The Mediterranean Sea Heat and Mass Budgets : Estimates, Uncertainties and Perspectives

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Abstract

This paper presents a review of the state-of-the-art in understanding and quantification of the Mediterranean heat and mass (i.e. salt and water) budgets. The budgets are decomposed into a basin averaged surface component, lateral boundary components (through the Gibraltar and the Dardanelles Straits), a river input component and a content change component. An assessment of the different methods and observational products that have been used to quantify each of these components is presented. The values for the long term average of each component are also updated based on existing literature and a first estimate of heat fluxes associated with the riverine input has been produced. Special emphasis is put on the characterization of associated uncertainties and proposals for advancing current knowledge are presented for each budget component.

With the present knowledge of the different components, the Mediterranean budgets can be closed within the range of uncertainty. However, the uncertainty range remains relatively high for several terms, particularly the basin averaged surface heat fluxes. Consequently, the basin averaged heat budget remains more strongly constrained by the Strait of Gibraltar heat transport than by the surface heat flux. It is worth remarking that if a short (~few years) averaging period is used, then the heat content change must also be considered to constrain the heat budget. Concerning the water and salt fluxes, the highest uncertainties are found in the direct estimates of the Strait of Gibraltar water and salt transport. Therefore, the indirect estimate of those transports using the budget closure leads to smaller uncertainties than the estimates based on direct observations. Finally, estimates of Mediterranean heat and salt content trends are also reviewed. However, these cannot be improved through the indirect estimates due to the large temporal uncertainties associated to the surface fluxes and the fluxes

through Gibraltar. The consequences of these results for estimates of the Mediterranean temperature and salinity trends obtained from numerical modelling are also considered.

1- Introduction

Due to its relatively small size its semi-land locked nature, and its geographical location, the Mediterranean Sea is very sensitive and responds relatively quickly to natural forcing and/or anthropogenic influences. Demographic growth, climate change and overexploitation are exerting exceptional pressure on the Mediterranean environment, its ecosystems, services and resources. Over 150 million people live along its coasts and the basin is of paramount importance for the socio-economics of the surrounding countries. The Mediterranean Sea also hosts a large biodiversity (8% of the known marine species with only 0.8% of the oceanic surface) and is a basin of contrasting ecosystems: from the strongly oligotrophic deep interiors to the fully eutrophic northern Adriatic. Moreover, the Mediterranean region has been identified as one of the hot-spots for climate change in the world (Giorgi, 2006), where expected changes will be largest. Some of the projected changes for the next decades include the warming and salinization of the basin as well as the rise of sea level (e.g Adloff et al., 2015). These changes have been recently detected (e.g., Schroeder et al. 2016, Schroeder et al. 2017) and can potentially have a strong impact on the coastal environment and the marine ecosystems (e.g. Macias et al., 2015). The Mediterranean also plays a role in the evolution of the global ocean, mainly because of its character as a concentration basin. The small catchment area (i.e. relatively low river contribution), the permanent dry conditions in the south and the dry conditions in the north under cold and dry northerlies in winter contribute to evaporation exceeding the contribution of precipitation. Therefore, the basin acts as a source of salty waters for the global ocean. For instance, saline waters outflowing from the Strait of Gibraltar (SoG) are then found in the Atlantic Ocean at a depth of ~1,000 m and might have an impact on the global ocean circulation (Reid 1979; Béthoux 1979; Millot et al. 2006). In this context, the understanding of the functioning of the Mediterranean basin is of paramount importance in order to be able to accurately predict its evolution during the next decades, and its impact on the global ocean. Among all the elements of that functioning, one of the first issues that should be addressed is the monitoring and characterization of the heat and mass content evolution. Although previous attempts have been done (e.g. Lacombe et al., 1981 or Rohling and Bryden, 1992), the significant improvements of the observational systems over the last decades motivate a review of the present knowledge on the Mediterranean heat and mass budgets.

The variations of heat and mass content of the water in the Mediterranean basin (Figure 1) are directly related to the fluxes of heat and mass across the boundaries of the basin water body. Therefore, the heat budget equation for the Mediterranean can be written as follows:

(1)

where ρ is the density of seawater, θ the potential temperature, Cp the specific heat capacity. V_{Med} and A_{Med} represents the Mediterranean volume and surface area while s_{Gib} and s_{Dard} are the vertical section at the Gibraltar and Dardanelles Straits. u represents the velocity of the flow normal to those sections. QH_{net} refers to the net heat flux through the sea surface (in W/m^2) and Q_{riv} is the river runoff. The last sum is done over all the Mediterranean rivers.

The left hand side term in (1) represents the changes in the Mediterranean heat content (HC), while in the right hand side we have the heat flux entering into the basin through its boundaries: the air-sea interface (SHF), the Gibraltar Strait (GHF), the Dardanelles Strait (DHF) and the rivers (RHF), respectively (heat exchanges across the seafloor are ignored).

The mass budget could be expressed in similar terms. In this case it is convenient to split the mass content into two terms: the water and salt contributions. Thus, the equation for water content (WC) can be written like (1):

(2)

where in this case P and E respectively represent precipitation and evaporation (note again that water exchanges across the seafloor are assumed to be negligible in comparison with the other fluxes at basin scale). It is also worth noting that the heat and water budget equations are closely related as far as water and heat fluxes at surface and lateral boundaries are strongly related since an evaporation driven change in water flux necessarily implies a change in latent heat flux.

As for the salt content (SC), it can only be modified by the salt fluxes (SF) through the lateral boundaries, in particular through the Gibraltar and Dardanelles straits, since the atmosphere, the rivers or the bottom, do not provide significant salt inputs to the ocean. Therefore considering that S represents the salinity of sea water the salt equation will be reduced to:

(3)

The estimates of the single components of the heat and mass (water and salt) budgets using the above formulae require ‘good’ information on fluxes at the boundaries that are subject to large uncertainties, as will be shown in the present paper. Therefore, we begin by asking two questions : What is a ‘good’ flux? And why should we improve the estimates of those fluxes? Ultimately, a good flux is one that provides an accurate representation of the exchanges at the given boundaries on the space and timescales of interest irrespective of whether the underlying approach is observation or model. However, it is instructive to examine how the requirements of the observational and modelling communities can differ for particular applications. For instance, for modellers, air-sea fluxes should be compared with "observations" (*e.g*. from buoys)

at fine scales in order to reconstruct the seasonal cycle of some parameters (Sea Surface Temperature (SST), Mixed Layer Depth (MLD)). Good fluxes should permit the correct simulation of the thermohaline circulation, at daily timescales and with spatial resolution of the order of several tens of km, as well as the oceanic deep convection in order to simulate properly dense water mass formation. A correct estimate of net heat or freshwater fluxes, despite compensation errors, should prevent model drifts, even if sometimes it is difficult to know if such drifts exist or not due to the lack of reliable data for model validation (see Section 6). To enable the accurate simulation of some oceanic features, such as the eddies (20-30 km in diameter) observed in the Levantine basin, it is unclear how accurate the fluxes need to be in order to enable accurate simulation. Furthermore, for modellers, the question of timescales is important to solve: fluxes must at least enable reconstruction of the seasonal cycle and not only the annual mean state. Smaller timescales (weekly to daily) can also be required because strong flux events influence directly the formation of new water masses (e.g. deep water formation). In summary, it is very difficult to establish an absolute criterion to determine if a flux is 'good' in all the situations. Therefore, a better approach would be to quantify the uncertainties associated to each flux, so that each user may decide if the quality of the estimate fits the specific needs.

A joint analysis of the different components merging information from different sources could potentially lead to a significant improvement of the knowledge of the system as a whole. In this context, the project "**Med**iterranean **S**ea **ma**ss and **h**eat **b**udget: Understanding its forcing, uncertainties and time evolution (**MED-MaHb)"** has been initiated under the auspices of ENVIMED/MISTRALS/HYMEX. The principal objective of this initiative is to build up a multi-disciplinary synergic community for Mediterranean Sea climate research aiming to work on a re-evaluation of the mass and heat budgets. This work is based on the assessment of the consistency and uncertainty of the different components in equations (1-3). This assessment is based on different approaches (in situ measurements, remote sensing, ocean synthesis and numerical models) to further understand its forcing and to analyze its time evolution. This initiative follows the objectives of the 6th CLIVAR research focus (www.clivar.org/about/research-foci#six), MedCLIVAR (http://www.medclivar.eu/), HyMEX (www.hymex.org) and MedCORDEX (www.medcordex.eu/).

The goal of the present paper corresponds to the first major step of implementation of the MedMAHB project: to review the state-of-the-art of approaches and estimates of each component of the Mediterranean heat and mass budgets. Particular emphasis will be put on: (i) identifying current advances in research and on the characterization of the associated uncertainties, and (ii) to propose a set of recommendations for future improvements of those estimates. All this will in turn deliver relevant contribution to future advancements of regional climate monitoring and predictions, which have been identified as key challenges for future climate research studies (e.g. IPCC, 2013). In sections 2 to 5 we address the characterization of the fluxes through each one of the boundaries: ocean-atmosphere interface, the Strait of Gibraltar, the Dardanelles Strait and the riverine inputs, respectively. In section 6 the changes in the global heat, water and salt content are characterized, while in section 7 we present a short review of the projected evolution of the different components during the next decades. A general discussion is presented in section 8 prior to concluding with some final remarks.

2- Mass and heat exchange through ocean-atmosphere interactions

2.1. Overview

Air-sea exchanges of heat and freshwater in the Mediterranean are central to our understanding of its oceanography and role within the wider climate system. The exchanges are difficult to determine as they require knowledge of the latent (equivalently evaporation), sensible, longwave and shortwave heat fluxes, as well as precipitation. This in turn requires estimates of the driving near surface atmospheric meteorological parameters (air temperature, humidity and wind speed), sea surface temperature and cloud amount. Attempts to estimate these fields are severely limited by lack of observations and the subsequent derivation of the flux terms by incomplete knowledge of air-sea exchange physics. Consequently, considerable uncertainty remains in Mediterranean air-sea heat and freshwater exchange fields across a wide range of spatial and temporal scales. These vary from basin mean quantities at decadal timescales to, for example, extreme heat loss associated with winter storms at hourly to daily timescales and spatial scales of tens of kilometres.

To date, the main obstacle to producing accurate ('good') fluxes from various sources (satellite, numerical weather prediction (NWP) models) is the lack of reference quality flux datasets at key locations (for example, high quality time series from surface flux reference site moorings of the type deployed at several other locations in the global ocean within the Ocean Observatories Initiative). To our knowledge, some observatories do exist in the Mediterranean which collect relevant meteorological variables and radiative fluxes (see for instance the FixO3 network, http://www.fixo3.eu/observatory/) although they have not been used yet to evaluate different flux datasets. An alternative approach may be to undertake systematic tests that rely on forcing ocean models with all the available flux fields and compare ocean model results with oceanic data in order to ellucidate which forcing dataset is the best. However, this approach is expensive in computational terms and face the problem of errors in the ocean model (i.e. the discrepancies between model and oceanic observations may come from sources other than the forcing fields). One of the main techniques used by modellers is to apply flux correction in the form of restoring or adjustment. That is, to impose a constraint in the fields used to force the ocean models in which the modelled sea surface temperatureand/or salinity are relaxed towards a reference dataset. This approach has proven to be useful to avoid model drifts but can only be applied if the required reference datasets are of good quality and available for all the modelled period (i.e. this would not be possible for climate projections) . Moreover, an additional question is then to know if at long timescales these corrections affect processes such as convection, the production of dense waters and the global circulation itself and its variability (Simmons and Polyakov, 2004).

2.2. Methods for Estimating Surface Fluxes

Different methods are used to estimate turbulent fluxes (latent and sensible heat, evaporation, wind stress, CO_2 fluxes, etc ...). Local estimates are derived from turbulence measurements of the wind (3 components, with a three-axis sonic anemometer), air temperature and humidity

fluctuations at high frequency (20 to 50 Hz sample rate) in the constant flux layer. At sea, typical approaches are the eddy correlation (EC) and the inertial dissipative (ID) methods, that employ measurements from surface moorings (*e.g.* Weller *et al*., 2012; Bigorre *et al*., 2013) or research vessels (*e.g.* Weill *et al*., 2003; Bourras *et al*., 2009).

The EC method is based on direct calculation of the covariance between vertical wind and temperature or humidity fluctuations, and requires correction of wind, temperature and humidity measurements for ship movements (Edson *et al*., 1998). This means that the angular and linear accelerations of the ship have also to be measured at high frequency with a six degrees of freedom accelerometer. These additional parameters are not necessary for applying the ID method, which is based on the closure of the turbulent kinetic energy equation in the inertial zone of the wind spectrum. In both cases, however, it is essential to apply corrections to take into account flow distortion and vibrations of the measurement platform (Yelland *et al*., 2002; Dupuis *et al*., 2003; Bourras *et al*., 2009). In the best case, measurements have to be performed facing wind (at angles less than 30-40°) and at reduced ship velocity (2-3 kts).

Other methods for estimating turbulent fluxes at sea can also be used, such as: a) the profile method (derived from high resolution radio-soundings in the constant flux layer) (Dupuis *et al*., 1995) and b) the eddy-accumulation method (Brut *et al*., 2004), which is well suited for estimating latent heat fluxes, CO2 fluxes or for scalar fluxes of minor and slowly reacting atmospheric species (DMS, ozone, etc.).

The method most commonly used by modellers is the bulk method (Fairall *et al*., 1996, 2003; Weill *et al*., 2003; Edson *et al*., 2013; among many others). It is based on the physical properties of the constant flux layer (*cf*. Monin and Obukhov, 1954). The method considers the existence of relationships between turbulent fluxes and "slow" (opposed to turbulent) meteorological and oceanic parameters (*i.e.* wind, temperature, humidity, sea level pressure and SST). The exchange coefficients connecting the fluxes and the slow parameters are determined from local estimates of turbulent fluxes using the ID or EC methods.

In general, experiments at sea only sample a limited range of conditions (*i.e.* medium wind speeds from 3-5 m/s to 15 m/s; weakly stable stratification to moderate unstable stratification). In these ranges, relatively good agreement exists between various parameterizations found in the literature. However, large disagreements exist at low and high winds and highly unstable stratification, which are largely undersampled (Figure 2). Moreover, the sensible heat flux transfer coefficient is poorly known: these restrictions limit the use of bulk algorithms. In numerical models, bulk formulae are usually extrapolated beyond the conditions under which in-situ estimates are made, and this results in significant spread between flux evaluations (see example given in Figure 3).

Some physical processes directly affect the estimates of fluxes and several bulk algorithms include sea surface properties such as surface wave characteristics, sea spray, white capping, stability, gustiness factor, the effects of skin temperatures, Webb correction, and cooling due to rain (*e.g*. Fairall *et al*., 2003). These effects are generally weak except in some specific cases. Of course, these additional parameterizations (as they depend on an increased number of basic

observables) increase the errors in the fluxes themselves. Note that for the Mediterranean Sea, frequent gale force (>20 m/s) winds from September to May, and related young sea state and sea spray (Andreas *et al*., 2014) need to be taken in account in the estimation of wind stress and latent heat fluxes.

Besides local flux estimates, fluxes can potentially be retrieved from satellite observations. From brightness temperature measurements in a particular frequency range, it is possible to estimate both the shortwave and longwave radiative fluxes (Brisson *et al*., 1994; Schluessel *et al*., 1995). Produced operationally at the Centre de Météorologie Spatial (CMS, Météo-France) and at fine time and space scales (1h and 0.1° longitude and latitude), longwave and shortwave radiations are compared systematically with local measurements (*e.g*. buoys in the Atlantic) (Le Borgne et al., 2007). As reported in Table 1 and reference herein, the comparison of daily satellite retrievals compare quite well with ship and mooring data in the Atlantic, with bias less than 2 W/m² and RMS ranging from to 10 to 20W/m², except for some specific cases (in presence of smoke aerosols or Saharan dust aerosols) . Unfortunately, a similar comparison in the Mediterranean has not been yet done.

SSTs are nowadays retrieved as a blend of satellite microwave sensors. They perform quite well even through cloudy conditions and infrared observations provide accurate SST analyzed fields (Donlon *et al*., 2002, 2007). For instance the accuracy of ENVISAT radiometer is in the range of 0.1-0.3°C (Embury *et al*., 2012) and systematically compared with in-situ measurements, SEVERI has a bias of -0.16°C with a root mean square error of 0.5°C (Marullo et al., 2007). Like for radiative fluxes, limitations also do exist in the presence of aerosols.

Wind stress is currently estimated from scatterometry which provide the wind relative to surface currents at spatial scales of the order of 10-15 km. Sampling problems exist with these retrievals since the time interval between two passages of the satellite at the same spot is of the order of some days and no data are available less than 30 km from the coast. Other limitations exist at weak wind speeds (less than 3 m/s) and for high winds (*e.g.* in tropical cyclones), as well as in presence of strong surface currents, of rain (Weissman *et al*., 2012), or due to the influence of waves on the stress. Moreover, the space resolution along oceanic fronts and mesoscale features (eddies, filaments) is generally too coarse.

