-comparison with GFZ-RL07

5.1.1 Temporal correlation and bias analysis in spectral/spatial domains

In this section, a straightforward comparison is made at the product level itself. To assess the temporal performance of CRA-LICOM, we performed a detailed comparison with the official GFZ-RL07 product. Temporal correlations and biases were evaluated in both spectral and spatial domains.

First, in the spectral domain, the Stokes coefficients of CRA-LICOM and GFZ-RL07 products were analyzed for all degrees up to 60, for example. Figure 6 presents the mean temporal correlation coefficients of the Stokes coefficients per degree. At lower degrees (previous to 15), CRA-LICOM exhibits correlations exceeding 0.8, demonstrating strong agreement with GFZ-RL07 for gravity signals on a medium to large spatial scale. The consistency of low degrees is important since it is known that the AO model, as well as the GRACE gravity field, has its major energy at those degrees. As the degree increases, the correlation gradually decreases but remains statistically significant at the confidence level of 99%. The peak correlation occurs at degree 5 for coefficient C (0.89) and degree 6 for coefficient S (0.90). Then, the standard deviation (STD) ratios of the two products were also analyzed to evaluate the variability biases; see the dashed curves in Fig. 6. CRA-LICOM generally

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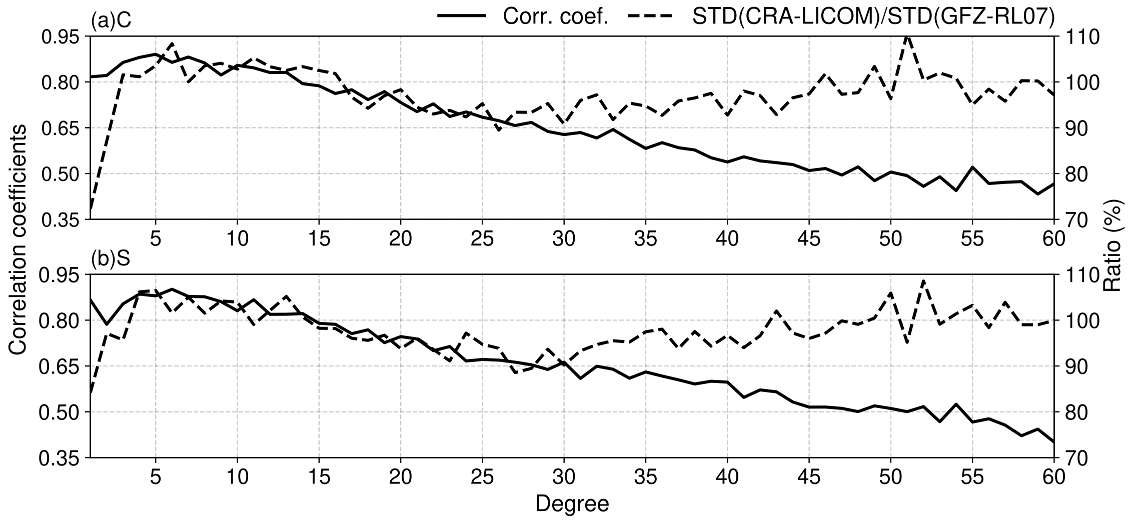

Figure 6. Mean temporal correlation coefficients (solid) and variation bias (dashed) for (a) C and (b) S at each spherical harmonic degree between CRA-LICOM and GFZ-RL07 during 2002-2024.

slightly underestimates the variability coefficients compared to GFZ-RL07, with STD ratios stabilizing over 90% in the whole spectrum. Despite a decline at lower degrees, the ratio is constantly increasing from degrees 30 to 100, eventually reaching around 100% at degree 100 (not shown), indicating CRA-LICOM's capability to reproduce high-frequency signals effectively.

Then, evaluations are made in the spatial domain by projecting two products onto pressure fields on a regular grid $1° \times 1°$. To be consistent with the 1-month resolution of the present satellite gravity mission, the time series pressure fields are decomposed into a frequency variability < 60 days and another frequency variability > 60 days. Figure 7a illustrates the temporal correlation coefficients for that < 60 days, from which we see: (1) nearly all are statistically significant at a 99% confidence level; (2) correlation generally decreases as the latitude increases, inferring a bias for unknown reason; (3) the correlations are substantially greater overland (0.80 on average) than over ocean (0.54 on average). The high correlation (0.80 on average) over land indicates an overall consistency of employed atmospheric forcing fields, despite a few exceptions, such as Central Africa and the Northern region of South America, where the correlation coefficient degrades to around 0.7. We attribute this degradation as a consequence of the remaining S_2 atmospheric tide in CRA-LICOM since the spatial pattern resembles that of S_2 at a high similarity, see Yang et al. (2021). In contrast, S_2 has been removed from GFZ-RL07. For oceanic regions, mid- to high-latitudes have a correlation coefficient above 0.6, while it is lower at mid- to low-latitude oceans, particularly the Atlantic Ocean, and marginal seas exhibit weaker correlations down to 0.4, indicating a larger discrepancy between the two ocean models in these areas.

Furthermore, Fig. 7c illustrates the root mean square error (RMSE) of global temporal variability at a time window of < 60 days. The global mean RMSE is found to be 2.25 hPa, while most pronounced biases are observed in the continental shelf regions of the Arctic Ocean, offshore China, and Hudson Bay, with RMSE peaks of up to 15 hPa. The elevated RMSE in