For retrieving turbulent fluxes from satellite, one requires near-surface wind, air temperature, humidity, SST and sea level pressure with the best accuracy as well as the use of a state-of-art bulk algorithm (*e.g.* Bourras, 2006). Several attempts have been performed to get turbulent flux fields from satellite (Chou *et al*., 2003) or a net heat balance. Nevertheless the determination of near surface moisture and air temperature still need to be improved in order to get reliable turbulent fluxes (Bourassa *et al.*, 2010; Kent *et al*., 2014).

Estimating rain is clearly important for the Mediterranean Sea. However, precipitation is difficult to measure over oceans and seas, both because of instrumental errors (flow distortion on the sensor, difficulty in differentiating sea spray and precipitation at high wind, etc..), and of sampling errors due to the representativeness of local measurement with the surroundings. This problem is more complicated in the case of convective precipitation, due to the highly

patchy and intermittent nature of rain. Moreover, precipitation measurements at sea are very scarce. All these factors contribute to the difficulty in evaluating satellite retrievals of oceanic precipitation or numerical model outputs or reanalyses with in-situ data. Satellite retrievals face the question of the temporal sampling window (e.g. where the scanned area is limited to once or twice a day sampling), a problem exacerbated in the case of convective precipitation. Some products combine satellite datasets with reanalyses (Xie and Arkin, 1996) although important biases remain in reanalyses (*e.g.* Serreze *et al*., 2005). Additionally some observational products based on in-situ reports from ships have been produced (e.g. ICOADS, http://icoads.noaa.gov/). However, these products suffer from the scarcity in time and space of the observations, so the sampling error can be very large. A complementary source of information can be the precipitation estimates from coastal radar measurements. At present many of the Mediterranean countries have radars that measure precipitation till 50-150 km from the coast (e.g. http://www.eumetnet.eu/radar-network). Although the whole basin is not completely covered by the radars they can provide useful information for the calibration/validation of other precipitation datasets. The main difficulty is that the data is managed by many different providers and there is not a centralized database where all the radar data is unified and homogenized. Furthermore, the radar measures reflectivity which has to be converted to precipitation intensity relies on more or less complex relationships that can induce uncertainties.

Finally, flux fields can be provided from operational NWP model outputs (e.g. ECMWF, NCEP, ARPEGE...). However, models have errors related to parameterizations used in the atmospheric boundary layer (including turbulent flux bulk methods), and to radiative schemes which affect cloud cover and radiative fluxes. Moreover, microphysical schemes and formation of clouds, rain bands etc. affect the estimate of precipitation and thus the fresh-water balance at the sea surface. An example of the uncertainties introduced by all those factors is the large discrepancies in the mean heat and water flux over the Mediterranean shown by different regional climate models (RCMs) even when all are driven by the same global reanalysis. For instance, Sánchez-Gómez et al. (2011) analysed an ensemble of atmospheric RCMs driven by the ERA40 reanalysis. While the long term mean of the surface heat flux in ERA40 was -15 W/m^2, the RCMs driven by ERA40 provided values ranging from -40 to +12 W/m^2. For the net surface water flux the results where similar: the ERA40 mean was 781 mm/yr and the RCMs results ranged from 602 to 1087 mm/yr.

2.3. Reliability of Surface Flux Estimates

Surface fluxes estimated by the ID or EC methods are generally compared to assess which method gives better results. For example, from the AMMA/EGEE campaign, Bourras *et al*. (2009) gets weak biases between the ID and EC wind stress and sensible heat fluxes (<2 $W.m^{-2}$) but larger biases affect latent heat fluxes when humidity fluctuations were measured with different instruments (18 Wm^{-2} when evaluated with a LICOR and 33 Wm^{-2} when evaluated with a refractometer). These results show that the evaluation of exchange coefficients is affected by some errors: such errors have thus to be evaluated when using a bulk algorithm.

Errors in turbulent bulk fluxes result from: 1. the accuracy of the input observables (humidity, wind velocity, temperature, SST, sea level pressure); 2. the bulk algorithm itself, which corresponds to the uncertainty affecting the exchange coefficients; 3. the iterative method used, since most bulk algorithms solve the constant flux layer equations iteratively until convergence. Generally point 2 and point 3 errors are computed as a percentage of the mean flux values.

Moreover, observable errors increase in cases of undersampling, for instance associated with diurnal cycles, when aggregating fluxes over heterogeneous areas etc. Note that the Mediterranean Sea experiences frequent summer SST oscillations associated with diurnal cycles of the order of 0.5°C (Stuart-Menteth *et al*., 2003; Marullo *et al*., 2014), a phenomenon largely increased under weak winds and strong diurnal heating conditions. The difference between bulk and skin SST may induce a difference of 0.2° C at night and several degrees during the day (Gentemann and Minett, 2008). According to Webster and Lukas (1992), a 1°C skin SST variation affects upward longwave radiation, sensible heat and latent heat fluxes by 1.3%, 23.3% and 16.2% respectively.

In Table 2, the expected errors are reported for the fluxes estimated at the LION buoy in the north-western Mediterranean Sea. The figures were obtained by taking the mean values of the observables and fluxes estimated with the COARE3.0 bulk algorithm at the buoy for the period 2011-2013. The error on the net heat flux mostly comes from the errors on the latent heat flux (12 W/m^2) and the net longwave radiation (7 W/m^2). The error on the net heat flux reaches 15 W/m^2, a figure close to those reported by Colbo and Weller (2009) in the sub-tropics and or by Cronin *et al*. (2014) for the tropical Pacific. When the bulk algorithm error is considered, the total error on the net heat flux rises to 20 W/m^2. It is important to notice that this estimate of the heat flux uncertainties has been obtained at the LION buoy and is not representative of the whole Mediterranean. Unfortunately, up to now it is not possible to directly assess the uncertainties associated with basin averaged fluxes because mooring observations are restricted to a limited number of sites that do not adequately sample the range of conditions across the basin.

Concerning the water fluxes various studies show that the bias between satellite retrievals of precipitation and rain gauge can be significant (up to 25% even for long time series, e.g. Bowman et al., 2009). Also Romanou et al. (2010) have estimated the uncertainties in the annual mean evaporation over the Mediterranean provided by HOAPS-3 to be ~20-35 % of annual estimate. There are no studies quantifying the uncertainty of products based on ship observations and only a comparison among products is possible. In particular, the different in-situ observational products analyzed by Mariotti et al. (2002) provide a range for the long term mean of precipitation (evaporation) of 310-700 mm/yr (1120-1570 mm/yr). The discrepancies among products can be due to the data included in the product as well as to the interpolation procedure used to generate the basin averages. Regarding the numerical models and reanalyses, precipitation depend largely on the dynamics (humidity transport) and on the parameterizations of warm and/or cold microphysics and radiative transfer, leading to significant biases in precipitation positioning and intensity in comparison with satellite datasets (Xie and Arkin, 1996; Béranger et al., 2006; Pfeifroth et al., 2013). As an example, the Mediterranean average

precipitation in the reanalyses analyzed by Mariotti et al. (2002) ranged from 300 to 500 mm/yr while the average evaporation ranged from 920 to 1200 mm/yr.

2.4. Evaluation of Flux Estimates

Evaluations of surface flux datasets rely on both large scale (e.g. budget closure, see Discussion section) and local (i.e. high quality surface flux reference site measurements) constraints (Josey and Smith, 2006). At local scales, some flux datasets exist that have gathered high quality data measured at sea and can potentially be used for model and satellite flux product evaluation (e.g. AUTOFLUX, Larsen *et al*., 2000; ALBATROS, Weill *et al*., 2003; SAMOS, Smith 2004). For the Mediterranean, the Fetch experiment specifically documented fluxes under low winds and under gale force winds in the Gulf of Lion, as well as associated sea states (Hauser et al., 2003). Surface flux reference site data from flux moorings at several locations in the Mediterranean Sea would clearly be highly desirable to complement those datasets. To our knowledge, only a very limited amount of data exist for the Mediterranean Sea, at periods of time of the order of one year or longer, which are necessary for evaluation of flux products. One candidate of such dataset is the network of Mediterranean buoys that is currently operated by different institutions. A description of most of the existing sites can be found in the FIX03 project website (http://www.fixo3.eu/observatory/). Most of the buoys do not measure precipitation, radiative fluxes or the parameters required to derive the turbulent fluxes but there are some exceptions (see Figure 1). The Lion and the Azur buoys operated by Météo-France (http://mistrals.sedoo.fr/HyMeX/) provide subdaily observations of those parameters since year 2000 (radiative fluxes after 2011). The W1-M3A buoy, located in the Ligurian Sea and operated by CNR (http://www.odas.ge.issia.cnr.it), have been measuring air-sea fluxes (turbulent and radiative fluxes and precipitation) at subdaily frequency since 2009 (Canepa et al., 2015). The E2-M3A buoy, located in the South Adriatic and operated by the OGS measure the same parameters except precipitation since 2014. The E1-M3A site, managed by HCMR (http://www.poseidon.hcmr.gr/) have also observations of the relevant parameters since 2009, although with large gaps in some of the parameters. Finally, the OOCS observatory operated by the CEAB-CSIC in the Catalan Sea (http://www2.ceab.csic.es/oceans/) also provides the same parameters (except precipitation) since 2009. Although most of the data is publicly available, they should be quality-controlled and homogeneized before they can be used as a reference dataset. Moreover, it has to be noticed that most of the observations are located in the northwestern Mediterranean, which is a problem in terms of the representativity of the reference dataset. Therefore, having a continuous monitoring of air-sea fluxes in the central and southern parts of the basin would be advisable

Additionally the measurements of atmospheric data from vessels (see for instance http://icoads.noaa.gov/) could be a complementary source of information. As mentioned above, the sparse spatial and temporal coverage of those data pose a serious problem for the estimate of basin average quantities. However, that dataset could be used to assess the quality of the products at the time the observations were taken, and thus identify which product (if any) is more reliable. Finally, radar observations of precipitation could be used to assess the quality of the products. However, as mentioned above, this is not straightforward as the reflectivity data

provided by different radars should be first homogenized and transformed into precipitation intensity.

A complementary approach is to use the budget closure at basin or subbasin scales for closing the observed heat and water budgets (Caniaux *et al*., 2005a, 2005b). The results obtained from this inverse method over the north-western Mediterranean basin are encouraging and the adjusted flux fields, including adjusted precipitation, could potentially be used as a reference for evaluation, model forcing or process studies (Caniaux et al., 2017).

The heat transport observations at the Strait of Gibraltar provide useful constraints on the basin mean averaged fluxes that have been the subject of many studies over the years (e.g. Gilman and Garret, 1994; Sanchez-Gomez et al, 2011; Schroeder et al., 2012). A common approach in those studies has been to use the results of Macdonald et al. (1994) who determined, from water temperature measurements at a few points of the water column across the Strait, a value for the constraint on the basin mean heat flux of –5.2+/-1.3 W/m^2 (where negative values imply an ocean heat loss). It is important to remark, though, that Macdonald et al. (1994) used less than one year of observations and did not use observations in the upper 100 m of the water column. Therefore, as acknowledged by those authors, the uncertainty provided is probably underestimated. Historically, observation based flux datasets have not been able to satisfy this constraint. Instead they have basin mean values typically in the range 20 to 30 W/m^2 (e.g. Gilman and Garrett, 1994). However, this has been recognised as being due in part to the use of radiative flux formulae that have been developed in large scale ocean basins (e.g. North Atlantic) and thus do not reflect conditions typically experienced in the Mediterranean. Consequently, various causes for this imbalance have been proposed including the neglected influence of aerosols on the shortwave radiation and the need for revised longwave parameterisations specific to the Mediterranean (Bignami et al., 1995).

Even with reasonable adjustments for these processes in place, the basin mean heat flux with the NOC ship based product is about 5 W/m^2 i.e. a positive bias relative to the Gibraltar based constraint of the order of 10 W/m^2 (Sanchez-Gomez et al., 2011). By combining the NOC based radiative fluxes with satellite based (HOAPS) estimates for the turbulent fluxes these authors were able to obtain a more realistic basin mean value of - 3 ± 8 W/m^2 for which closure with the Macdonald et al. (1994) value is obtained within the stated error bounds. Sanchez-Gomez et al. (2011) also considered budget closure across a range of regional climate models and find a large range but a mean value that does satisfy the Gibraltar constraint; for the ensemble of models considered the basin mean net heat flux is -7 $\pm$ 21 W/m^2 datasets. The large spread indicates the need for substantial further effort to reduce model uncertainty but such models may ultimately provide a useful indicator of the processes responsible for the biases in observation based Mediterranean Sea air-sea flux datasets.

2.5. Perspectives

Based on the published results cited above, the NOC/HOAPS estimate from Sánchez-Gómez et al. (2011) of the basin average Mediterranean surface heat flux of -3 $\pm$ 8 W/m^2 is closest to the Gibraltar based value of –5.2 $\pm$ 1.3 W/m^2 (Macdonald et al. , 1994). In a further study, Pettenuzzo et al. (2010) attempt to use various satellite and in situ observational data to construct spatially

varying corrections to the ERA-40 reanalysis air-sea fluxes and obtain a climatological mean net surface heat flux of −7 W/m^2 (note they do not provide an uncertainty range) which is also consistent with the Gibraltar based value. We take both the Sánchez-Gómez et al. (2011) and Pettenuzzo et al. (2010) values to be indicative of plausible surface flux based estimates of the basin mean heat budget in the discussion in Section 8. It should be stressed that there is no single 'best estimate' set of fluxes for the Mediterranean, hence our decision to select these two papers as a reflection of recent efforts and ongoing uncertainty in the surface flux fields. As these estimates have been obtained with knowledge of the Strait of Gibraltar constraint, it is not useful to employ the surface flux estimates to indirectly estimate the other heat budget components. Regarding the surface water flux the basin average is estimated to be -750 $\pm$ 190 mm/yr. The uncertainty here has been estimated as the spread of the different products analyzed in Mariotti et al., (2002) and Sánchez-Gómez et al. (2011).

At present different products based on observational and modelling approaches are available for the Mediterranean air-sea fluxes. Qualitatively, the strengths and weakness of each one is relatively well known but the bottleneck that prevents the improvement of such products and a more accurate estimate of the associated uncertainties is the lack of reference in-situ datasets. The gathering and homogenization of all the above mentioned datasets could lead to such reference dataset and eventually to an improved estimate of the surface fluxes.

3 - Exchanges through the Strait of Gibraltar

3.1 Overview

The vigorous water exchange through the Strait of Gibraltar (SoG hereinafter) is an issue of compelled analysis in the study of the mass and energy budgets of the Mediterranean Sea, since it is its most important open lateral boundary and the only connection to the global ocean. Not surprisingly, its study has aroused great interest and much effort has been paid to put reliable figures to the size of the exchanged flows.

The highly irregular topography of the SoG (Figure 4), and the remarkable strength of the tidal currents interacting with this topography pose severe difficulties for estimating the size of the exchanged flows and their variability (Lacombe and Richez, 1982; Armi and Farmer, 1988; Bryden et al., 1994; García Lafuente et al., 2000; Sannino et al., 2004; Sánchez Román et al., 2009; Sannino et al. 2009; Sánchez-Garrido et al., 2011; Sannino et al., 2015). The attempts to provide a reliable estimate are traced back to the beginning of the 20th century (see Table 3) and can be sorted out into three categories. In the first one are the theoretical approaches based on the volume and salt conservation for the Mediterranean, which require the previous knowledge of the net evaporation over the basin (or at least a realistic guess of it) as well as the salinity difference between the connected water bodies. The weakness of this approach is the above-mentioned poorly constrained values of the evaporation . More sophisticated approaches include energy considerations that bring the problem to the field of the hydraulics (Bryden and Stommel, 1984; Armi and Farmer, 1985; 1988; Bryden and Kinder, 1991). The second category gathers the attempts based on observations, whose history is more recent due to the formidable

technical challenge of deploying scientific equipment in such a harsh environment. The third one is the numerical modelling (Sannino et al., 2004; 2007; 2009; 2014a; Sánchez Garrido et al., 2011; Peliz et al., 2013; Boutov et al., 2014) whose usefulness is arguable as models require feedback from observations to be calibrated. Table 3 shows the values provided by the first two categories and illustrates how the theoretically-based estimations are systematically larger than those based on observations. Estimations from numerical models are not included in Table 3 because of their reliance on the observations used for calibration. However, numerical models are highly valuable to address other flow properties and to draw important conclusions, as it will be shown below. Table 3 also provides an estimation of the net flow through the SoG as the mean of the values based on observations (no-shaded rows), which is 0.065Sv (1Sv = $10^6 m^3/s$). Taking the standard deviation of these values as the accuracy, then the net flow through the SoG from historical observations would be Q_{net}=0.065±0.033 Sv, which could be rewritten as Q_{net}=0.82±0.42 m/yr in terms of the rate of net water loss (E-P) in the Mediterranean basin (1 Sv across the SoG is equivalent to 1260 mm/yr of evaporation in the basin).