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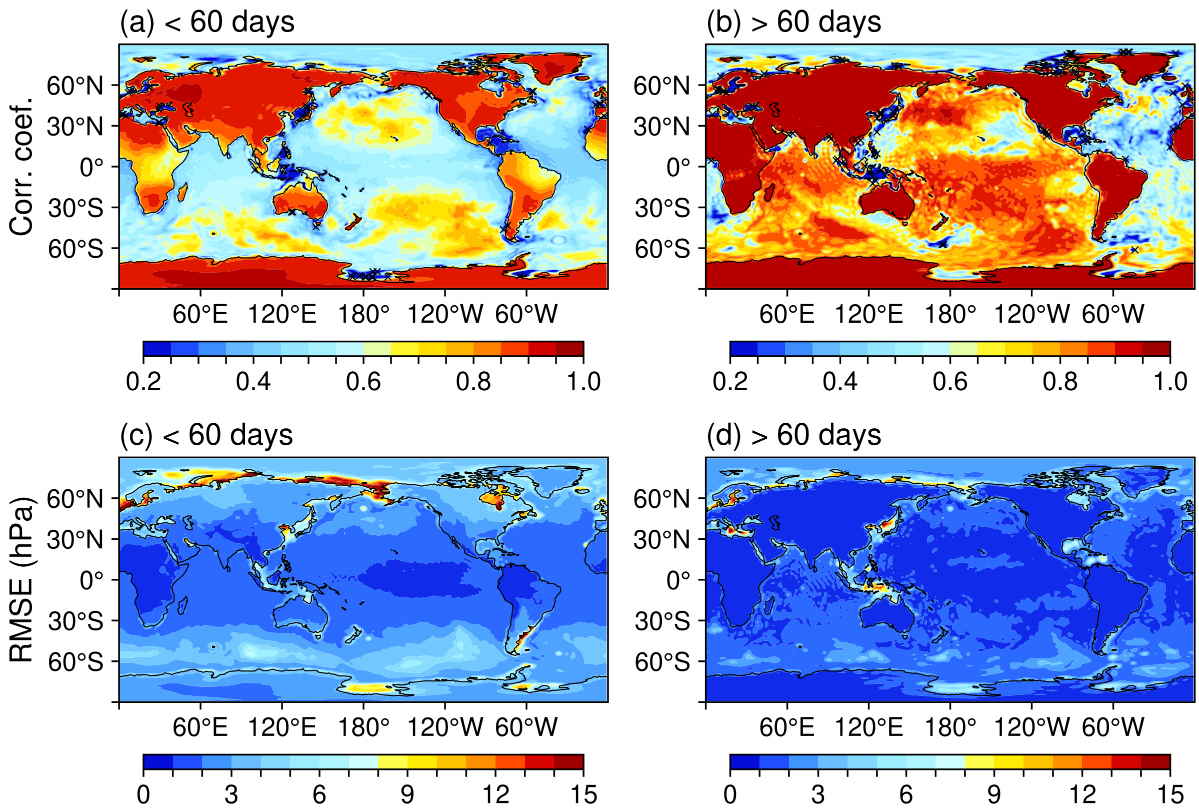

Figure 7. Temporal correlation coefficients (in terms of synthesized pressure fields) between CRA-LICOM and GFZ-RL07 during 2002–2023 for periods (a) < 60 days and (b) > 60 days. Panels (c) and (d) show the corresponding root mean square error (RMSE, hPa). In (a) and (b), locations marked with "x" indicate data not statistically significant at the 99% confidence level. Frequency bands are separated using fourth-order Butterworth filters.

the Ross Sea can be attributed to LICOM's inability to simulate OBP in this region. Furthermore, notable biases are evident in the Southern Ocean, where the RMSE averages around 4 hPa, smaller than the STD value, approximately 8 hPa. The observed biases elsewhere may stem from differences in how models handle topography and key sea channels, leading to error accumulation along Western boundaries. These findings suggest that CRA-LICOM effectively captures consistent temporal variability amplitudes across most regions, except the marginal seas near continental shelves. Figures 7b and 7d demonstrate the correlations and biases for periods >60 days, revealing stronger correlations and smaller biases globally, while this has little impact on satellite gravity due to its spectrum being slower than the aliasing frequency. The average correlation coefficients are 0.89 over land and 0.70 over oceanic regions, with a global mean RMSE as low as 1.30 hPa. This confirms the improved agreement between CRA-LICOM and GFZ-RL07 for longer periods. However, model uncertainties are more pronounced at higher frequencies, which challenges OBP simulations in the context of de-aliasing satellite gravity observations.

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5.1.2 On-orbit validation via prefit KBRR-residuals

The observation of satellite gravity on orbit along the track, for example, the inter-satellite K-band range rate (KBRR), is extremely sensitive to the geophysical process over regions where the satellites fly (Ghobadi-Far et al., 2020). Therefore, KBRR, especially its residuals after removing essential background geophysical signals, including AO, can be an effective indicator of the quality of the AO model (Zenner et al., 2010; Yang et al., 2018). Here, we use data from Sect. 2.2.2 to calculate the KBRR residuals for an initial diagnosis of two AO products, ie, GFZ-RL07 and CRA-LICOM. All data processing to obtain the final KBRR residuals is manipulated by our internal open source Python software (namely PyHawk, https://github.com/NCSGgroup/PyHawk.git, see also Wu and Yang (2024)), which indeed has achieved a complete data processing chain from Level-1b raw data to Level-2 temporal gravity fields.