3.2 Recent estimations of the exchanged fluxes from observations

The volume transport, or flow, through a given section is defined by:

(4)

where subscript *i* is for Atlantic or Mediterranean flows, u_i is the water velocity (considered positive to the east, negative to the west), and the integral is extended to the cross area of the section from the lower (*LS*) to the upper (*US*) limit. These limits are the seafloor and the interface for the Mediterranean water (*i=M*), and the interface and the free surface for the Atlantic water (*i=A*). The net flow $Q_{net}=Q_A+Q_M$ is a small quantity achieved as the sum (actually the difference, since Q_M is negative) of two much greater numbers, making it very sensitive to the accuracy of the summands. This is a relevant issue regarding the computation of net fluxes through the SoG.

The final figure provided by Sammartino et al. (2015) for the outflow at the westernmost section of the SoG (Espartel sill section, ES, see Figure 4) using this approach was -0.85 Sv. The outflow variability has to be considered at different time scales. It is dominated by tidal fluctuations whose standard deviation at ES is ± 0.39 Sv, quite close to the amplitude of the M2 tidal constituent over there. For long-term studies, in which tidal variability is filtered out, this interval is unrealistic. The main remaining source of variability after removing tides stems from the atmospheric forcing and shows a standard deviation of 0.15 Sv, less than half the previous value (Figure 5). For annual time scales, the meteorologically-driven fluctuations are averaged out and the variability is further reduced (± 0.08 Sv), the seasonal cycle prevailing over the rest of contributions. It shows maximum outflow in late winter-early spring and minimum outflow in late summer (Figure 6). At even longer time scales, the interannual variability drops to ± 0.03 Sv, which is the uncertainty assigned by Sammartino et al. (2015) to their long-term estimation of

the outflow at ES based on ten years of observations. A very small trend of -4.6×10^{-4} Sv/yr was reported by these authors, although it was not significant at the 95% confidence level.

The conversion from volume to mass transport is easy since the Mediterranean water density variations around a central (3D spatially-averaged) value of ρ_0=1028.5 kg m^{-3} in the SoG are smaller than 1 kg m^{-3}. It gives a relative variation ~40 times less than the relative uncertainty in the estimated flows and it can be assumed as constant. With that value for ρ_0, the mass outflow at ES would be $(8.74\pm0.31)\times10^8$ kg s^{-1}. Because of this straightforward conversion, we will only speak of volume transport (flows).

Heat (*QH*) and salt (*QS*) transport are computed similarly to the flows in eq. (4):

$$\rho \tag{5}$$

(6)

where subscript *i* is again for layers, ρ_i is the density, ϑ_i and S_i are the potential temperature and salinity, respectively, and $c_{p,i}$ the specific heat of seawater at constant pressure, which depends very slightly on ϑ_i and S_i. The units for QH_i and QS_i are W and kg s^{-1}, respectively. The net heat and salt transport is the sum of the inflow and outflow contributions. Notice that the negative sign of u_M implies negative heat and salt Mediterranean transports (QH_M, QS_M).

The practical problem to compute these flows from observations is the necessity of having a good representation of the spatial distribution of ϑ and *S*. First, there is the need of sampling the water column to have vertical profiles of ϑ and *S*, due to their strong dependence with depth in the SoG. The same practical problem exists for the flow computation using eq. (4), but this problem can be sorted out by a bottom-moored uplooking ADCP, as it is the case in the ES monitoring station. The horizontal variability could be addressed in a similar way as in *Sammartino et al*. (2015) for the outflow computation if a time series of a single vertical profile of ϑ and *S* were available, which is not the case. Presently, such time series do not exist and the only feasible estimations of heat or salt transports based on observations come from simultaneous measurements of the *u*, ϑ and *S* fields collected by CTD probes with ADCP installed on it during oceanographic surveys. The resulting estimation has the important drawback of being strongly contaminated by the time variability of the velocity field. Figure 7 is a clear illustration of this situation for the case of heat transport. None of the displayed estimations are even close to the expected mean value. Yet more, the estimations carried out during the same oceanographic cruise at the two sampled sections differ noticeably because the sections were sampled at different moments of the tidal phase. The same is applicable to the salt transport, since the main source of variability in eq. (4) to (6) is u_i, especially at tidal time scales. With this fact in mind, the best way of addressing heat and salt transports nowadays, or at least some aspects of them, is numerical modelling.

3.3 The exchanged fluxes from numerical models

Numerical models offer useful insight on relevant features of the exchange hydrodynamics, although the accuracy of the size of the flows computed from their outputs must be considered cautiously. Different factors affect the reliability of numerical models of the SoG, which have to deal with an unusually high spatial and temporal variability and short-scale pronounced bathymetric features. Tidal forcing must be explicitly included because the highly non-linear dynamics caused by the strong flow-topography interaction produces a tidally-driven increase of the long-term exchange (Bryden et al, 1994; García Lafuente et al., 2000; Sannino et al., 2004). A good assessment of the effect of this topographic control on the water flow requires a precise bathymetry of the model domain associated with a very high spatial resolution in both horizontal and vertical directions. From a intercomparison modeling study, Sannino et al., (2014a) were able to assess the minimum requirements needed by an ocean model to reproduce a reliable hydraulic regime of the exchange through the SoG. The requirements suggested are: a minimum horizontal resolution of 500 m and a minimum vertical resolution of about 10 meters. Moreover, while the non-hydrostatic feature is crucial for the generation and propagation of the well-known internal bore and associated large-amplitude internal waves, it is not strictly necessary for simulating the hydraulic regime. Such a very high resolution demands expensive computational systems, which finally provide global figures and spatial patterns that would be impossible to infer from the limited pool of observations. For instance, they resolve the cross-strait structure of the flow, a fact employed by Sánchez Román et al. (2009) and Sammartino et al, (2015) to refine the outflow estimated from observations at ES using the vertical profiles of velocity collected at a single point of the section.

Two hindcasts produced by the Massachusetts Institute of Technology general circulation model (MITgcm) model adapted to the SoG domain (see inset in Figure 4) are now applied to estimate heat and salt transports. Details about the model initialization and validation can be found in Sanchez-Garrido et al. (2011, 2013) and Sammartino et al. (2014). The first hindcast covers the period from March 12 to June 24, 2011, and the second one from September 1 to November 17, 2011.

Table 4 summarizes results from these hindcasts, which are discussed next. A first topic to deal with is the change of the inflow and outflow from one section of the SoG to the other (columns Q_A and Q_M), rising the question of which value is the best one, if any. The classical salt ($Q_M S_M + Q_A S_A = 0$) and volume ($Q_A + Q_M = E$) conservation relationships, valid for steady state exchange and often used to get estimations of the flows (see caption of Table 3), can be combined to give $(S_M - S_A) \times Q_M = S_A \times E$. Since S_A and E are constants, the product $\Delta S \times Q_M$ ($\Delta S = S_M - S_A$) must be constant too, a result pointed out by Bryden et al. (1994), who termed it Outflow Salinity Transport (*OST*). They remarked the fact that a value of the outflow is not meaningful unless it is accompanied by the salinity difference under which the exchange is computed.

In the SoG this subtlety minds because the entrainment that the active layer[1] exerts on the passive layer increases the exchanged flows (García Lafuente et al., 2013) and diminishes their averaged salinity difference with regards to what happens in Camarinal Sill (CS hereinafter). Last column of Table 4 confirms that the *OST* is the same in all sections in both runs with an accuracy greater than 2% (*OST* =–1.81 Sv PSU), which is not true for the outflow. The discussion above also suggests that the reference flows should be those measured in CS because it is the section of maximum salinity contrast, which implies that it is the section where the entrainment influence is the least. And, therefore, it shall be the section with the minimum size of the exchanged flows. This result is of practical interest for the numerical models that do not resolve the SoG domain properly. Such models are computing flows that correspond to those at CS in our discussion, as they do not resolve the entrainment. According to the results in Table 4, the outflow at ES should be reduced by 12% to obtain the representative value at CS. This percentage is greater than the 5% put forward by García Lafuente et al. (2011) based on near bottom salinity observations at ES and CS. With a 12% reduction, the outflow at CS during the period analyzed in Sammartino et al. (2015) should be -0.77±0.03Sv. The same considerations apply to the inflow Q_A in Table 4, but not to the net flow, that must be constant throughout all the sections of the SoG in each hindcast. Column Q_{net} in Table 4 suggests that, in fact, it is.

In the light of this discussion, the question then arises as to why the flows are not observed at CS if it is the most representative location? The answer may be found in Table 5, which shows the eddy-flux contribution to all the transports during hindcast II. Eddy fluxes have their origin in the positive correlation between tidal currents and interface vertical oscillations. They increase the long-term exchange (Bryden et al., 1994; Vargas et al., 2006; Sannino et al., 2004, 2007, 2015; Naranjo et al., 2014) and modify the way in which the exchange is achieved. The eddy-fluxes contributions to the long-term volume, heat and salt transports depend on the section, being maxima at CS where it represents between 35% and 40% of the total transport in hindcast II (Table 5; similar outcome is produced in hindcast I -not shown), which agrees well with the findings of Bryden et al., (1994). They are greatly reduced at the ending sections of the SoG. Considering the difficulty to estimate eddy-fluxes from observed ADCP velocity profiles, and that their estimations are affected by noticeable uncertainty, it is concluded that CS is not well-suited to observe the outflow. ES gathers more appropriate conditions to this aim, whereas the inflow is better monitored near the eastern exit of the SoG.

Figure 8 shows a fragment of the heat and salt transports produced by the numerical model and helps to explain the large spreading of the observation-based transports displayed in Figure 7. It confirms that the transports are fully dominated by the tidal variability, a result even more evident in CS section (not shown), where the transports change sign periodically within each layer due to the flooding of the hydraulic control. At ES, the transport within the Mediterranean layer is always negative except for very short and sporadic moments, while the change of sign is the rule for the Atlantic and, hence, the net transports.

[1] We shall refer to as *active* and *passive* the layers that flow faster and slower, respectively. West of Camarinal sill, the *active* and *passive* layers are the Mediterranean and Atlantic layers, respectively. The opposite happens to the east of Camarinal sill where the *active* layer is the Atlantic one.

The mean values of heat and salt transports in both hindcasts at the selected sections of the SoG are shown in Table 4. An uncertainty has been ascribed to each mean, which has been estimated as half the standard deviation of the subinertial time series, in agreement with the analysis carried out in Sammartino et al. (2015) [2].

Within the uncertainty interval, the net salt transport (QS_{net}= -1.5 $\pm$ 6.5 10^6 kg s^{-1}) is null. Actually, it is slightly negative, suggesting a prevalence of the Mediterranean layer, but the mean of the net transport is less than one order of magnitude the transport within each layer. It could also have been the result of an enhanced negative barotropic fluctuation that affected the hindcast periods. Whatever the cause, the assigned uncertainty, which is 3 to 7 times greater than the mean, suggests that the salt transport is balanced or, if it is not, the numerical simulation cannot give an accurate figure for the imbalance. In other words, the classical salt conservation relationship ($Q_M S_M + Q_A S_A = 0$) would hold.

The net heat transport is towards the Mediterranean Sea with a mean value greater than the associated uncertainty and almost identical numerical values of ~11.5 TW in hindcast I and ~10.5 TW in hindcast II in the three sections (Table 4). Dividing the heat transport by the area of the Mediterranean Sea (2.5 10^{12} m^2), we obtain the equivalent heat flux through the Mediterranean Sea surface required to balance the heat transport across the SoG. The resulting figures are 4.6 W m^{-2} and 4.2 W m^{-2} with a similar uncertainty of ~4 W m^{-2}. They are in good agreement with the only estimate carried out from the already mentioned direct temperature observations at the SoG (Macdonald et al., 1994) and with reported values in the literature based on Mediterranean Sea heat budgets (Bunker et al., 1982; Garrett et al., 1993).

3.4 Perspectives

Estimations of the size of the exchanged flows through the SoG and their uncertainty based on observations require long-term monitoring stations such as the one currently running at the westernmost section of ES. The quality and accuracy of the time series employed to this aim would be improved by deploying, at least, a second twin monitoring station at a different site of the section that allows for addressing explicitly the cross-strait structure of the velocity field and, hence, for performing more robust computations of the flow. On the other hand, the peculiar hydrodynamics of the region strongly suggests the selection of specific locations for deploying the monitoring stations. Considering that eddy-fluxes are hard to estimate from observations, the best places for monitoring the inflow and outflow are the eastern and westernmost boundaries of the Strait, respectively, because the influence of the eddy-fluxes is minimized in these sections. While the outflow is rather satisfactorily monitored by the single station

[2] The ascribed uncertainty should reflect the medium-to-long term variability of the transports, that is, the variability after removing short–term fluctuations, which are the tidal and meteorologically driven fluctuations. The length of the hindcasts is not enough to assess the uncertainty of the long-term scale, but it does allow for the satisfactory removal of tidal variability. Sammartino et al. (2015) shows that the medium-to-long term variability of the outflow is about half the variability of the subinertial series obtained after filtering out the tidal variability. This has been our criterion to assign the uncertainties to all variables in Table 4. Should we have taken the standard deviation of the original series, the uncertainty in, say, the heat transport through ES during hindcast I would have been 120.7 TW instead of 5.43 TW showed in Table 4. The first number is not a realistic long-term uncertainty at all.

currently deployed at ES, the inflow is not. Notice that, nowadays, the inflow estimations are almost uniquely based on an indirect approach using the hydrological budget of the Mediterranean, which is not well constrained, and the direct observations of the outflow (see for instance Soto Navarro et al., 2010). Monitoring the inflow requires a good sampling of the top tens of meters of the water column through which a considerable fraction of the inflow occurs. Monitoring stations based on moored up-looking ADCPs miss most of this layer because of the acoustic reflection at the sea surface. Stations based on down-looking ADCPs located at the surface are unfeasible due to the heavy ship traffic. The development of High Frequency Radar stations (HFR) that measure the surface currents open new possibilities for observing the inflow, although the accuracy of HFR observations need to be further improved. A HFR network installed by Puertos del Estado (Spain) in the eastern part of the SoG (http://www.puertos.es/en-us/oceanografia/Pages/portus.aspx) is presently running and providing surface velocity fields that can be suitably used, in combination with moored up-looking ADCPs sampling the interior, to obtain long time series of the inflow. Recently Jordà et al. (2016) have demonstrated that a reduced order optimal interpolation scheme is a feasible approach to combine both sources of information and to get accurate estimates of the inflow.

While the new monitoring systems have obvious potential to improve the volume transport estimates, the continuous sampling of temperature and salinity necessary to estimate heat and salt transports at the same sections where velocity is observed poses challenges that are not overcome yet. The approximately four-month interval sampling of the ES and easternmost sections using a CTD probe equipped with a Lowered ADCP (LADCP) to measure velocity and the properties of the water column simultaneously, which is accomplished routinely by the Spanish Institute of Oceanography since the beginning of this decade, is a significant step forward, but insufficient due to the strong contamination introduced by tides. Supra-tidal sampling interval is required to filter out this noise. One possibility would be to install ADCP in the commercial vessels that cross the Strait routinely. This would provide a large amount of information about the structure of the velocity field and, after a proper filtering out of the tidal signal, an estimate of the water transports. Concerning the hydrographic properties, the required information could be obtained by deploying gliders programmed to sail across the Strait. Two important problems must be envisaged: the heavy ship traffic and the noticeable currents that will probably drift the glider away from the programmed track. An alternative solution that could partially overcome those problems would be to deploy a moored profiler that could take vertical profiles of temperature and salinity autonomously. This solution, however is not exempt of technical difficulties either.

4- Exhanges trough the Dardanelles Strait

4.1 Overview

The Turkish Straits System (TSS), located at the junction of the Asian and European continents, provides a restricted connection between the large inland basins of the Mediterranean and Black Seas (Figure 9). The Sea of Marmara (surface area 11,500 km^2) connects the Aegean and Black Seas through the Dardanelles (western limit, length 75 km, min. width 1.3 km) and Bosphorus (eastern limit, length 35 km, min. width 0.7 km) Straits. Three elongated depressions

(max. depth ~1350 m) interconnected by sills (depth ~600 m) and adjoining continental shelves of 40 and 10 km wide, respectively on the southern and northern sides, are found within the Sea of Marmara. The entire TSS is two-layer stratified (Andersen and Carmack, 1984; Ünlüata et al., 1990), with a sharp pycnocline at 25 m separating water masses of Black Sea and Mediterranean Sea origin. The wind-mixed upper layer transits through the system as it responds to seasonal changes, while the lower layer waters transported from the Aegean Sea (Beşiktepe et al., 1993, 1994, 2000), supply limited oxygen to the deep basins (Beşiktepe et al., 1993, 1994).