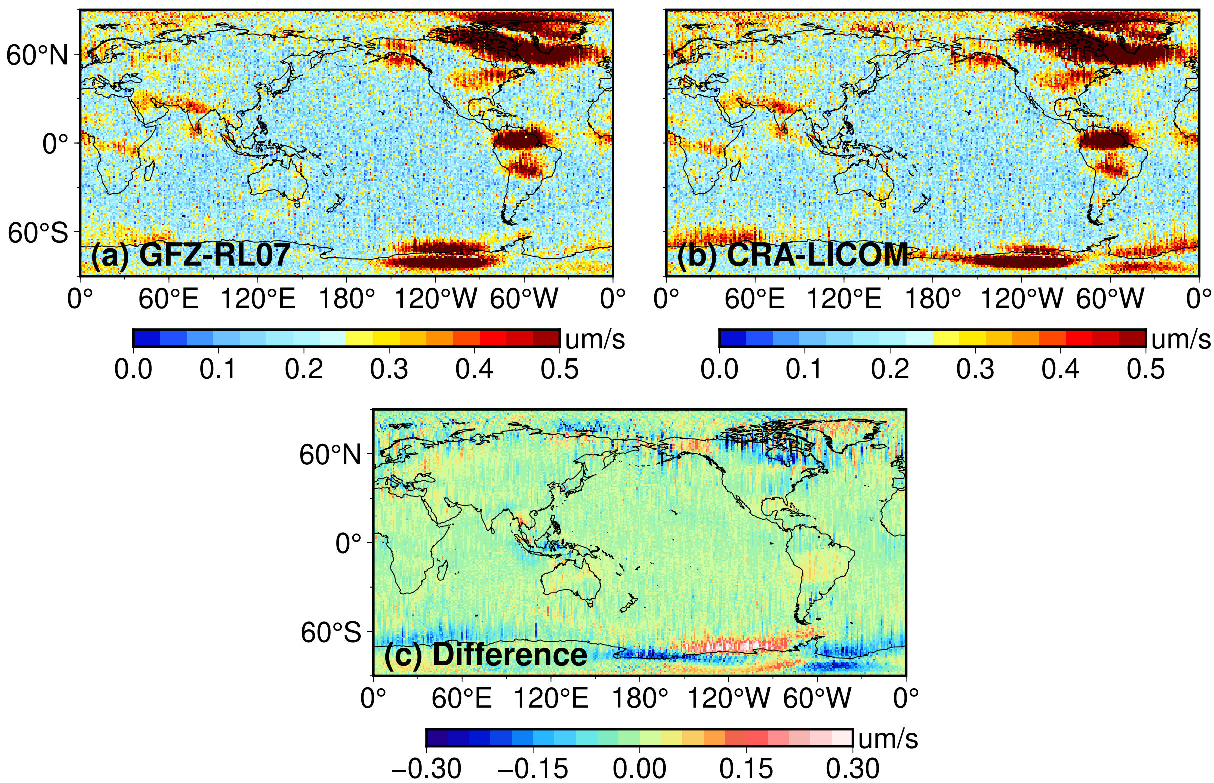

Figure 8. Prefit KBRR-residuals for GRACE using AO product, i.e., (a) GFZ-RL07 and (b) CRA-LICOM, respectively. One-month KBRR-residuals on December of 2010 were firstly assembled as gridded RMS (root mean square) by GRACE's ground track (mid of twin satellites) and projected into a map of $1^\circ \times 1^\circ$. The grid with a negative value at the map (c, GFZ-RL07 minus CRA-LICOM) may indicate where GFZ-RL07 outperforms CRA-LICOM and vice versa.

Figure 8 illustrates the spatial map of the KBRR residuals in terms of gridded root mean square (RMS), where an arbitrary month is selected as an example. Comparing Fig. 8a to b, both scenarios, as expected, have shown an ability to capture plausible

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signals, for example, over polar regions, glaciers, and tropic areas. The two scenarios also agree with each other very well; for example, the spatial correlation coefficient is 0.92 and the relative difference of their global mean RMS are as small as 4.7%. Since AO is utilized as a prior model to be removed from KBRR, one can always expect a smaller KBRR residual if the prior model better reproduces the reality. To this end, for the difference data in Fig. 8c, we exclude meaningless noise by setting a threshold of 0.1 $um\ s^{-1}$ (the accuracy limit of KBRR); as a result, for the remaining data, the proportion (relative to the whole map) of positive and negative grids is 2.08 % and 4.07 %, respectively. The small proportion indicates that KBRR is insensitive to a majority of their differences, while GFZ-RL07 slightly outperforms CRA-LICOM due to a higher proportion of negative.

5.1.3 Temporal gravity recovery and its error analysis

As one step further than in the previous section, we recover Earth's temporal gravity fields up to a degree/order of 60 for five years, which is also one of the ultimate goals of satellite gravity. The period from 2005-2010 is selected as an example to take advantage of GRACE's stable performance to obtain a convincing result. Then, a series of standard post-processing of obtained gravity fields is made, which include (but are not limited to): (1) conversion of gravity field to equivalent water height (EWH) at a gridded map of $0.5° \times 0.5°$, (2) replacement of low-degree Stokes coefficients (Loomis et al., 2020), (3) spatial filtering, DDK3, to damp the noise (Kusche, 2007; Yang et al., 2024b), (4) removal of glacial isostatic adjustment (Caron et al., 2018) etc. Subsequently, a linear regression to extract the climatology, as well as the residuals, is performed to indicate the signal and noise level of the obtained gravity field. All the aforementioned post-processing and signal/error analysis are achieved by our in-house open source Python software (called SaGEA, https://github.com/NCSGgroup/SaGEA), which follows a standard workflow to handle the Level-2 gravity fields, please see Liu et al. (2025b) for more details.

Figure 9 illustrates the comparison between the official latest CSR gravity product (where GFZ-RL07 AO model is employed as background model) against our gravity field (denoted as CRA-LICOM as well for brevity) obtained with CRA-LICOM AO model. From the first two rows, one can see that both products reach a comparable signal level, whether for the secular trend (spatial correlation as high as 0.97) or the annual amplitude (correlation of 0.91), confirming that CRA-LICOM is a qualified alternative AO product for present satellite gravity. In terms of the noise level, one can see from Fig. 9 that our gravity solution has a substantially higher noise than the CSR solution; however, be aware that this error is still among the range of GRACE gravity solutions (Chen et al., 2021). Moreover, the error pattern as revealed by Fig. 9e-f altogether resembles that of Fig. 7 in many places, such as the Ross Sea, Indonesia, the coastal area of the Arctic Sea, the black sea, the Baltic Sea, etc., confirming that (i) AO uncertainty has a major contribution to the ultimate gravity field's uncertainty and (ii) the current AO model is less plausible in these areas. Comparing Fig. 9e-f, we find that one may have apparently larger noise than another somewhere, e.g., Hudson Bay. For areas where our product appears more noisy, the reasons are likely twofold: (i) our ability to recover temporal gravity field from GRACE is not as good as CSR due to the fact that CSR has the highest signal-to-noise ratio even among existing official products, and (ii) potential model imperfection/inconsistency at specific places as indicated by Fig. 9f, and the justification about this point will be detailed in Sect. 6 as well.

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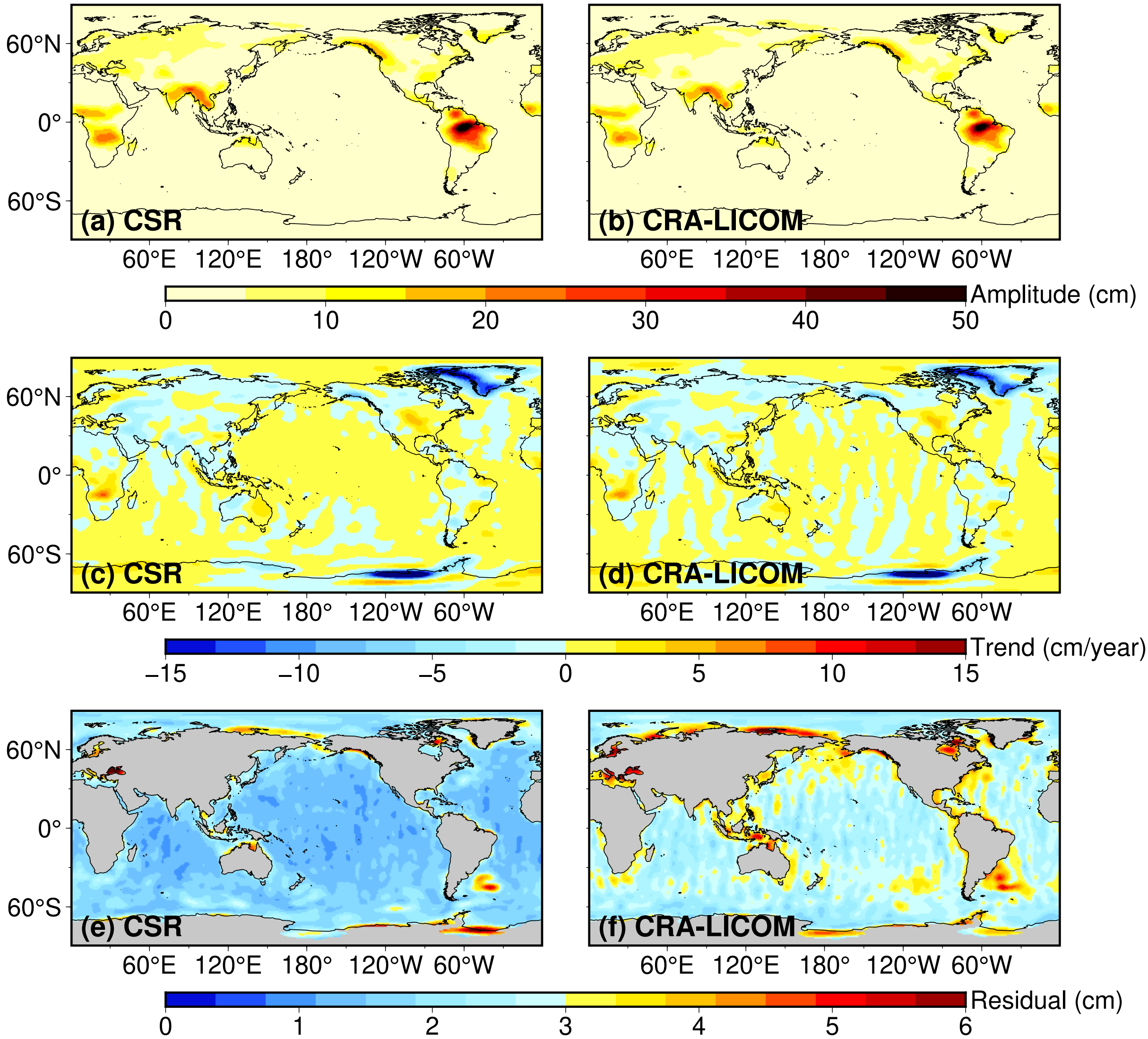

Figure 9. Temporal gravity field recovery (2005-2010, in terms of equivalent water height) using CRA-LICOM and its comparison against the latest CSR product. The signals, in terms of annual amplitude and secular trend, are demonstrated in Fig. 9(a,b,c,d); the error (or noise) level, indicated by standard deviation of ocean residuals with climatology removed, is present in the last row.

5.2 Validation against OBP recorders and Argo observations

Direct observational data from OBP recorders were used to validate the OBP simulated by LICOM. Figure 10 illustrates the non-tidal OBP STDs from DART and CRA-LICOM, both of which demonstrate stronger OBP variability in the North Pacific compared to other regions. The OBP STDs simulated by LICOM are generally smaller than the in situ data. Specifically, the mean STDs for 68 locations for DART is 2.87 hPa, while 1.78 hPa for LICOM. This systematic bias may partly be caused by the

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uncertainties in the model and forcing. However, the insufficient sampling of DART may also be attributed to the discrepancy. We found that about 85% of observational sites (58 out of 68) are located near the land or in the marginal seas, where the uncertainty of the model is particularly pronounced. We have evaluated the simulation against the near-land and open-ocean sites separately. Here, we define the near-land sites as the sites within 1500 km of the coastline. The relative bias of the OBP STDs between CRA-LICOM and DART is -34.4% for near-land and -26.5% for the open ocean.

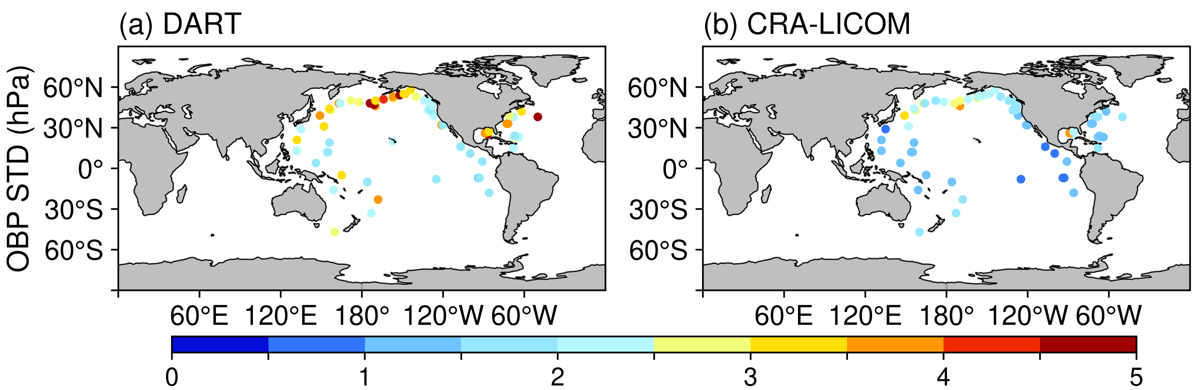

Figure 10. Non-tidal OBP STDs (hPa) from (a) DART and (b) CRA-LICOM during 2002-2023.