Besides the density (hence baroclinic pressure) differences between the two basins, atmospheric pressure and wind effects in the adjacent basins create additional pressure gradients that maintain the exchange through the Bosphorus (Özsoy et al., 1998). Turbulent entrainment and interfacial mixing modify waters in transit through the two straits (Beşiktepe, et al., 1993, 1994; Gregg and Özsoy, 1999, 2002; Gregg et al., 1999; Özsoy et al., 2001), with the highest rates of change of upper layer salinity occurring in the southern Bosphorus and also in the western part of the Dardanelles.

Current-meters and both ship-mounted and bottom mounted ADCP measurements in the Bosphorus (Pektaş, 1953; De Filippi et al., 1986; Gregg and Özsoy, 1999; Çetin, 1999; IMS-METU, 1999; Özsoy et al., 1998, 2009; Gregg and Özsoy, 2002; Jarosz et al., 2011a,b, , Özsoy and Altıok, 2016; Tutsak, 2013; Tutsak et al., 2016; Özsoy et al., 2016c) and Dardanelles (Jarosz et al., 2012, 2013), as well as bottom pressure measurements at their junctions (Book et al., 2014) have revealed many different time-scales of oscillations in the TSS, ranging from inertial, semi-diurnal, diurnal to several days periods influenced by the adjacent basins (Yüce, 1993; IMS-METU, 1999; Tutsak, 2013; Tutsak et al., 2016; Özsoy et al., 2016c). Tidal oscillations are exceptionally small, on the order of ~10 cm in the TSS, especially east of the Nara Pass of Dardanelles. Basin oscillations with periods of 2-5h have been observed in sea level records (Alpar and Yüce, 1998). Coupled Helmholtz mode oscillations of the Black Sea and the TSS (e.g. Ducet et al., 1999) with 14.7d and 1.9d periods and a two-layer exchange adjustment time scale of 42d have been estimated (Özsoy et al., 1998). The spectral analyses of current velocity measurements provided by Jarosz et al. (2013) appear to support oscillations at those low frequencies.

The narrow geometry and topographical complexity of the straits support fast currents leading to nonlinear processes of hydraulic controls and turbulence. With hydraulic controls operating at a sill outside its Black Sea exit and a contraction and a sill in the southern part, the Bosphorus is a good example of the maximal exchange regime (Farmer and Armi, 1986), operating in the full range of weak to strong transient barotropic component. The Dardanelles Strait flow appears to be in the submaximal regime, subject to a single hydraulic control at the mid-strait contraction (Ünlüata et al., 1990).

4.2. Observations

CTD station data collected by the R/V BILIM of the IMS-METU has the highest rate of sampling during the 1985-2001 when intensive studies were carried out at seasonal to annual repetition periods in the TSS (Ünlüata et al., 1990; Latif et al., 1991; Özsoy et al., 1998) and has been continued sporadically in many other scientific cruises since then. A most detailed joint experiment in the Bosphorus and its Black Sea exit, covering also some turbulence and fine

structure measurements was carried out in 1994 (Gregg et al., 1999; Gregg and Özsoy, 1999, 2002, Özsoy et al., 2001). Detailed mapping of the Bosphorus currents and hydrography was undertaken in 1998-1999, when ADCP and CTD measurements were extended to cover small bays and bends (IMS-METU, 1999). Recent measurements have been acquired from automated coastal meteorology, sea level and ADCP stations (http://moma.ims.metu.edu.tr) operated under a coastal network (Özsoy et al., 2009, Tutsak, 2013, Tutsak et al., 2016). During this experiment, a unique set of continuous current profiles were obtained at 15 min. intervals, continued for several years (2008-2012) from a cabled bottom mounted ADCP system deployed near the narrow southern section of the Bosphorus. The sea-level stations established in the project have been operational since 2008. Unfortunately the ADCP measurements have been concentrated mostly at the Bosphorus area and measurements other than CTD are relatively rare and results based on existing data have not been sufficiently reported in the Dardanelles Strait, except for the analyses reported in IMS-METU (1999). A bottom deployed ADCP was placed in the Dardanelles Strait in 2008, though the measurements were terminated after the cable had been pulled out and damaged by fishermen. Further detailed measurements were obtained during the recent SESAME European project (2007-2008) and the following TÜBİTAK research projects (Özsoy et al., 2015). Additionally, a remarkable experiment surveying the full Turkish Straits System with multiplatform observations has been carried out during 2008-2009, by the CMRE (Centre for Maritime Research and Experimentation, formerly NATO Undersea Research Centre, NURC) in collaboration with the Turkish Office of Navigation, Hydrography and Oceanography (SHOD) and the US Naval Research Laboratory (Jarosz et al, 2011a,b, 2012, 2013; Chiggiato et al., 2012; Book et al., 2014; Gerin et al., 2013). As part of this experiment, a one-year time series of bottom mounted ADCP in the straits allowed the computation of water fluxes. These measurements (Jarosz et al., 2013) confirm the order of magnitude of the earlier measurements and provide detailed characterization of the multiple scales of motion in the TSS. A critical review of the past time-series measurements through the TSS can be found in Özsoy and Altıok (2016a,b).

4.3 Modelling results

Although a summary of the modeling is beyond the present scope, it is worth mentioning the grand challenge of predicting the TSS exchange currents and material fluxes that is being taken in a number of steps, using models of increasing complexity. In terms of the complexity of its dynamics and the observational and modeling challenges posed, the TSS is very similar to the Strait of Gibraltar, and on many accounts even more challenging because of the hydrodynamic coupling between three adjacent sea basins and two narrow straits. Advanced three-dimensional modeling studies of the Bosphorus Strait (Sözer, 2013, Sözer and Özsoy, 2017) based on ROMS, as well as curvilinear grid MITgcm (Sannino et al., 2017) and unstructured grid FEOM (Gürses et al., 2016) models of the entire TSS have recently been used for solving the circulation of the entire TSS through the use of high performance computation, thereby fully resolving the current dynamics of the narrow Bosphorus and Dardanelles Straits. Modelling results obtained so far on the structure of the currents and on the exchanged water fluxes through the TSS confirm the ranges obtained in experimental measurements, and verify the limits when either the upper or the lower layer gets blocked. It is also shown that the Marmara Sea circulation and the exchange with the adjacent basins are sensitively determined by the nonlinear, strongly stratified, turbulent dynamics of the exchange flows at the straits. These

results show the sensitivity of the response to the net flow through the TSS, as well as other immediate and remote forcing factors, confirming the need for integrated modelling of the entire TSS, with the Marmara Sea, the Dardanelles and Bosphorus Straits coupled together, in order to account for, and eventually be able to fully understand the dynamical interactions and exchange between the adjacent basins of the Mediterranean and Black Seas.

4.4 Water fluxes

In the Black Sea, the total freshwater input (P=300 km^3/yr and R=350 km^3/yr) is twice as large as the loss term (E=350 km^3/yr) which means that the annual net water budget (Q=P+R-E) is positive and results in a net flux towards the Mediterranean. Making use of an assumption of steady state mass balance, and multi-year averages of salinity measurements at the junctions of the straits, Ünlüata et al. (1990) have computed the annual average fluxes through the TSS from a two-layer box model, updated later with a seasonal version provided by Tuğrul et al. (2002) and Beşiktepe (2003). An updated description is provided here based on the review given in Schroeder et al. (2012).

The annual two-layer volume fluxes of the TSS are shown in Figure 10a. The exchange through the TSS increases in the spring-early summer, and weakens markedly in autumn (within a margin of about ~40% of the annual mean) in response to the freshwater input to the Black Sea (Tuğrul et al., 2002). The input into the Marmara Sea by the upper layer flow of the Bosphorus is estimated as 657 km^3/yr and the outflow into the Black Sea in the lower layer is estimated as 318 km^3/yr (Figure 10a) based on the measurements during the years 1985-1995 (Tuğrul et al. 2002). The input into the Aegean Sea by the upper layer flow of the Dardanelles was found to be 1331 km^3/yr and the outflow into the Marmara Sea in the lower layer was 1010 km^3/yr in the same period (Tuğrul et al. 2002). The net flow at both sections is then 321 km^3/yr which, if divided by the area of the Mediterranean Sea (2.5×10^{12} m^2) results in an equivalent contribution to the Mediterranean budget of 128 mm/yr.

Bosphorus fluxes computed from ADCP data (Özsoy et al., 1998) show the same seasonal behaviour as reviewed above, but reveal maxima of about Q_1 = 1600 km^3/yr and Q_2 = 630 km^3/yr for the upper and lower layers respectively, including blocked cases, indicating instantaneous fluxes 2-3 times larger than the annual mean. Despite large scatter in data due to sampling, overall average values of Q_1 = 537 km^3/yr and Q_2 = 112 km^3/yr were computed, thus leading to a net average value of 425 km^3/yr (170 mm/yr of equivalent sea level change in the Mediterranean Sea). It has to be noted that the lower layer flux Q_2 is possibly underestimated as a result of data loss near the bottom, which implies net transport to be overestimated.

Based on two layer (1D) modeling results, Oğuz et al. (1990) concluded that the upper or lower layer flows would be blocked when the net flux exceeds -580 km^3/yr or 800 km^3/yr, in respective directions. The different sources of data and model results in Figure 11 reveal -630 km^3/yr and 950 km^3/yr respectively as the approximated limiting values beyond which either the lower or the upper layer are expected to be blocked. Figure 11 compares various sets of observations (Merz and Möller, 1928; Özsoy et al., 1998) and results from two different models (Sözer, 2013; Sözer and Özsoy, 2017; Sannino et al., 2017), establishing the relative upper and lower layer exchange fluxes as a function of the net flux imposed on the Bosphorus Strait. Although the

fluxes reported in Figure 11 are for the Bosphorus Strait rather than the Dardanelles, they indicate the range of fluxes expected and the net flux Q_1-Q_2 is of course the same between the two Straits assuming that volume in the Marmara Sea does not change significantly at long time scales.

Indirect estimates of the net transport through the Bosphorus (i.e. equivalently through the Dardanelles) based on long-term water fluxes and sea-level variations in the Black Sea (Stanev and Peneva, 2002) indicate seasonal anomalies (Figure 10b), with variations of the order of ~75% of the mean, that is considerably larger than the ~40 % estimated by Tuğrul et al. (2002) from the seasonal based mass budget. These differences arise from data uncertainties as well as differences in the averaging (monthly versus seasonal) applied to the data. However, both methods yield the same general pattern of seasonal variability.

Jarosz et al. (2012, 2013), based on ADCP data gathered during 12 months in 2008-2009 in both ends of the Dardanelles, have found that the input into the Aegean Sea by the upper layer flow of the Dardanelles was 1157 km^3/yr and the outflow into the Marmara Sea in the lower layer was 999 km^3/yr, the net flux therefore amounting to 158 km^3/yr (63 mm/yr of equivalent Mediterranean sea level change). Figure 12 shows the fluxes collected in 2008/2009 at the southern section of the Dardanelles Strait and it is worth remarking the large variability observed in the flows. The volume fluxes exchanged between the Aegean and Marmara Seas via Dardanelles Strait are subject to a wide range of oscillations that are especially amplified in winter months and often two times larger than the mean fluxes (Jarosz et al. 2013). It is also observed that the net barotropic fluxes are even of greater amplitude and oscillate about a much smaller mean value. The minimum and maximum values of the upper layer flux respectively were -2593 and 149 km^3/yr during the measurement period. For the lower layer the extremes were -434 and 3375 km^3/yr, and for the net fluxes one finds -2368 and 3513 km^3/yr for the respective minimum and maximum values.

Concerning the uncertainties in the estimates of the net water fluxes (reviewed by Özsoy and Altıok, 2016), up to our knowledge there are no accurate estimates of them yet. The retrospective evaluations of long-term average fluxes should take into account the instrumental accuracy, the uncertainties in the estimates of the fluxes from limited observations and the influence of inter-annual variations. Therefore, up to now only the comparison among different published estimates (see Table 6) can be used as a rough approximation of the uncertainty. By doing this, the net water flux is 275 + 59 km^3/yr towards the Mediterranean (or 110 + 24 mm/yr in equivalent sea level change in the Mediterranean Sea).

4.5 Heat and salt fluxes

To our knowledge there are no published estimates of the heat and salt fluxes through the Dardanelles Strait. In order to get a first estimate of those fluxes we do a simple computation using published values for the temperature and salinity of the upper and lower layers. In particular, we estimate the net heat flux as $(Q_1 \cdot T_1 + Q_2 \cdot T_2)$ and analogously the net salt flux as $(Q_1 \cdot S_1 + Q_2 \cdot S_2)$, where subindex 1 and 2 represent the upper and lower layer, respectively. For the temperature and salinity we use T_1=14ºC, T_2=16ºC, S_1=35 psu and S_2=39 psu, similar to the values observed in the Dardanelles (e.g. Jarosz et al. 2013). Then, in order to take into account the time variability in the water fluxes and the uncertainty in those estimates we assume a random error

of 200 km^3/yr for Q_1 and Q_2, which is based on the std of observed fluxes by Jarosz et al. (2013) and the spread among estimates shown in Table 6. Similarly we consider an uncertainty of 2ºC for the temperature and 2 psu for the salinity, although sensitivity tests have shown that uncertainties in the temperature and salinity have a second order importance in front of the uncertainties in the water fluxes. The estimated inflow, outflow and net heat fluxes are 0.90 $\pm$ 0.15 W/m^2, -0.80 $\pm$ 0.17 W/m^2 and 0.10 $\pm$ 0.16 W/m^2, respectively (here positive indicates towards the Mediterranean). The estimated inflow, outflow and net salt fluxes are 1.35 $\pm$ 0.20 10^6 kg/s, -1.18 $\pm$ 0.22 10^6 kg/s and 0.20 $\pm$ 0.22 10^6 kg/s, respectively.

4.6 Perspectives

At present there are no monitoring systems that can provide continuous estimates of the fluxes through the Dardanelles Strait. However, previous studies have shown that this is technically feasible. A monitoring system similar to the observational system used by Jarosz et al. (2013) could allow an accurate estimate of the water fluxes in the Dardanelles Strait. That system should be complemented with routine measurements of temperature and salinity profiles at least on the limits of the Strait in order to compute heat and salt fluxes. However, up to our knowledge there are no plans to implement and maintain such system. A complement to the observational approach would be the operational implementation of numerical models fully resolving the dynamical and mass balances of the adjacent basins and straits (i.e. similar to those presented above). Those models could be calibrated and/or validated using a more limited set of observations (i.e. from short term monitoring periods) and then provide continuous estimates of the heat and mass fluxes into the Mediterranean.

5- Riverine input

5.1 - Overview

The rivers carry to the sea the residual of precipitation and evaporation over the entire continental catchment of the Mediterranean Sea. This area covers about 7.9 10^6 km², or three times the area of the sea itself and goes from tropical Africa up to Russia. This area is drained by some of the largest rivers of the world, like the Nile (2.9 10^6 km²) or the Danube (0.8 10^6 km^2). Large parts of this drainage basin are located in arid areas (northern Africa), thus leading to relatively low net rainfall (precipitation minus evaporation) and river discharge. The Mediterranean catchment is also strongly populated and men since millennia have regulated the flow of water in order to make the best usage of it. Today, most of the Mediterranean countries practice an irrigated agriculture which essentially takes water from rivers in order to enhance evaporation and crop productivity. In the countries on the southern shores more than 80% of the net rainfall is used for irrigation (Margat 2008). It is thus a process which cannot be ignored when quantifying the riverine input.

5.2 - The reliability of observational based estimates

The largest rivers are being gauged by national agencies but such data is not always publicly available or reliable. The societal importance of the water flux available in the rivers also makes this information sensitive from an economical or geopolitical point of view. Therefore the data is not always made available or only after having been treated to remove the impact of some

agronomic or industrial activities. Thus many of the data publicly available are either outdated or with incomplete information on possible treatments applied to remove the impact of water usage. Obviously taking into account the contribution of un-gauged rivers is even more difficult. Especially in semi-arid regions where overland flows are the prime source of water for irrigation. Concerning the data realiability, on the one hand a gauging station has to be as close as possible to the mouth of the river in order to be as much representative as possible of the entire catchment, but on the other hand it has to be unaffected by estuarine processes and thus free of saline intrusions or tidal effects. A further complication to the observational based estimates is that many large rivers in the Mediterranean region form a delta. Observing the river discharge upstream the delta neglects the strong evaporation on such fertile grounds, which are often the heartland of agricultural activities and thus of human water usage.