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Another indirect validation is performed by investigating the key variables of ocean simulation, i.e., the temperature and salinity, which together define the density and eventually influence the bottom pressure. Figure 11 presents the difference in temperature and salinity in terms of temporal mean and STD between LICOM and Argo for the upper 2000 meters during 2005–2020. We note that either temperature or salinity is computed as the vertical mean of the ocean up to 2000 meters, which shall not be mentioned again for readability. As a reference, we report the global mean of the following variables from Argo: the temporal mean and STD of temperature is 6.22 °C and 0.14 °C; the temporal mean and STD of salinity are 34.69 psu and 0.014 psu; their spatial distribution can be somewhat inferred from the contours of Fig. 11 as well. Compared to this reference, the bias (CRA-LICOM minus Argo) in terms of global mean is much smaller: the temporal mean and STD of temperature is 0.025 °C and -0.030 °C; temporal mean and STD of salinity are 0.025 psu and -0.0027 psu. The bias in terms of relative percentage is 0.4%, 21.4%, 0.1%, and 19.3%, respectively, for these four variables. Although CRA-LICOM exhibits a slightly smaller variation (STD), be aware that the observations of Argo suffer from considerable uncertainty as well. Apart from this, all other evidence demonstrates an accurate simulation of the temperature and salinity of CRA-LICOM against in situ observations across the majority of the oceans, which further confirms the model's ability to capture upper-layer density and reproduce ocean states and variability. Furthermore, the variables simulated by LICOM are comparable to those of other leading ocean models (Tsujino et al., 2020; Treguier et al., 2023; Chassignet et al., 2020), providing a solid foundation for effective OBP simulations. However, the spatial heterogeneity revealed in Fig. 11 should also be taken into account. In particular, an increased bias could be seen in the tropical Pacific, the western coasts of the mid- to high-latitude Atlantic, and

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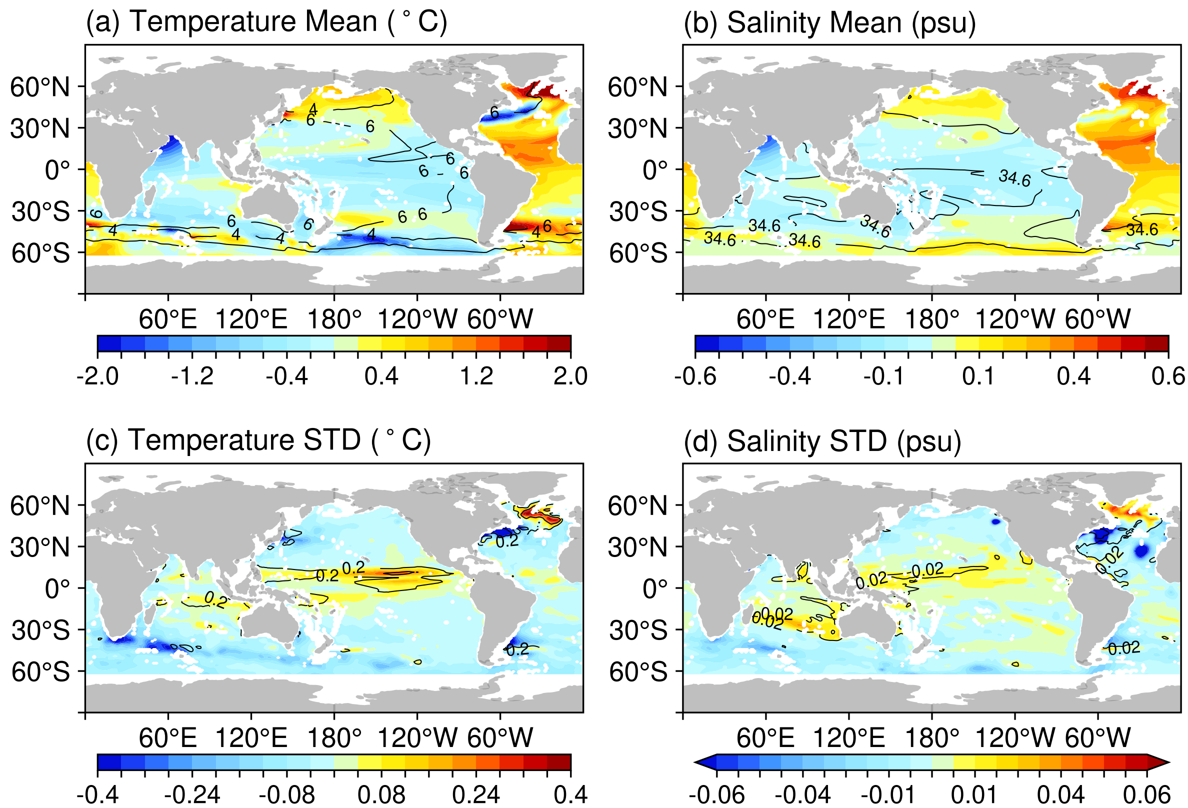

Figure 11. All the gridded value is derived as the average over vertical dimension by thickness-weight for the upper 2000 meters. Time mean (top panel) and STD (bottom panel) are obtained from the period of 2005-2020. Across all subfigures, the selected variable from Argo is illustrated in contours as the reference, while the difference (i.e., CRA-LICOM minus Argo) is visualized in a shaded manner. The unit of temperature and salinity is °C and psu, respectively.

the southern ocean, indicating a greater uncertainty or potential problems of simulated temperature and salinity in these places. The next update of CRA-LICOM will focus on areas with significantly stronger bias or weaker STD.