The World Meteorological Organization (WMO) sponsored Global Runoff Data Centre (GRDC) has assembled river discharge data from most basin agencies around the world and thus offers today the most complete data set. But this data comes without indication on the quality of the observations. It is unknown to which precision the basin agency has monitored the transect of the river or which assumptions were used in the velocity measurements. Nevertheless, GRDC data can be the basis for an observational derived freshwater input into the Mediterranean. This effort has been undertaken at CEFREM (Centre de Formation et de Recherche sur les Environnements Méditerranéens) by identifying for each coastal point on a regular latitude longitude grid the available data and computing an estimate of the annual mean influx (Ludwig et al. 2009). As the last station on the river before its mouth can be before the confluence of other tributaries, some corrections are applied to add the flux which has not been observed. In the CEFREM data set only about 63% of the total flux is based on observations. For coastal points where un-gauged rivers end, the authors have applied a simple annual water balance model which assumes that the total annual rainfall over the basin minus an estimate for average evaporation yields the flux into the ocean. The limitation to annual mean fluxes is justified by the fact that human regulation of rivers mostly affects the seasonal cycle of discharge. The CEFREM is available in 2 versions which differ by the resolution at which they have been established and the period they have covered. In both versions the same methodology is used but different atmospheric data are considered for the water balance model.

The riverine input estimate by CEFREM is currently the best observational based product available. It estimates that the total freshwater influx is 780 km^3/yr for the Mediterranean and Black seas over the period 1960-1969 and 710 km^3/yr for the period 1991-2000 (Both averages are computed with version 1 of the data, see Figure 13). A reduction is also found when the total discharge is separated in different subbasins. For the period 1960-1969 the discharge in the western Mediterranean, the eastern Mediterranean and the Black Sea are 137, 253 and 390 km^3/yr, respectively. For the period 1991-2000 the values are 115, 213 and 382 km^3/yr, respectively. This reduction is attributed by the authors to changes in the climate forcing using a simple precipitation/runoff model. The CEFREM products inherit the uncertainties inherent to the observed discharge discussed above and which cannot be quantified because of the lack of meta-information on observational methods or the post-processing applied to remove the

human influence. It is also limited by the simple runoff model used for un-gauged basins and the fact that only annual mean values are proposed.

5.3 - The case of the Nile

To illustrate the complexity of the human management of the rivers it is interesting to examine in more detail the case of the Nile. This river has the third largest drainage area in the world but has probably always contributed less fresh water to the Mediterranean than the Danube. Over the last centuries a number of dams have been built to regulate its flow, produce hydro-power and develop irrigation based agriculture. In its lower basin the first major dam to be built was the lower Aswan dam which was filled in 1902. The Aswan gauging station reports a mean annual discharge of 112 km^3/yr over the 1871-1900 period (Vörösmarty et al. 1998). The same station reports for the period 1903-1960 an annual mean value of 88km^3/yr indicating a reduction of the water carried by the river. In 1976 the high Aswan dam reached its capacity and the downstream station reported for the final period over which data is available (1976-1984) an average discharge of 59km^3/yr.

But the abstractions of water are not limited to the dams built upstream of the city of Aswan. Vörösmarty et al (1998) and GRDC report data for the 1976-1984 period at a station just before the delta (El Ekhsase) and which displays and average flow of 40km^3/yr. Obviously a significant fraction of the water observed in El Ekhsase will evaporate within the delta or reach the sea via canals, drains or groundwater. Nixon (2003) estimates that in 1984, 14 km^3/yr reached the sea.

Since the high Aswan dam was filled, agriculture and other water usages have increased in Egypt. Based on the reported discharge at the Aswan station, the infrastructure could have made available for Egypt's development 29km^3/yr (difference between the 1903-1960 and 1976-1984 periods) which are carried north through a network of channels. If this water had been evaporated over the area of Lake Nasser (5000 km^2) and the irrigated surfaces in Egypt (3 10^4 km^2 FAO 2005) this would have meant an increase of 2.27 mm/d of the flux to the atmosphere. Considering that maximal evaporation of crops in similar climates are rarely above 6mm/d during the growing season (Duchemin et al. 2008, Garatuza-Payan et al. 1998), this seems to be a unrealistic increase of annual mean evaporation following the construction of the Aswan high dam and hints to the fact that a large part of the water lost on the way to the sea could have infiltrated into the soils. This water would then flow to the Mediterranean after passing through the aquifers.

This discussion of the Nile river suggests that probably a significant fraction of the water entering Egypt still reaches the sea but it is not accounted for by the gauging station as it flow to the ocean through canals, drains and aquifers. Smaller arid basins of the Mediterranean will probably pose similar observational problems but at smaller scale.

We propose here to quantify the annual mean riverine input into the Mediterranean basin based on the CEFREM data as it is based, for a large part, on observed discharges. In particular we use the average of the V1 and V2 products which differs in the period spanned, the spatial resolution and the atmospheric forcing used to infer the discharge of un-gauged rivers. The observed discharge in monitored basins is the same in both version of the product. The average discharge

for the period 1960-2010 we obtain is 767 km³/yr, which can be split in 125, 250 and 392 km³/yr for the western Mediterranean, eastern Mediterranean and Black Sea. As the direct uncertainty cannot be estimated it is suggested to use the uncertainty from the ORCHIDEE model as a surrogate (see next section). It has to be reminded that a significant fraction of the freshwater inflow in the CEFREM data is based on a simple water balance model because of a lack of hydrological data. Based on this hypothesis we obtain an uncertainty of 107km³/yr for the entire basin. It is important to note that this assumed error of 14% hides a very contrasted picture. In the basins where the hydro-meteorological data are reliable, ORCHIDEE simulates correctly the major rivers and the error can be as small as 10%. In contrast for Mediterranean sub-basins affected by irrigation and poor observational networks, the errors can be in excess of 100%.

5.4 - Modelling riverine inputs

Riverine input can also be obtained through global hydrological models. Their objective is to simulate the continental water cycle under given atmospheric conditions by representing in as much detail as possible the transfer of water and its usage. These models are often validated and calibrated on observed discharge and thus can be considered as the most reliable predictions of river flows. One such model is WaterGAP3 (Döll et al. 2003) which has been used to study in detail the water cycle of the Mediterranean basin (aus der Beek et al. 2012). It is able to predict, at a daily time step, realistic discharge values for most rivers and provide an estimate of the water withdrawals for irrigation. WaterGAP3 predicts a riverine input into the Mediterranean (including the Black sea) of 924 km³/yr for the period 2002-2009.

Alternatively, forcing a state of the art land surface model (LSM) with bias corrected atmospheric conditions allows also to produce estimates of the freshwater inflow into the Mediterranean. The LSMs have been developed to represent continental processes in climate models and include complex parameterizations for plant and bare soil evaporation as well as for hydrological processes and simulate the water and energy exchange at a sub-diurnal time step. In some cases they have even introduced representations for irrigation and floodplains in order to estimate water taken from the streams to sustain or augment the evaporation of crops and wetlands. One such model is ORCHIDEE developed at the Institut Pierre Simon Laplace (IPSL). It has been validated over a number of catchments of the world, and most recently over the Amazon basin (Guimberteau et al. 2012). Using the WATCH atmospheric forcing data bases (WFD and WFDEI) (Weedon et al. 2011, Weedon et al. 2014), the total fresh water flowing into the Mediterranean can be estimated with ORCHIDEE. Such an exercise is presented in Figure 13 where the simulations with different atmospheric forcings are compared to the estimates of CEFREM. It has to be noted that the model reproduces well the observed inter-annual variability of the riverine input and the uncertainty in the forcing introduces errors which are similar to the difference between the two versions of the CEFREM product.

These integrated numbers over the entire basin hide a number of large discrepancies. This can be well illustrated again with the case of the Nile. WaterGAP3 and ORCHIDEE produce discharges of over 300 km³/yr which is well above the values observed at Aswan before the dams or the modern irrigation infrastructure of Egypt were installed. This can be explained by the limited capacity of the models to represent the floodplains of the upper Nile, the lack of regulation by the Nalubaale Dam at the outflow of Lac Victoria and not the least the poor atmospheric forcing

over the tropical part of the basin. More generally, ORCHIDEE overestimates the inflow into the ocean of arid basins and thus underestimates the continental evaporation there. Certainly the irrigation parameterization of ORCHIDEE is too simple (de Rosnay et al. 2003) but one may also wonder if the observations represent correctly the water cycle in these basins which are not often not well monitored.

5.5 - Heat flux contribution of rivers

To our knowledge there is no database of temperatures for streams flowing into the Mediterranean. This is generally a topic which is lacking attention for the research community, probably because of the complexity of the processes involved. We are not aware either of efforts under way to include this parameter in land surface models or global hydrological models. Some isolated effort have been undertaken, as for instance by Beaufort et al. (2015) for the Loire, but a generalization to all water flows on continents is desperately lacking.

A first approximation for the total contribution of rivers to the heat budget can be derived assuming that the temperature of the river waters is similar to the temperature in the sea. For each river considered by ORCHIDEE we obtain the closest sea surface temperature from the SST product of Marullo et al. (2007). Then, we estimate the heat flux for each river and do the integration for the whole basin. In order to take into account the different sources of uncertainty in this estimate we first consider that the temperature in the river can be $\pm$ 5ºC the sea temperature. Then, we also assume that the uncertainty in the river flow is a 15% of the total value (section 5.3). The result of this estimate show a marked seasonal cycle (see Figure 14) with a maximum in November (1.1 $\pm$ 0.2 W/m^2) and a broad minimum from May to July (~0.3 $\pm$ 0.1 W/m^2). This annual minimum is relatively constant along time, while the annual maximum shows significant variability ranging from 1 to 2 W/m^2. The long term annual average of the heat contribution from the rivers is 0.6 $\pm$ 0.4 W/m^2.

5.6 - Perspectives

In view of the difficulty of accessing discharge data close to mouth of the rivers flowing into the Mediterranean and knowing how much of the total flux they actually capture, the observationally based approach will always be affected by large uncertainties. Models are plagued by large errors today, but they offer a good potential for improvements. The quality of their atmospheric forcing will certainly improve. Developments will enhance their ability to represent irrigated areas and floodplains as well as groundwater processes. With irrigation playing such an important role in arid basins, reducing the uncertainty in the manmade enhancement of evaporation will be key in refining our estimate of the riverine inputs. Furthermore satellite information can be used to validate lake levels, irrigated areas or the total continental water storage. Also, observed discharges throughout the basin can be used to verify the models or correct the predictions with assimilation methods. Finally, the temperature of the river waters should be measured, at least in the largest rivers, in order to estimate the heat flux associated to the river runoff. Once that data will be available it could be compared to sea surface data from satellite observations to identify up to which extent the sea temperature can be used as a proxy for the river temperature.

6 - The Mediterranean Sea heat and mass content

6.1 - Overview

Direct estimates of the Mediterranean heat and salt contents can only be obtained from in-situ measurements of temperature and salinity, respectively. Those measurements represent a discrete set of observations with typically large gaps in the spatial and/or the temporal coverage. Thus, the in-situ observations have to be averaged using more or less sophisticated approaches in order to produce basin or sub-basin integrated quantities. Therefore, the accuracy of the heat and salt content estimates will depend not only on the accuracy of the measurements but especially on their spatial and temporal distribution as well as on the properties and the tuning of the algorithm used to integrate those measurements. For instance, Vargas-Yáñez et al, (2009) have shown that the method used to define anomalies (e.g. removing a monthly or a yearly climatology) can lead to large differences in the estimates of the western Mediterranean temperature and salinity trends (i.e. more than 50% depending on the method used). Also, Vargas-Yáñez et al. (2012) have shown that computing trends from an interpolated product instead of using the raw profiles can also lead to large discrepancies (e.g. from non significant trends for the upper layer temperature in the western Mediterranean to a significant trend of 0.02ºC/yr, respectively).

The water mass content has to be estimated indirectly since it cannot be directly measured. In order to better understand the links between the changes in the volume and mass of the basin we present first some basic concepts. For a water column of area A, a change in the sea surface elevation (η) can be due to changes in the density of the water column (ρ; steric component) and/or to changes in the mass of the water column (*m*):

— — — —— — (7)

Assuming a simple linear approach for the (temporal and spatial) changes of density with temperature and salinity —— we get:

— — — —— —— —— —— (8)

where α and β are the thermal expansion and haline contraction coefficients, respectively. Equation (8) expresses that changes in the volume of the water column can be associated to changes in the thermal content (thermosteric), the salt content (halosteric), the mass content of salt and the mass content of water. It is important to highlight that any change in the salinity of the water column implies a change in the salt content. For more details on these concepts and the implications for the Mediterranean basin the reader is referred to Jordà and Gomis (2013).

Concerning the quantification of the changes in the water mass content, on one hand gravimetric observations (as those provided by the Gravity Recovery and Climate Experiment or GRACE mission) provide a measure of the changes in the total mass. However, it has to be noted that this includes both changes in the water content and in the salt content (last two terms in (8)). I.e.: GRACE cannot distinguish if the mass change is due to a salinification of the basin or to an increase in the amount of water. Therefore, a complementary estimate of the changes in the salt content (e.g. from the in-situ observations mentioned above) has to be provided in order to deduce the changes in the water mass content. On the other hand, observations of sea level (i.e. from altimetry or tide gauges), can be used to measure changes in the total volume of water in the basin (the left hand side term in (8)). In this case, though, it has to be noted that changes in the volume may be due to an expansion/contraction of the water column without necessarily implying a change in the water mass content. Therefore, the thermosteric component, the halosteric component and the changes in the salt content of the basin have to be estimated with complementary datasets.

In the following we summarize the different types of temperature and salinity observations that are currently available for the Mediterranean, the choices that are typically used to integrate those observations to compute the heat and salt content, the different methods to estimate the water content changes and, finally, we discuss the contribution of numerical modelling.

6.2 – In situ observations of temperature and salinity in the water column

Since the first decades of the XXth century different kinds of instruments were available to measure temperature and salinity of the water column. The first measurements were obtained with Nansen bottles which were equipped with reversing thermometers. This system provided observations of temperature and salinity at a limited number of vertical levels and a non-negligible uncertainty existed in the estimate of the depth at which the measurements were obtained. Also, the Mechanical BathyThermograph (MBT) was widely used until the end of the 70's. The MBT provided profiles of temperature vs depth for the upper part of the water column. Its accuracy was low (~0.05°C) and recently a bias in the MBT data has been identified which can induce an error of up to 0.2-0.4°C (Gouretski and Koltermann, 2007). MBT was substituted by the expandable BathyTermograph (XBT) which is still used in different applications. The XBT provides profiles of temperature *vs.* depth down to large depths but its accuracy is low (~0.05-0.1°C) and also suffer from biases in the determination of the depth, which translate into errors of up to 0.2-0.4°C. In this case, however, there is less consensus about how to correct those biases on the historical data (Cowley et al., 2013). Most MBT/XBT data can be found in the databases although usually uncorrected. However, it is worth noting that Vargas-Yáñez et al. (2010) have shown that including or not MBT/XBT data in heat content estimates of the western Mediterranean does not significantly change the results.

Also since the early '70s, oceanographic vessels used to collect temperature and salinity data from surface waters underway (thermosalinometers). This procedure has been extended to "ships of opportunity" (SOOP) such as ferries or liners that cover a repeated route across the Mediterranean (e.g. MedGOOS).

At the end of the 70s, when CTD (Conductivity, Temperature and Depth) instruments reached a reasonable accuracy, they started to be widely used everywhere, including the Mediterranean. CTDs provide a continuous vertical profile of temperature and conductivity (which is converted into salinity) at a given station, from the surface to the bottom. Up to now, it is agreed that CTDs provide the most accurate measurements of temperature and salinity (accuracy of 0.001°C and 0.001 psu, respectively). Generally both salinity and conductivity sensors are calibrated on a regular basis and the measurements are checked on board with reversing thermometers and with water samples that are analyzed directly with a salinometer. Ship-based hydrography represents the only method for obtaining high vertical resolution measurements over the whole water column. Presently the Mediterranean Argo network (described below) samples only the upper 2000 m of the sea. If one considers that 20% of the Mediterranean volume lies below 2000 m, where significant changes have been recently observed, it appears clear that ship-based full-depth CTD data become essential to reduce uncertainties in freshwater, heat and sea level budgets and to assess the long-term steric contribution to Mediterranean sea level (Schroeder et al., 2015). In addition, ship-based hydrographic measurements provide a standard for validating new autonomous sensors and a reference data set for other observing systems such as Argo profiling floats, expendable bathythermographs, and gliders. However, even if there is a large amount of CTD data available in the international databases, there is still a non-negligible fraction of the total dataset acquired that is only available upon request submitted to the responsible institutions.