5.3 Validation against Argo and Altimetry observations

On the one hand, satellite gravity (e.g. GRACE) can well reveal the total mass change of the ocean, i.e., water from land/glaciers into the ocean; on the other hand, the accompanying monthly mean ocean bottom pressure product (i.e., GAD; see Sect. 3.2.3) reflects the change in the dynamic mass of the ocean. By convention, these two components together can be a measure of the manometric ocean (Gregory et al., 2019), and consequently, GRACE+GAD has been widely used for the investigation of the change in global mean ocean mass (GMOM) (Uebbing et al., 2019). In addition, enforced by the ocean budget equation, Argo-induced steric ocean change and Altimetry-induced sea level rise, if combined together, can be another reliable source of

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450 estimating GMOM change. Therefore, we used Argo+Altimetry as an external independent observation to validate the GRACE solution as well as our GAD (CRA-LICOM) product. Details on Argo + Altimetry can be found in Sect. 2.2.3.

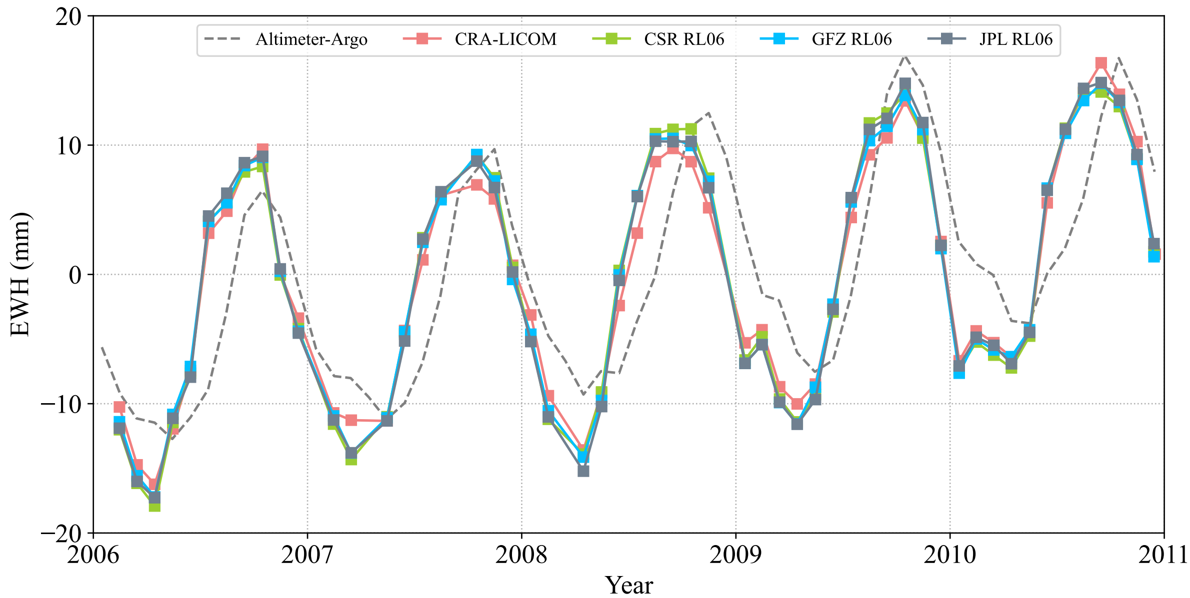

Figure 12. Global mean ocean mass change inferred from Altimeter-Argo and various GRACE solutions. The AO product GFZ-RL07 is added back to GRACE's official gravity solutions, including CSR, GFZ, and JPL release 06, to obtain the manometric sea level. Instead, our AO product, i.e., CRA-LICOM, is added back to our GRACE gravity solution (see Sect. 5.1.3) for consistency.

Table 1. Secular trend and (semi-)annual amplitude of GMOM change inferred from Altimeter+Argo, or from GRACE+GAD.

2006-2010	Linear trend (mm/year)	Annual amplitude (mm)	Semiannual amplitude (mm)
Altimeter-Argo	2.47 ± 0.13	9.24 ± 0.26	2.06 ± 0.25
CRA-LICOM	1.87 ± 0.18	10.81 ± 0.36	1.08 ± 0.35
CSR RL06	1.86 ± 0.18	11.87 ± 0.37	0.87 ± 0.36
GFZ RL06	1.79 ± 0.19	11.62 ± 0.38	0.77 ± 0.37
JPL RL06	1.89 ± 0.19	11.92 ± 0.38	0.97 ± 0.36

Here, we select three official GRACE Level-2 gravity field products (CSR RL06, JPL RL06, GFZ RL06, see Sect. 2.2.2 for more details) other than ours for a comparison. The gravity fields are first projected onto a $1^\circ \times 1^\circ$ gridded EWH map, then the GAD is added back, and finally the GMOM is derived for the global open ocean with a buffer area of 300 km to
455 avoid leakage error from continents to oceans (Chen et al., 2018). The dynamic variation of the time series is illustrated in Fig. 12, and the climatology indices are reported in Table 1. From Fig. 12, we see an overall agreement between Altimeter + Argo and GRACE + GAD, despite the obvious phase delay that is common in previous studies. In addition, there is still a non-

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negligible discrepancy between these two measurements for trend and annual amplitude, but this can be assumed as uncertainty (Uebbing et al., 2019). However, one can find out that various GRACE+GAD products agree with each other very well, where the differences between them are apparently within the uncertainty (1 sigma). Therefore, we suggest that CRA-LICOM has accepted accuracy for scientific applications without special caution, particularly for large-scale studies.

6 Limitations

Despite the satisfactory accuracy of CRA-LICOM for scientific application, there is still a non-negligible discrepancy between CRA-LICOM and the official GFZ-RL07 product. Although a part of the discrepancy can be attributed to an inevitable uncertainty of both the forcing field and the ocean model, we also recognize that the current version of CRA-LICOM has some potential limitations that need to be addressed here and considered in the next round of updates.

One major challenge is that the atmospheric forcing field employed, at its current version, has a coarser vertical and temporal resolution than ECMWF's latest reanalysis product used by GFZ-RL07. Therefore, multi-layer atmospheric reanalysis in terms of pressure level has been adopted for our atmospheric gravity modeling, which is likely not able to accurately reflect the upper air anomaly (Swenson and Wahr, 2002). Considering the fact that the impact of upper air anomaly is non-negligible, the model-level rather than pressure-level forcing field is recommended (Yang et al., 2021; Shihora et al., 2022). In addition, the sampling rate of our forcing field is only available for up to 6-hour, which restricts the number of feasible tides, for example, a major atmospheric tide S_2 (at a frequency of 3-hour) is not allowed for the insufficient sampling rate. Likewise, many other smaller atmospheric tides, as well as oceanic tides, are not estimated and removed from CRA-LICOM, while this has been done in GFZ-RL07. As a consequence, the deficiency in atmospheric and oceanic tides will eventually influence the non-tidal counterparts. Furthermore, the 6-hour resolution of CRA-40 may also limit the representation of high-frequency variations in OBP simulations compared to the 3-hour atmospheric forcing fields used in GFZ-RL07 products.

Another challenge comes from the complexity of ocean model (LICOM) configurations. While our model's native horizontal resolution (equivalent to approximately 1°, see Appendix A) is comparable to the MPIOM model (1° on average, see Dobslaw et al., 2017) used by GFZ-RL07, the resolution of our model (due to different grid strategy) appears insufficient to accurately simulate currents in western boundary regions, equatorial zones, and the Southern Ocean, where simulated OBP errors are relatively larger. The 30 vertical levels employed in LICOM are also inadequate to resolve the first baroclinic mode (Stewart et al., 2017). Furthermore, relatively coarse horizontal/vertical resolution can reduce the accuracy of ocean bottom topography, affecting the magnitude and spatial patterns of simulated OBP as well (Chen et al., 2023). In addition to the resolution, the ocean mask that defines the distribution of ocean and land should also be responsible for the biases observed in marginal seas between two products. In particular, the Black Sea and the Caspian Sea are defined as land areas in our current configuration as a result of their small sizes. Other differences in ocean masks include the Antarctic ice shelves and the Arctic Ocean coastal area (particularly near the Beaufort Sea), where we may need a more accurate definition of LICOM. Finally, we claim that the current LICOM configuration used in this study lacks tidal mixing and self-attraction and loading (SAL) feedback to ocean dynamics (Ray, 1998; Kuhlmann et al., 2011), which also contributes to simulation errors. Although nontidal OBP is the main

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focus, the potential interactions between general ocean circulation and tidal flow regimes are non-negligible and should be taken into account (Thomas et al., 2001; Li et al., 2015); and Ghobadi-Far et al. (2022) also emphasized that SAL significantly affects coastal regions and enclosed seas, such as the Gulf of Carpentaria. These model configurations are subject to the next round of updates.

7 Conclusions

We establish a new high-frequency atmospheric and oceanic gravity de-aliasing product, called CRA-LICOM, with a resolution of 6 hours and 50 km and a coverage of 2002-2024 at a global scale. Various inter-comparisons and validations confirm that CRA-LICOM can well represent Earth's high-frequency mass change and has sufficient accuracy to achieve the goal of de-aliasing for present satellite gravity missions. Specifically, we draw the conclusions as follows.

1. CRA-LICOM has confirmed that AO is the dominant source of high-frequency gravity signals (much larger than H), especially within the spectrum of aliasing, i.e., periods < 60 days (twice the monthly sampling rate of GRACE).
2. CRA-LICOM is generally consistent with the official GFZ-RL07 in terms of the dominating long-wave gravity signal, where a high temporal correlation (> 0.8) is found in the spectrum up to degree 15. Further spatial analysis confirms that the discrepancies are mainly within the aliasing spectrum (< 60 days), which challenges gravity recovery. However, the two products demonstrate improved long-term consistency, i.e., the global mean temporal correlation coefficient increases from 0.62 to 0.77, and the global mean RMSE decreases from 2.24 to 1.30 hPa when transitioning from periods < 60 days to > 60 days.
3. Inconsistency of atmospheric/oceanic tidal constituents between CRA-LICOM and GFZ-RL07 degrades the temporal correlation of their non-tidal counterparts. For better consistency, one must not add back the atmospheric tide S_2 for orbit determination or GRACE gravity recovery using CRA-LICOM in practice.
4. Validation of the ocean model confirms that LICOM effectively captures the ocean state and variability, including temperature (mean bias < 0.4%) and salinity (mean bias < 0.1%), across most regions. However, significant biases are observed in the North Atlantic, Southern Ocean, and along the western coasts of certain ocean basins, which likely contribute to the underestimation of OBP STD by LICOM, with values approximately 30% smaller than those of in situ observations. This also challenges the use of CRA-LICOM for scientific inference in these areas.
5. Temporal gravity recovery from GRACE using CRA-LICOM demonstrates a fairly high agreement (correlation coefficient > 0.9) with GRACE's latest official products from CSR, JPL, and GFZ. Independent validation with Argo and Altimetry further confirms the ability of CRA-LICOM in large-scale ocean applications and its consistency with other official gravity products.
6. As an independent product, CRA-LICOM could be a promising alternative to the official GFZ-RL07 product to be used in geoscience studies (GNSS, GRACE, and other geodetic techniques). In particular, a full-time-scale uncertainty could be

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produced through an inter-comparison of CRA-LICOM and GFZ-RL07, which could also be a valuable complementary to GFZ's uncertainty product (Shihora et al., 2024). A better understanding of the uncertainty of the AO is essential for improving the current GRACE (-FO) as well as the design of the next-generation satellite gravity mission.

8 Data availability

CRA-LICOM products are freely available at https://doi.org/10.11888/SolidEar.tpdc.302016 (Liu et al., 2025a). The products include Stokes coefficients for 6 hours (ATM, GLO, OBA, OCN), the corresponding monthly variables (GAA, GAB, GAC, and GAD), and the atmospheric tides.

Appendix A: Fundamentals of LICOM model

LICOM is a global general circulation model developed by LASG/IAP since the late 1980s (Zhang and Liang, 1989). LICOM3.0 is currently the ocean component of two climate system models participating in CMIP6 (Coupled Model Intercomparison Project Phase 6): the Flexible Global Ocean-Atmosphere-Land System Model version 3 with a finite-volume atmospheric model (FGOALS-f3; He et al., 2020) and the version with a grid-point atmospheric model (FGOALS-g3; Li et al., 2020). In this study, we employ LICOM3.0 coupled with the Community Ice Code version 4 (CICE4) through NCAR flux coupler 7 (Craig et al., 2011; Lin et al., 2016), previously used for two Ocean Model Intercomparison Project (OMIP) experiments (Lin et al., 2020), to simulate OBP.