Since the turn of the century, profiling floats have been deployed in the Mediterranean as part of the international Argo program (Poulain et al., 2007) to measure profiles of temperature and salinity at intervals of 5 or 10 days, between the surface and maximal depths of 2000 dbar. Nearly 30,000 profiles have been collected throughout the Mediterranean basin between 2001 and 2015. The number of floats operating simultaneously has reached a maximum of about 80 in mid-2015 and future deployments aim at maintaining an active Argo fleet of 60-70 floats in the Mediterranean Sea. The Argo data are accurate (±0.002ºC, ±0.002 psu, ±2 dbar for temperature, salinity and pressure respectively, for SBE 41 CP model) and although the pumped measurements are not as reliable in the upper 5 m it provides a reasonably well distributed coverage reaching regions which are traditionally not surveyed by oceanographic campaigns. Also, the Argo data are centralized and subject to homogeneous quality checks. However, it is worth noting that the calibration of the instrument remains a very sensitive issue. More information on the Mediterranean Argo float data can be obtained at the Coriolis data center and the Mediterranean Regional Argo Center (http://nettuno.ogs.trieste.it/sire/medargo/).

During the first decade of the XXI century several autonomous underwater vehicles (gliders) have started to be operated in the Mediterranean. These vehicles allow to obtain a very spatially dense set of quasi-vertical profiles with reasonably good accuracy (0.004ºC and 0.01 psu for temperature and conductivity respectively, for the GPCTD model). Also for these platforms, calibration is an important issue, along with problems related to the fact that not all CTDs mounted on gliders are pumped (which leads to a systematic salinity offset between descending and ascending profiles). The glider data are quite unified in the databases and efforts are being devoted to homogenize the quality control applied to the data. Additionally it has to be

mentioned that the very large contribution of the gliders during the last years can be misleading (see Figure 15): glider data can be redundant because profiles can be very close to each other, hundreds of meters apart. Concerning the salinity measured by the gliders it has to be mentioned that both are affected by the thermal lag problem (Garau et al., 2011), which causes that the conductivity observations can be inaccurate when the instrument crosses a strong vertical temperature gradient. This kind of problem has been extensively studied and corrected for the standard ship-based CTDs in the past. Lueck and Picklo (1990) have proposed a numerical model of the thermal inertia of the Sea-Bird conductivity cells. Analyzing temperature-salinity diagrams, Morison et al. (1994) have estimated the correction to be applied to the thermal lag error. However the Sea-Bird CTD probes installed on gliders do have additional problems that make it difficult to apply the traditional corrections. In particular, (i) they are not always pumped, so the water flux velocity within the sensors depends on the glider velocity, and (ii) the sampling frequency on gliders is reduced compared to ship-based CTD casts (0.5 Hz vs 24 Hz). Garau et al. (2011) have therefore proposed a new method for the thermal lag correction for unpumped CTDs on gliders, which takes into account the variable velocity of the instrument itself.

Finally, it is worth mentioning that continuous time series of temperature and salinity at given depths have been obtained using moored instruments (see for instance the CIESM HydroChanges Programme, Schroeder et al., 2013). The accuracy of the temperature measurements can be as good as those of the CTD (at least the more recent ones), while the accuracy of the salinity measurements is lower (~0.01 psu). Also, in certain areas and depths, biofouling may be a critical issue in the stability of the sensor response with time. The temporal resolution of these data is generally very high, but the drawback of the mooring data is the discrete coverage in depth and that often present temporal discontinuities, either due to malfunctioning of the instruments or to discontinuities in the monitoring program. Also, it is possible to have drifts in the salinity measurements that should be corrected. Conversely, its very good temporal coverage (typical temporal resolution is 1 hour at least) allows to filter out the high frequency signals that can contaminate the climate signals. An additional problem of the mooring data is that they are not unified in the databases and most of the data have not been delivered to them.

The relative importance of all the above mentioned types of measurements has evolved with time (see Figure 15), from the beginning, when only bottles and MBT where used till the present when the Argo and glider data have become the largest contributors to the databases. This evolution also implies that the typical accuracy of the temperature and salinity measurements have evolved with time. This is especially important prior to the mid 60s when only bottle and MBT observations were available. Also, it is important to underline the fact that until the late 70s the only information for salinity was provided by bottles. This heterogeneity in the type of observations requires a multiplatform cross-calibration in order to homogenize the information but this is often not available.

Besides the instrumental error we have to consider the representativity errors, which are usually much larger than the instrumental ones. Typically, we are interested in monthly or yearly estimates of the temperature/salinity 3D field in order to infer the basin heat/salt content.

However, the measurements obtained at a given time may not be representative of the monthly/yrly average due to the influence of higher frequency processes (i.e. typically mesoscale features). As an example, the rms difference between hourly temperature observations at 100 m depth with respect to the monthly mean in a point located in the NW Mediterranean is 0.1 ºC, much larger than the nominal accuracy of the instrument. Also, the number of observations during summer is much higher than during winter for the ship-based observations, so a careful treatment of the seasonal heterogeneities should be applied in order to avoid a bias towards the summer values when year-averaged estimates are computed.

The spatial coverage is also inhomogeneous across the basin and in depth (Figure 16), mainly because the hydrographic data, except the Argo and some mooring data, were obtained in the framework of targeted studies and not with the aim of monitoring long term signals. Some regions, especially the southern part of the basin, are rarely sampled (or if sampled the data are not in public databases). Also, the upper layers are better sampled than the intermediate or deeper layers, because of technical limitations or specific goals of the surveys (e.g. lowering a CTD until the bottom floor requires significantly more operation time and for most biological applications it is not needed). Although the spatial and temporal variability of the temperature and salinity fields at those depths is much lower than in the upper layers, this lack of information can hamper the ability to identify long term trends in the deeper layers (Llasses et al., 2015).

Long-term ship-based repeated full-depth CTD stations are a useful tool to assess the relative importance of different layers in the water column in determining the variability of the integrate heat and salt contents. In the Mediterranean such a station is for instance the DYFAMED station (bottom depth 2000 m), located in the north-western Mediterranean, east of the Gulf of Lion. Schroeder et al. (2010) used these CTD data, collected almost every month by the Observatoire Océanologique de Villefranche sur Mer Service d'Observation (http://www.obs-vlfr.fr/sodyf/) to observe the temporal evolution of heat and salt contents in the Ligurian Sea, for the period 1995–2008 (see Figure 7 in Schroeder et al., 2010). Further analyses of the time series concerning the integral heat and salt contents and the contribution of the different layers to the total variance of heat and salt contents, reveal that the heat content changes of the whole water column are mostly explained (> 93% of explained variance) by the heat content changes of the first 1000 m, so that glider and Argo datasets could be sufficient to guarantee to capture this signal (obviously if a spatially dense enough network was established). However, concerning the salt content changes of the whole water column, the variance explained by the first 1000 m is only about 70%, so there is at least a 30% of the variance which is missed when not using full depth profiles. This is important to be taken into account when choosing the depth range for the assessment of the heat and salt content changes in the Mediterranean Sea.

Concerning the public databases where the hydrographic observations can be retrieved, the data obtained until the 90's were gathered and quality controlled in the frame of the MEDAR/MEDATLAS project (http://www.ifremer.fr/medar/). The SeaDataNet project (www.seadatanet.org/) also gathers the metadata of most of the available observations and provides a portal for contacting the data providers. Also, a large part of the data are available in the World Ocean Database (https://www.nodc.noaa.gov/OC5/WOD/pr_wod.html) and in the Hadley Center repository (http://hadobs.metoffice.com/en4/). The homogenized and quality

controlled Argo dataset can be retrieved from the Coriolis data center (www.coriolis.eu.org/), while the glider data are stored and disseminated trough the EGO-network (http://www.ego-network.org/) and the Coriolis data center. Moreover, there are other data portals that include, at least partially, those datasets (e.g. EMODNET, http://www.emodnet-physics.eu/; MonGOOS http://oceanobs.mongoos.eu/); MyOcean-Copernicus, http://marine.copernicus.eu/). It has to be reminded, though, that not all the existing data are included in those databases and that the consistency among databases is not always guaranteed. The data distributed to the data centers have not always passed a quality control prior to their submission (raw data are often submitted). Moreover, the quality control applied in the different center may significantly differ. Finally, the calibration of the instruments is a very important element to ensure consistency among different observations, but unfortunately it is not always correctly tracked.

6.3 – Computing heat and salt content

The available discrete temperature and salinity data have to be integrated in order to produce estimates of heat and salt content. Often, before computing basin averages the discrete data is interpolated on a regular grid to reduce the impact of the heterogeneous spatial distribution. The interpolated field is typically obtained through a weighted average whose general expression is:

(9)

That is, the interpolated value at point j is a combination of a first guess (or background field) at that location plus a weighted sum of the observed anomalies). There are different ways to define the weights w_i . They can be constant, a function of distance or obtained as the result of a statistical optimization (e.g. optimal interpolation, Bretherton et al., 1976). The way the background field is defined is also an open issue. Sometimes it is simply zero or a low order polynomial although often the background is defined as a long term average and depends on the month of the year and the spatial location.

A note of caution is required concerning the interpolation of the discrete data. The averaging often can mask problems with the observations, so it is very important that the observations considered have been quality controlled before the interpolation. Also, the analysis of the uncertainty linked to the interpolated product is not straightforward. There are different sources of uncertainty: observational and representatively errors, inhomogeneous data sampling and the limitations of the interpolation/averaging method. Considering all of them is a difficult task and often the uncertainties are underestimated. Also, some interpolation methods produce a formal error map that depends on the analysis choices and the selected parameters, so they are up to certain level arbitrary. Furthermore, once the 3D field is integrated to produce the heat or salt content value, that uncertainty should be propagated, but there is not a simple way of doing so (i.e. the spatial covariances between the errors at different grid points should be characterized somehow).

Some representative examples of different approaches followed to integrate the heat and salt content are the following. D'Ortenzio et al, (2005) and Houpert et al., (2014), in the study of the

seasonal mixed layer depth and the heat storage rate have followed a similar approach. Data from all available platforms has been used. In a first step the data was binned in 1º boxes, then the median was obtained for each box and month and a spatial filter was applied to remove small scales. The uncertainty is estimated from the standard deviation (std) of temperature at each box and then propagated into the heat storage rate estimate using a MonteCarlo approach.

Other authors have generated monthly 3D maps of temperature and salinity using algorithms based on optimal interpolation (e.g. Rixen et al., 2005; Llasses et al. 2015; and Jordà et al., 2016). In those cases the data from all the available platforms were used and interpolated in a ~0.2º grid. The background was generated from the same dataset and kept either constant or evolving on time. The uncertainty analysis is generated as the formal OI error (Rixen et al., 2005) or more accurately using surrogate data from a numerical model (Jordà et al., 2016). It must be mentioned that there are also global products that cover the Mediterranean area and which provide gridded 3D maps at a spatial resolution of 1º like the one from the JMST Center (rda.ucar.edu/datasets/ds285.3/, Ishii and Kimoto, 2009) or the Hadley Center EN4 (hadobs.metoffice.com/en4, Good et al., 2013).

In order to illustrate the type of results that are provided by the above mentioned products the Mediterranean 3D basin averaged temperature and salinity time series obtained from Jordà et al. (2016) are shown in Figure 17 along with the uncertainty estimates. It can be seen that the uncertainties before the '80s are larger than the interannual variability, thus suggesting that probably there are not enough observations to characterize that variability. After the '80s the temperature variability can be better tracked and a warming trend seems clearly captured. The salinity changes are typically more difficult to capture and only after year 2000, with the contribution of Argo data the uncertainty range could be narrowed. The relatively high uncertainties in the salinity are due to two factors: first, there are less salinity observations (i.e. until the 80s salinity was measured using bottles); second, the salinity field is dominated by short scale features (i.e. the variations at short spatial scales <~50-100 km are more important that those at longer spatial scales), so it is more difficult to identify the large scale features with discrete measurements (Llasses et al., 2015).

6.4 Observational estimates of the water content.

Changes in the volume of the basin can be obtained from two main sources of information. On one side there are the coastal tide gauges, which measure sea level relative to the coast. Some of the records span more than 100 years (e.g. the Marseille tide gauge, Wöppelmann et al., 2014) and for the last decades there is a good spatial coverage along the northern side of the basin. Conversely, very few data along the southern coast is included in the databases. The accuracy of hourly measurements is ~1 mm and for the monthly averages the accuracy is higher than 0.01 mm. Most of the tide gauge data are available through the Permanent Service for Mean Sea Level (www.psmsl.org, Holgate et al., 2013). The tide gauge data has to be quality controlled to avoid spurious peaks and changes in the reference level as well as to remove the effects of land movements. The drawbacks of this type of data is that they provide observations at discrete locations which are not necessarily representative of the basin average.

The other main source of information concerning the volume of the basin is satellite altimetry, which is operational since 1993, and measures sea level relative to the geoid along given tracks. Then the data from different satellites can be merged and interpolated on regular grids to produce weekly maps of sea level anomalies. Along-track and gridded products are distributed by several data centers (e.g. AVISO, www.aviso.altimetry.fr ; CSIRO, http://www.cmar.csiro.au/sealevel/sl_data_cmar.html; University of Colorado, http://sealevel.colorado.edu/). The accuracy of along-track is estimated to be ~2 cm while the gridded maps accuracy is estimated to be ~2- 4 cm for the weekly maps. It has to be noted, though, that the number of available satellites has not been constant in time, so the accuracy of the maps was not constant either. Also, for long-term trend estimates there can be an error due to instrumental and processing errors of ~1 mm/yr at regional scales (Henry et al., 2014).

The long time coverage of the tide gauge data can be combined with the good spatial coverage of altimetry in order to produce long time series of basin averaged sea level. This has been done using reduced order optimal interpolation techniques (e.g. Calafat and Gomis, 2009, Meyssignac et al., 2011, Calafat and Jordà, 2011). These techniques combine the spatial covariance patterns of sea level variability characterized from satellite altimetry or models with the time variability provided by tide gauges. Using this approach the basin averaged sea level for the last six decades has been reconstructed and the associated uncertainty has been estimated in less than 1 cm for the monthly values (Calafat and Jordà , 2011). Then, using hydrographic data to quantify the three first terms in the RHS of equation (8) the time series of water content can be derived (Figure 18, adapted from Jordà and Gomis, 2013).

The changes in water mass content can be also derived using gravimetric observations. The GRACE mission measures changes in the gravity field which in turn can be converted, after several corrections, into mass variations. In order to reduce the noise in the mass estimates a smoothing filter with a characteristic length scale of ~300-500 km is typically applied, so the spatial resolution of the GRACE data products is relatively coarse. Also, the Mediterranean region suffers from the handicap of being surrounded by a complex coastline which can contaminate the solution (e.g. Calafat et al., 2010; Fenoglio-Marc et al., 2012). Different products derived from GRACE can be obtained from GRACE-Tellus data center (http://grace.jpl.nasa.gov/). The accuracy of the monthly maps is estimated to be ~1 cm of equivalent water thickness (Johnson and Chambers, 2013), although a detailed estimated of the products accuracy for the Mediterranean has not been yet conducted. Once the mass estimates are available, they have to be corrected for the changes in the salt mass content in order to infer the changes in the water content (Eq. 8). Unfortunately, the uncertainties in the salt content change are large: the uncertainty for the salt content change during the period 2004-2010 is $\pm 0.32\ 10^6$ kg/s (Figure 17), which translated to equivalent water thickness would be $\pm$ 3.9 mm/yr much larger than the estimated mass change ~ 1 mm/yr (Calafat et al., 2010).

6.5 Contribution of numerical modelling

The numerical models are tools that can also give a different insight in the evolution of the heat and mass content of the Mediterranean. Numerical simulations covering the whole Mediterranean run with different numerical codes, with several spatial resolutions (from 1/6º

to 1/36º) and temporal coverage (from 20 to 60 years) are presently available in different repositories (e.g. www.medcordex.eu; Ruti et al., 2015; http://marine.copernicus.eu/). Here we specifically consider free-surface 3D primitive equation models which provide temperature, salinity and sea level fields as outputs. As a note of caution it has to be mentioned that the state-of-the-art models for the Mediterranean use the Boussinesq approximation. This implies that models cannot reproduce the effects of a net thermal expansion, so such effect has to be added a posteriori as a spatially constant but time varying factor (Greatbach 1994). Besides this, the main limitations for the realism of modelled sea level is that most models usually does not account for the atmospheric pressure effects and that the contribution of the global ocean must be accurately prescribed through the lateral boundary conditions.

The models distribute inside the basin the heat and mass that is introduced through the surface and the lateral boundary conditions. The internal dynamics generated by the model can partially modify the air-sea fluxes through changes in the sea surface temperature and/or salinity, the characteristics of the lateral outflow through the Strait of Gibraltar and/or the Dardanelles Strait and up to certain extent the inflow in those straits. However, in general, it is found that the evolution of the model simulations is strongly constrained by the forcings imposed at the boundary conditions (e.g. Harzallah et al., 2015). Also, imbalances between the initial conditions, the forcing through the sea surface and/or the lateral boundaries can lead to spurious long term drifts that may be misinterpreted as climate change signals. In summary, if the numerical models were correctly configured and the forcing fields were accurate enough, they could provide an accurate representation of the evolution of heat and mass content and the mechanisms behind it. Unfortunately those two assumptions are difficult to fulfill and at present the numerical estimates of the Mediterranean heat and mass content are not considered more reliable than observational based estimates. However, it is worth mentioning that several international initiatives (e.g MedCORDEX, ENVIMED/TANGRAM) are devoting lots of efforts to advance in this field in order to produce more accurate and reliable numerical simulations.

A complementary tool that merges the benefits of numerical model with observational-based estimates are the ocean reanalyses. By reanalysis we refer to numerical simulation in which a data assimilation procedure has been implemented to incorporate the information from observations into the model trajectory. The reanalyses can be viewed as a sophisticated way of interpolating data providing, in statistical sense, the best estimate of the true state of the ocean. Therefore, in theory, they should provide a more realistic picture of the ocean state that statistically based interpolations as those discussed in section 6.3. However, they also suffer from some of the above mentioned problems that models have concerning long term drifts and the impact of errors in the forcing fields as well as the lack of data to constrain the model in certain regions and/or periods. For the Mediterranean, there exists only a limited number of reanalyses (Adani et al., 2011; Hamon et al., 2015) and up to our knowledge no multi-product assessment has been performed yet. In any case, they are a promising alternative for a better estimate of the Mediterranean heat and mass budgets.

6.6 Perspectives

The recent improvements in the observational networks, especially since the development of the Argo program in the Mediterranean starting in the early 2000s, has allowed a significant

reduction of the uncertainties associated to heat and mass content estimates in the Mediterranean. However, some improvements could be still brought through a coordinated effort for the collection and homogenization of hydrographic data that are not yet in the public databases. In that sense it is important to highlight the existing discrepancies in the historical data collected in different databases. A careful review and homogenization of the past observations including a tracking of the calibration procedures would also be very beneficial, especially for those observations obtained in low-sampled areas or periods.

Concerning ship-based hydrography, until now the Mediterranean has been sampled at irregular intervals during national expeditions, while it has been neglected by the global sampling programs such as WOCE and GO-SHIP (www.go-ship.org). Med-SHIP is a new program that aims at filling this gap, by regularly sampling the Mediterranean along two meridional transects in each basin (western and eastern, to be repeated every 3 years) and one zonal transect from the Strait of Gibraltar to the easternmost Mediterranean (to be repeated every 6 years (last repetition in 2011; see Tanhua et al., 2013) and next one planned for 2017-2018). In particular the zonal transect, repeated on a low-frequency basis is thought to allow the assessment of long-term variations in the heat and salt budgets, with a focus on the deeper layers that are less influenced by small-scale features (Schroeder et al., 2015). It is important to mention that Med-SHIP aims at following GO-SHIP standards, which include a fully open-access data policy.

Regarding the distribute network of open ocean moorings there are different initiatives that gather some of the existing data (a large fraction of the mooring data is kept by the institutions and not held in the databases). For instance, the FIX03 program (http://www.fixo3.eu/observatory/) includes 10 sites in the Mediterranean that obtain hydrographic observations. The data is available in the data center of each observatory. Also, a complementary network is the CIESM HydroChanges Programme, which its ultimate goal is to detect long-term variations, abrupt changes and events, especially of the deep waters. An special concern in HydroChanges is to ensure a careful calibration of the existing sensors on long-term observatories especially for those measuring at large depths where the variations are much smaller than in the upper layers. Very importantly, future commitments of the HydroChanges community includes the set-up of a common database and metadatabase, or alternatively, to promote the submission of the collected data to existing databases (e.g., the Mediterranean in-situ TAC, MonGOOS). However, as happens with the ship based data, most of the existing data are not in the databases yet.

A more important issue that has to be mentioned is that, with the exception of Argo, there are few long–term monitoring programs that can allow a correct observation of climate signals. Therefore, a first concern is that the Mediterranean Argo network should be maintained and sustained. Several Mediterranean countries and the Euro-Argo ERIC (www.euro-argo.eu) are already committed to guarantee this sustainability. Besides this, an internationally coordinated effort that ensure a regular monitoring of key regions of the Mediterranean could allow a much better estimate of the Mediterranean mass and heat content evolution. This also includes the deployment of floats capable of collecting data as deep as 4000 dbar in selected areas of the Mediterranean to measure the variability of the heat and salt content below 2000 dbar. The

1601 determination of those key regions would require a devoted study that exceeds the goals of this
1602 paper.
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1604
1605 **7- 21st Century scenarios for the budget components**
1606 ***7.1-Overview***
1607 An important concern about the Mediterranean heat and mass budget is the future evolution
1608 of their different components in a context of global warming. The Mediterranean is located at
1609 the frontier between the arid climate of North Africa and the continental European climate. So,
1610 small perturbations of the conditions either in Europe or in Africa may induce significant changes
1611 in the Mediterranean. This section aims at giving a short overview of the projected changes in
1612 the different components of the Mediterranean heat and mass budgets. These projections are
1613 based on numerical simulations that aim to reproduce the evolution of the climate under certain
1614 scenarios of the future greenhouse gases (GHG) concentrations. On one hand, the
1615 Mediterranean is not isolated from the global climate system, so in order to properly simulate
1616 the evolution of this region it is important to simulate the evolution of the whole globe. In this
1617 sense, the global climate models (GCMs) are the first required step. However, on the other hand
1618 the Mediterranean has several characteristics (i.e. large spectrum of mesoscale processes, water
1619 mass formation, strait dynamics,...) that require a relatively high spatial resolution in the models
1620 in order to produce a realistic behaviour. Therefore, regional climate models (RCMs) nested into
1621 GCMs are necessary to properly model the Mediterranean and to produce regional projections.

1622 It is also important to highlight that the projections are inherently affected by a significant degree
1623 of uncertainty. First, they are based on assumptions on the unknown future GHG concentrations.
1624 Second, being the result of numerical simulations, the projections are affected by uncertainties
1625 linked to the accuracy of the modelling system. In particular to uncertainties in the GCMs and
1626 also in the RCMs. And finally, the climate system has a natural multidecadal variability which is
1627 not linked to the GHGs concentrations and that is reproduced by the models. In other words,
1628 some changes simulated by the models may be due to natural variations and not to changes in
1629 the GHGs concentrations. This has to be taken into account when the projection only span few
1630 decades, as the changes due to GHGs and those due to natural variations are more difficult to
1631 distinguish. A common technique to characterize the uncertainties in the modelling system and
1632 to filter out the natural variations is to produce a large number of simulations forced by the same
1633 GHG concentrations. Then, the ensemble average is considered to be the best approximation to
1634 the real evolution while the ensemble spread is a measure of the uncertainties linked to the
1635 models and the natural variability. Unfortunately, the computational cost of the regional
1636 simulations is high, so at present the number of Mediterranean regional projections is reduced
1637 and the results must be interpreted with care.

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1639
1640 ***7.2- Water fluxes through the surface and the lateral boundaries***
1641 Table 7 summarizes projected changes of the different component of the freshwater budget
1642 obtained by different authors. Most of the compiled studies predict a decrease in precipitation
1643 over the basin with a reduction between -5 and -15% although some projections reach up to a -
1644 28 % reduction. Also, the evaporation tends to increase a similar amount with some models

projecting up to 18 % more evaporation. River runoff and Black Sea water net inflow both decrease respectively up to -87 % and -102 %. It is interesting to notice that the latter means that some studies even predict that the Black Sea would become an evaporative basin and thus the net water flow through the Dardanelles Strait would change its direction (from the Mediterranean to the Black Sea). Finally, the net water inflow through Gibraltar could increase up to 60 %, as the result of the increase in the freshwater deficit. Adloff et al. (2015) have also shown that the changes in the components of the water budget are strongly linked to the chosen socio-economic scenario: for higher GHG emissions, the water flux response is larger. [Note that the other studies referenced in Table 7 are based on a single scenario].

7.3- Heat fluxes through the surface and the lateral boundaries

Table 8 compiles the projected changes in the heat fluxes through the surface and the Strait of Gibraltar. The surface heat loss of the Mediterranean is decreasing in all available climate projections. This positive surface heat change ranges from +25 % to +118 %, which means that some model predict that the Mediterranean could even gain heat through the surface in the future. As for the water fluxes component, the changes in surface heat fluxes are tightly related with the GHG concentrations in the scenarios. It is worth noting that results from the CIRCE project compiled in both Dubois et al (2012) and Gualdi et al. (2013) refer to changes over the 2021-2050 period. Much larger changes are expected for the period 2061-2099 if the CIRCE simulations would have been performed up to 2099. Concerning the net heat exchange through Gibraltar, the ensemble of the CIRCE simulations displays a range between -11 % and +3 % (2021-2050 vs. 1961-1990), while the simulations carried up to the end of the 21st century predict an increase from 0 % to +40 % (2070-2099 vs. 1961-1990). The increase in the heat flux is linked to a combination of (i) an enhanced water flux required to compensate the increase in the water losses through the surface inside the Mediterranean and (ii) the increase in the temperature of the Atlantic waters flowing into the Mediterranean.

7.4- Heat and mass content changes

Most of the studies on the future climate of the Mediterranean Sea have projected a significant warming and salinization at the end of the 21st century (Table 9). This is consistent with the above mentioned results on the surface fluxes. An increase of the global temperature would also lead to a reduction of the heat loss through the sea surface and to an increase of the temperatures of the incoming Atlantic waters through the Strait of Gibraltar. This altogether would lead to an increase of the Mediterranean heat content. Also, an increase of the water loss trough the sea surface would lead to a salinization of the basin, since more salty water will enter the basin to compensate the fresh water deficit. However, the expected salinity of the Atlantic waters is not a straightforward issue. On one hand, some global models project an increase in the northeast Atlantic salinity due to increase evaporation in that region. On the other hand other models suggest that the ice melting in the Arctic would imply a water freshening that can be advected towards the northeast Atlantic. In the latter case the waters entering the Mediterranean through the Strait of Gibraltar may be fresher and could compensate, at least partially, the effects of enhanced evaporation.

Although most models agree on the general picture, the projected amount of change is quite diverse among the different studies. Somot et al. (2006), using a single simulation forced under

the SRES-A2 scenario, have projected an increase in the basin averaged temperature (T3D) of 1.5°C and of 0.23 psu in the salinity (S3D) in 2099 with respect to the values in year 2000. Carillo et al. (2012), performed two simulations for the period 1950-2050 under the scenario SRES-A1b. The difference between the two simulations was on the Atlantic boundary conditions, specially on the salinity. The simulated changes for the period 2001-2050 with respect to the period 1950-2000 was +0.38°C and +0.42°C for T3D and +0.03 and -0.08 psu for S3D, for the two simulations respectively. Adloff et al. (2015) performed an ensemble of six projections under three different GHG scenarios (B1/A1b/A2) and with varying Atlantic, riverine and atmospheric conditions. Their results projected an averaged change for the period 2070-2099 with respect to 1961-1990 of +0.93°C to +1.35°C for the temperature and +0.28 and +0.52 psu for salinity (Table 9). Finally, Macías et al. (2015) have projected an increase for the period 2096-2099 with respect to 2014-2017 of +1.30 and +2.50 ºC (+0.6 and 0.9 psu) for the temperature (salinity) using respectively the scenario RCP4.5 and RCP8.5. Part of the discrepancies among these studies are linked to the model characteristics and the way they simulate basic features of the basin. Another part should be attributed to the different configuration of the experiment (model chosen to provide boundary conditions and/or GHG scenario) and the diagnostics performed (analysed period).

The sea level in the Mediterranean will be affected by changes in the thermal content of the basin through the thermal expansion. The changes in the salt content are expected to have a minor effect (Jordà and Gomis, 2013). Concerning the thermal expansion, Carillo et al. (2012) computed that thermosteric sea level would reach 5cm and 7cm by 2050, although one should keep in mind that the largest part of the oceanic response to climate change appears in the second half of the 21st century. The simulations of Adloff et al. (2015) projected a thermal expansion from 45 cm to 61 cm by the end of the 21st century. However, it has to be kept in mind that the Mediterranean sea level is strongly affected by the changes in the nearby Atlantic (see for instance Tsimplis et al., 2013). This link with the Atlantic has clear consequences for the future evolution of the Mediterranean. First, the changes in the Mediterranean sea level will be modulated by the Atlantic: if the Mediterranean sea level rose much more than the nearby Atlantic the increase in the along-strait pressure in Gibraltar will translate into a reduction of the net flow in the basin and thus a reduction of sea level. Second, in a similar way sea level changes in the Atlantic will be transferred to the Mediterranean. In that sense, recent projections for the nearby Atlantic (Bouttes et al., 2013) show that sea level in that region will be ~10 cm higher than the global sea level rise. This, in turn, it is projected to be between 50 and 150 cm by 2100 due to the combination of thermal expansion and water addition from land based ice melting. Therefore, it should be expected that the Mediterranean sea level will rise much more than what is projected considering only the basin thermal expansion (e.g. Jordà et al., 2012).

7.5- Perspectives

As important as the projected changes is the uncertainty associated to those projections. In consequence it is of substantial importance to get accurate estimates of them. As mentioned above, one source of uncertainty is related to the modelling system itself. Mediterranean models do still have room for improvement and several efforts are being carried out in that line (e.g. MedCORDEX initiative , Ruti et al., 2015 ; the ENVIMED-TANGRAM project). A better representation of the present climate would increase the robustness of the future projections. Another source of uncertainty is related to the information introduced through the boundary

conditions, mainly the projected changes in the nearby Atlantic and in the atmospheric forcing. In order to tackle this it is necessary to have an ensemble of simulations forced by a range of feasible conditions at the boundaries. Also, such ensemble could be used to filter out the influence of the natural variability on the projected changes. And last but not least, it has to be kept in mind that the future GHG concentrations are unknown, so different scenarios should be taken into account. All this requirements demand a huge amount of human and computational resources and can only be achieved through sustained and coordinated efforts at the international level. In this sense, to maintain programs like MedCORDEX is essential to ensure the improvement of Mediterranean future projections.

8- Discussion

A key question that can be addressed is whether the attained accuracy of the different components of the Mediterranean Sea heat and mass budgets is sufficient. As mentioned in the introduction, the answer obviously depends on the application where those estimates have to be used. In our case, we are primarily interested in determining if we can close the heat and mass budgets with the state-of-the-art estimates and to assess in which components there are more room for improvements.

A first consideration is that the budgets described in equations (1-3) are independent of the timescale considered and the relations must hold for any averaging period of time. In practice, though, the chosen period may affect the reliability of the different components. For instance, when a long period is considered, the heat and mass content change in equations (1-3) is closer to zero (i.e. the basin is close to the equilibrium), which would simplify the closure of the budget. However, the estimates of the different components can be unreliable for past periods when the observational systems were scarce. Conversely, it is possible to find a short period where the quality of the different components is better estimated (i.e. because of having a better observational system). However in that case the content change cannot usually be neglected and has to be estimated somehow. Moreover, the accuracy of the trends (i.e. the rate of content change) is directly linked to the record length, so shorter periods are prone to show higher uncertainties.

A second issue closely linked to the closure of the budget is the quantification of the uncertainty associated to each component. This will determine the accuracy of the closure and the capability to indirectly estimate the different components. Ideally, the characterization of the uncertainty should involve the comparison with independent observations. Unfortunately this is almost never possible due to the lack of such observations, so other alternatives are envisaged. Sometimes the temporal standard deviation is used as a measure of the error in the mean. Strictly speaking, the statistical uncertainty of the average of a collection of N samples $\{X_i\}$ is defined as $std(\{X_i\})/N^{0.5}$ when the samples are assumed to be uncorrelated. Therefore, this quantity would measure the inaccuracies in the estimate of the mean provided we have a short record (i.e. that is what we should use if we wanted to estimate the long term average of a quantity when only few years of data are available). However two important remarks are needed. The first one is that this would be correct only if the samples are uncorrelated. In practice this requirement is almost never satisfied as the geophysical variables have a long term memory. For autocorrelated time series (i.e. those with low frequency signals on it), N should

be replaced by N*, the effective degrees of freedom. To determine N* is not straightforward and depends on the relation between the length of the record with respect to the characteristic time scales on it. Therefore, in those cases where the records are short the uncertainties obtained from the std are probably underestimated. The second remark is that the std is only a measure of the statistical uncertainty associated to the averaging, but by no means can be viewed as a measure of the total uncertainty. The latter should include also the uncertainties linked to the accuracy of the data and/or the methodological approach used to estimate the derived quantities (i.e. the fluxes). This part of the uncertainty is not easy to quantify but it is very important as it is usually the dominant contributor. A common approach has been to use an ensemble of products (e.g. surface heat fluxes from different models) and consider that the ensemble spread encompasses the observational and methodological errors. However, this approach can highly underestimate the uncertainty: most models/reanalyses are not independent as they are based on similar codes and the observations used to constrain and/or to calibrate them are often the same. Moreover, the number of available products is typically small so the inter-product spread is probably underestimating the actual spread. Alternatively other approaches base on sensitivity experiments using synthetic data from numerical models to test the accuracy of the methodology and the propagation of the observational and representativity errors to the derived quantities. This approach can lead to more realistic estimates of the uncertainties. Some examples can be found in Sánchez-Román et al. (2009) for the estimates of the Gibraltar outflow or Llasses et al. (2015) for the Mediterranean heat and salt content. Unfortunately this strategy is not always easy to implement and at present is not available for most of the components of the budgets.