LICOM3.0 employs an orthogonal curvilinear coordinate system and a tripolar grid with a resolution of about 100 km with two poles located on land in the Northern Hemisphere at (65°N, 65°E) and (65°N, 115°W), which addresses the singularity issue at the North Pole inherent in traditional longitude-latitude grids. The horizontal grid employs the Arakawa B grid system with a resolution of approximately 1°, while the vertical eta coordinate system comprises 30 levels. These levels have a thickness of 10 m in the upper 150 m, gradually increasing to 713 m near the ocean floor. The bathymetry of the model is derived from ETOPO2 bathymetry data (https://ngdc.noaa.gov/mgg/global/etopo2.html).

For equation discretization, the central difference advection scheme is applied in the momentum equations, while time integration uses the leapfrog method combined with a Robert filter. The tracer equations adopt a two-step shape-preserving advection scheme (Yu, 1994; Xiao, 2006) and semi-implicit vertical viscosity / diffusivity (Yu et al., 2018). The model computes the vertical viscosity and diffusivity coefficients using the scheme proposed by Canuto et al. (2001, 2002), while horizontal viscosity is represented using a Laplacian formulation, with coefficients set at 5400 $m\ s^{-2}$. To account for mesoscale eddy effects, LICOM3.0 employs the isopycnal tracer diffusion scheme of Redi (1982) and the eddy-induced tracer transport scheme of Gent and Mcwilliams (1990). In addition, the chlorophyll-a-dependent solar penetration scheme developed by Ohlmann (2003) is implemented.

Appendix B: Auxiliary experiments

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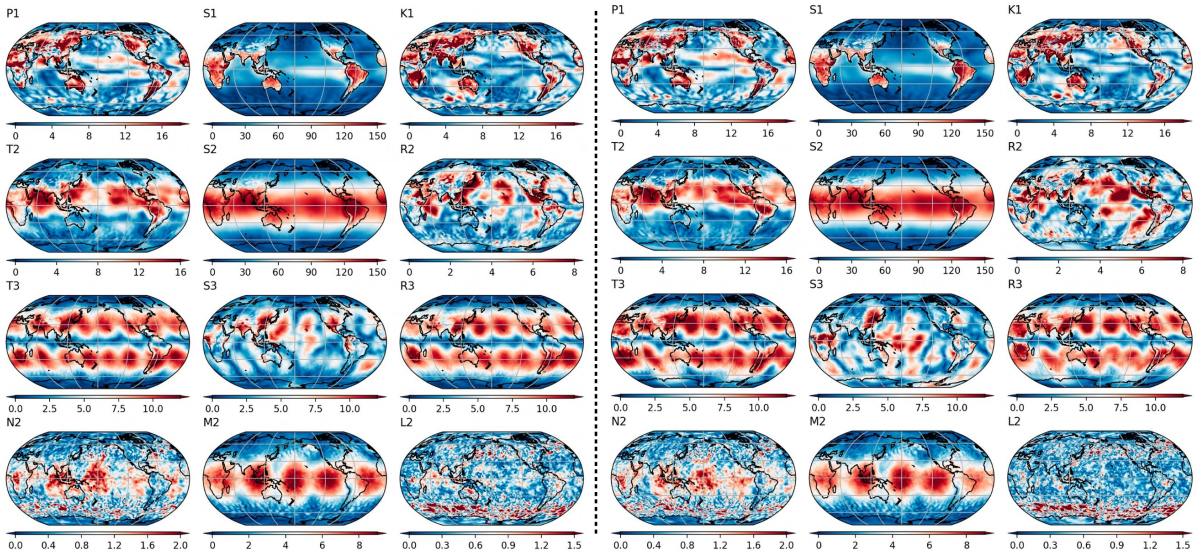

Figure B1. The left panel presents 12 atmospheric tides obtained from GFZ's official product, while the right panel presents our tides obtained from ECMWF-reanalysis (ERA-5) over the period 2007-2014. All tide lines are illustrated in terms of pressure amplitude [Pa]. See Dobslaw et al. (2016) for the definition of all tides.

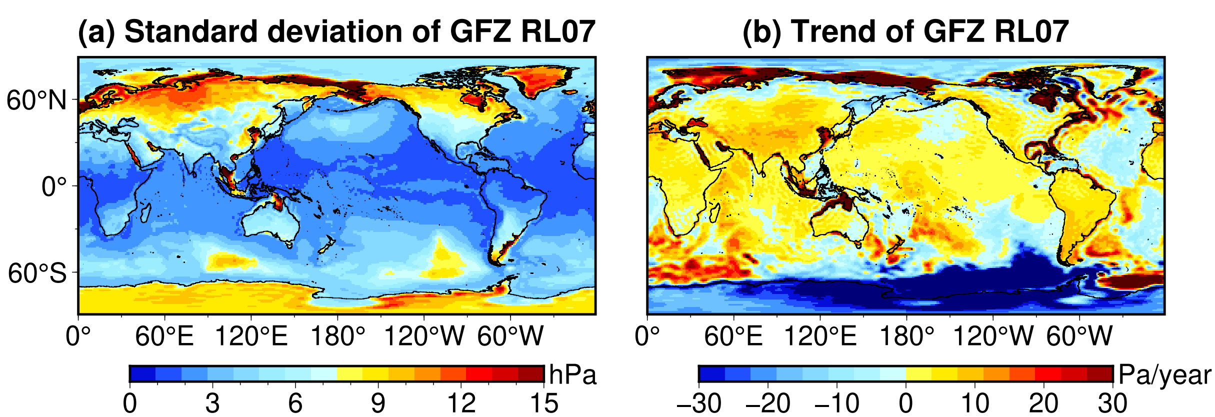

Figure B2. Equivalent pressure fields synthesized from GFZ-RL07 during 2002-present: (a) the standard deviation (hPa), (b) the secular trend (Pa/year).

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Competing interests. The authors have no competing interests to declare that are relevant to the content of this article

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