The next step is to summarize the different estimates of the components in (1-3) along with their associated uncertainties in order to evaluate in which cases the budgets can be closed. For illustrative purposes we consider a long term average for the period 1980-2010 as well as the short period 2005-2010. The summary of the results for the heat budget (eq. 1) is presented in Table 10. The long-term mean surface heat flux is estimated to be -3 ± 8 W/m^2 from ship/satellite observation based (Sánchez-Gómez et al., 2011) and where the uncertainty is computed as the temporal std. This value also encompasses the -7 W/m^2 value obtained by Pettenuzzo et al. (2010) from atmospheric reanalyses..It has to be noted that the selection of these two estimates has been done implicitely considering the Gibraltar constraint. Without that knowledge, the actual uncertainty in the mean surface heat flux would be much larger. For the period 2005-2010 we have analysed the results from the MedCORDEX ensemble of regional ocean models. The results indicate that during that period there was no significant difference with the long term mean or the intermodel spread. For the net heat flux through the Strait of Gibraltar, the estimated value is 4.5 ± 4.1 W/m^2. This value arises from a numerical simulation, as far as no direct observations are available, and the uncertainty has been estimated as half the std of the subinertial time series. Alternatively, assuming an uncertainty of 15 % in the inflow, outflow or the temperatures would lead to a similar degree of uncertainty. For the period 2005-2010 there is no information that can be used to assess significant changes in the heat flux through Gibraltar. For the Dardanelles Strait we use the estimate presented in Section 4.5 of 0.1 ± 0.2 W/m^2, which is based on published values for the temperature and transport of the two-layer system. The uncertainties have been derived assuming an uncertainty of 2ºC in the temperatures and 200 km^3/yr in the transport. The latter is based on the observed std of the

low frequency time series. Again, there is no data that allow to assess a significant change during the 2005-2010 period. For the rivers, the estimate performed in section 5.5 was 0.6 ± 0.4 W/m^2 where the uncertainty has been obtained assuming an uncertainty in the river runoff of 15% and 5 ºC in the river temperature. No significant difference is observed for the period 2005-2010. For the heat content change the value estimated for the period 1980-2010 is 0.3 ± 0.3 W/m^2 while for 2005-2010 is one order of magnitude larger 3.5 ± 3.0 W/m^2.

The summary of the water mass budget components is presented in Table 11. The averaged water flux through the surface is estimated to be -0.8 ± 0.2 m/yr based on published results (see section 2.5) where the uncertainty comes from the interproduct spread. For the period 2005-2010, the reanalyses and RCM models show a slight increase in the water loss, being -0.9 ± 0.2 m/yr. The Gibraltar net water flux is estimated to be 0.8 ± 0.2 m/yr based on historical estimates and with the uncertainty computed as the spread of those estimates. For the shorter period 2005-2010, there is no direct estimates of the flux but based on the indirect computation of Soto-Navarro et al. (2010), the net water flux should have increased to 0.85 ± 0.2 m/yr during that period. The net water flux through the Dardanelles is estimated to be 0.1 ± 0.02 m/yr , with the uncertainty computed also as the spread of historical estimates. For the period 2005-2010 there is no information to assess eventual changes in that component. The river contribution is considered to be 0.2 ± 0.01 m/yr and none of the independent estimates show significant changes on the mean during the period 2005-2010. The change in the freshwater content for both periods is not significantly different from 0: 0.5 ± 1.0 mm/yr and 1 ± 3 mm/yr, respectively (note that these values refer to the change in the freshwater content and not the total sea level; i.e. the thermal expansion effects are not considered).

Finally, the different components of the salt mass budget are presented in Table 12. The mean net salt flux through Gibraltar is estimated to be -1.5 ± 6.5 10^{6}kg/s and no information is available to assess if any change existed during 2005-2010. The salt flux through the Dardanelles Strait is estimated to be 0.2 ± 0.2 10^{6}kg/s with the uncertainty computed assuming an error of 2 psu in the salinity of the exchanged waters (see section 4.5). Again, no information is available to assess eventual changes during the period 2005-2010. The salt content change is estimated to be 0.1 ± 0.1 10^{6}kg/s for the period 1980-2010 and to increase up to 0.3 ± 0.2 10^6 kg/s during 2005-2010.

Comparing the direct estimates of the different components (Tables 10-12) it can be seen that the heat, water mass and salt mass budgets are closed within the uncertainty limits. This means that the each component could be indirectly estimated from the other components using the equations (1)-(3). However, this would only make sense for the component showing the largest uncertainty. In particular, for the period 1980-2010, the surface heat fluxes can be better constrained using this approach (uncertainty reduced from ± 8.0 W/m^2 to ± 4.3 W/m^2), as well as the net water flux through the Strait of Gibraltar (uncertainty reduced from ± 0.4 m/yr to ± 0.2 m/yr; that is from ± 0.03 Sv to ± 0.02 Sv), and the net salt flux through the Strait of Gibraltar (uncertainty reduced from ± 6.5 10^6 kg/s to 0.2 10^6 kg/s). For the shorter period 2005-2010 the same components are better estimated indirectly than from direct computation. It is worth noting that for that shorter period, the common assumption that the Gibraltar heat transport is mostly compensated by the heat trough the sea surface is not accurate. The changes in the basin

heat content have to be taken into account as far as they are comparable to the Gibraltar contribution.

The picture obtained for the heat budget shows that the dominant terms are the fluxes through the sea surface (-4.9 $\pm$ 4.3 W/m^2) and the fluxes through the Strait of Gibraltar (4.5 $\pm$ 4.1 W/m^2). The contribution of the heat content change (i.e. the temperature trend) is smaller but not negligible (0.3 $\pm$ 0.3 W/m^2) and should be taken into account when short periods (~few years) are considered. A new result is that the contribution from the rivers, albeit small is non-negligible (0.6 $\pm$ 0.4 W/m^2) and during some periods can be comparable to the heat fluxes through the Strait of Gibraltar (see section 5.5). The largest uncertainties in the heat budget are associated to the surface fluxes and the Gibraltar fluxes (~4 W/m^2). In relative terms, the uncertainties associated to the Dardanelles are also very high (i.e. roughly twice the estimated value). However, as that component represents a small contribution to the total budget those uncertainties are small in absolute terms. Concerning the water mass budget, the flux through the sea surface is the largest contributor (-0.8 $\pm$ 0.2 m/yr) and is compensated by a combination of the flux through Gibraltar (0.4 $\pm$ 0.2 m/yr), the river runoff (0.2 $\pm$ 0.01 m/yr) and the flux through the Dardanelles (0.1 $\pm$ 0.02 m/yr). It is worth noting that in this case all those components are of similar magnitude while the change in the water content (i.e. linked to the non-thermosteric sea level change) is much smaller. Concerning the uncertainties, those associated with the surface fluxes and the fluxes through Gibraltar are one order of magnitude larger than the others. Finally, for the salt mass budget, our estimates suggest that the contribution of the fluxes through the Dardanelles strait would be larger than the estimated salt content change, so the Gibraltar contribution would be negative (salt export to the Atlantic). Nevertheless it has to be noted that the uncertainties in the fluxes through the straits are large, so these results have to be taken with caution.

These results also indicate to which components more efforts should be devoted in the future. Although the budgets can be closed with the existing estimates, the associated uncertainties are still very large. It seems clear that the largest improvements in the characterization of the Mediterranean budgets would be brought by the reduction of the uncertainties in the surface heat fluxes and in the fluxes through the Strait of Gibraltar. For the surface fluxes more studies devoted to exploit the available in-situ data and to provide a reference dataset to evaluate the air-sea fluxes from different products are required. Complementary, a better monitoring of the Gibraltar Strait, even if it would require a large coordinated effort, would allow to better constrain the air-sea fluxes estimates and to better explain the changes in the salt content of the basin. However, some notes of caution are required. The net volume and salt transports are the difference of two very similar quantities and, therefore, their size is close to zero. Actually, if the uncertainty is identified with the error of the measurement, neither of them would be statistically different from zero (Table 4). The fact that these two net transports are more than one order magnitude less than the intervening terms in the summation makes them be very sensitive to small errors in the estimate of each term. Even when the net heat transport is obtained with the same procedure, it exhibits a clear difference with the net flow and salt transport: it is only a little smaller than the terms in the summation and its time-averaged value is slightly greater than the associated uncertainty. The reason for this difference stems from the

marked temperature difference between the Atlantic and the Mediterranean waters, which is greater than the salinity difference in relative terms. Thus, while the net salt transport is virtually zero, the heat transport is positive.

Another important consequence of the budget analysis performed before is that the Mediterranean can be used as a test case for the evaluation of air-sea heat flux products. The indirect estimate of the mean surface heat flux provides a strong constraint that can allow to elucidate which products are significantly biased. That constraint could even be refined with complementary monitoring of Gibraltar heat fluxes and, to a less extent, of the rivers and Dardanelles heat fluxes. Moreover, during the last years air-sea fluxes have been measured in several open sea locations (see section 2.5), which would provide additional constraints for the surface heat flux products once the available data is homogenized and quality controlled. Conversely, for the net surface water flux the indirect estimate cannot be considered an effective constraint as far as the associated uncertainties are too large. An accuracy of ~0.015 Sv in the estimates of net flow through the Strait of Gibraltar would be needed to improve the direct estimate of the surface water flux. In order to reach such value, an accuracy of ~0.01 Sv would be required for the inflow and outflow (i.e. ~ 1 % of the absolute value). Until that level of accuracy is attained in the estimates of the Gibraltar water fluxes, the only way to constrain the surface water flux products is through the development and exploitation of the open-sea observational network.

An open question that is usually discussed is if there should be a preferred period for the budgets closure. The results presented in this paper suggest that since 2004 the uncertainties in the heat, salt and water content change are reduced thanks to the new observing systems. However, no clear improvement has been brought to the other components of the budgets which in fact dominate the uncertainty. The estimates of the surface fluxes and the net fluxes through the Strait of Gibraltar have not been significantly improved during that period, so the accuracy of the budget closure is basically the same. It has to be noted that some observational systems that could bring new improvements to those fluxes have been deployed recently (e.g. surface current radar observations in Gibraltar (see 3.4) or air-sea flux measurements at open sea buoys (see 2.4)). Therefore, the period from 2014 onward could be potentially preferred but only once the new datasets have been used to improve the air-sea fluxes and the fluxes through the Strait of Gibraltar.

Finally, it is also worth noting that the indirect estimate of the heat and salt content is less accurate than what is obtained from in situ observations. The main reason is the large uncertainties associated to the surface heat flux and the heat and salt fluxes through the Strait of Gibraltar. An important consequence of this is that the heat and salt content changes obtained from numerical models would be less reliable than the observational estimates. Although the models can reproduce the heat and salt redistribution inside the Mediterranean, the changes in the total heat and salt content are strongly determined by the surface and lateral boundary conditions, which are highly uncertain. As an example, Vargas-Yáñez et al (2010) have estimated from in-situ observations a change in the western Mediterranean heat content of 0.3 W/m^2 for the period 1940-2000. This is an order of magnitude lower than the uncertainty associated to the mean surface heat flux used to force the ocean models. In consequence, the

characterization of temperature and salinity trends in the Mediterranean from models should be avoided unless until the uncertainties in the forcings are significantly reduced.

9- Conclusions

In this paper we have presented a review of the state-of-the-art of the understanding and quantification of the Mediterranean heat and mass (i.e. salt and water) budgets. The different components of the budgets have been described in detail and a critical review of the different methods and observational products that allow their quantification has been presented. Special emphasis has been put on the characterization of the associated uncertainties, and several proposals for the advance in the current knowledge are presented for each component of the budgets.

The *surface fluxes* are mainly characterized using atmospheric reanalyses or regional climate models as far as little observational data are available over the sea. The uncertainty associated to those fluxes is the most sensitive part of the heat and water mass budget as they are one of the dominant components of the budgets. In that sense, an increase of the observations of air-sea fluxes on the sea (i.e. with moored open sea buoys), as well as a coordinated effort to combine observational discrete data with numerical model outputs is thus recommended. The *fluxes across the Strait of Gibraltar* are also an important component of the budgets even if there are some important lacks. In particular, there are very few direct measurements of the inflow (and thus of the net water transport) and the heat and salt transports are not routinely monitored. However, and even with the technical challenges associated with the monitoring of such complex region, a sustained monitoring of the Gibraltar fluxes would be very beneficial as it would allow to better constrain the surface heat and water fluxes and to indirectly determine the changes in the Mediterranean salt content. The uncertainties associated to the *fluxes across the Dardanelles Strait* are even relative larger than for the Strait of Gibraltar. However, due to its small contribution to the total budget this is not a major shortcoming for the closure of the Mediterranean budgets. Nevertheless, it is obvious that they have a large impact at regional scale, so any improvement in the monitoring of those fluxes would lead to a better knowledge of the dynamics in the Eastern basin. The observations of the *riverine input* are often either outdated or with incomplete information on possible treatments applied to remove the impact of water usage. Also, about a 40% of the total flux cannot be estimated from observations. Different types of numerical models can complement the observational estimates producing high frequency time series as well as estimates in places where observations are not available. However they are strongly affected by uncertainties in the water fluxes over land. A first estimate of the heat flux associated to the river runoff has also been produced showing that rivers can significantly contribute to the heat budget. Concerning the *heat and salt content change*, the present day observational system has allowed a significant reduction of the uncertainties, mainly since the introduction of the Argo floats. However, it has been also identified that more efforts should be devoted to better integrate the existing information, especially that collected and stored at national level, which is not always available in the public databases. The changes in the water content are derived from estimates of total sea level

changes that have been corrected for the expansion of the water column (i.e. using hydrographic data). Alternatively, satellite observations of mass changes can be converted into water content changes using information on the salt content change. However, the large uncertainties associated to salt content estimates make this approach still useless in practice.

The state-of-the-art of the different components allow the budget closure in the range of the uncertainty. This allows to indirectly estimate some of the components of the budget with better accuracy than the direct estimate. In particular the Mediterranean has proven to be a suitable test case for the evaluation of air-sea heat flux products (but not for the water flux products). The accuracy of the indirect estimate of the long term mean is higher than the interproduct spread and could even be reduced if a monitoring of the Gibraltar heat flux was set up. Conversely, the indirect estimates of the heat and mass contents are more uncertain than the direct estimates. The consequence of this is that long term temperature and salinity trends obtained from numerical modelling must be call into question. The heat and salt content evolution in numerical models are strongly constrained by the surface and lateral forcings, which are not accurate enough to lead to accurate modelling of the temperature and salinity trends.

Finally, this review has revealed that the quantification of the uncertainty associated to each component should also be improved. In most cases it has only been assessed using the temporal std as a proxy of the error in the mean or using the spread of different products to account for methodological and observational errors. However, in most cases these approaches lead to an underestimation of the uncertainties.

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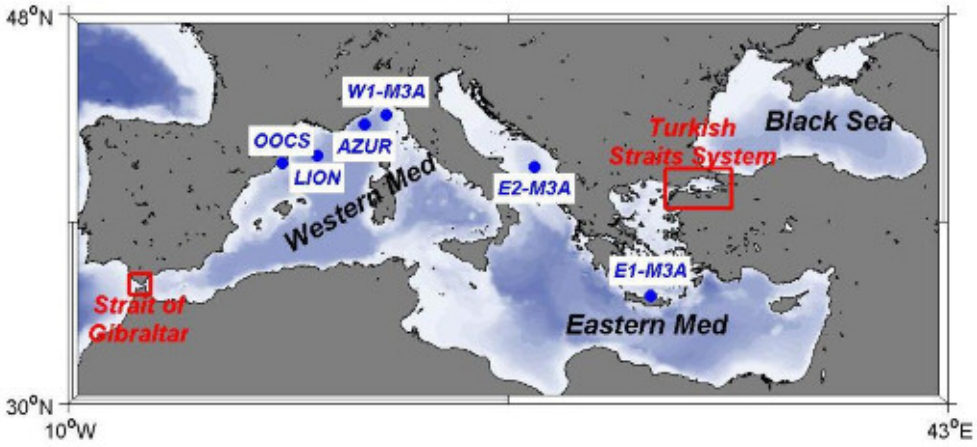

Figure 1: Mediterranean bathymetry with the location of the Strait of Gibraltar and the Turkish Straits System (for detail see Figures 4 and 9 respectively). The blue dots indicate the location of the open sea buoys described in section 2.

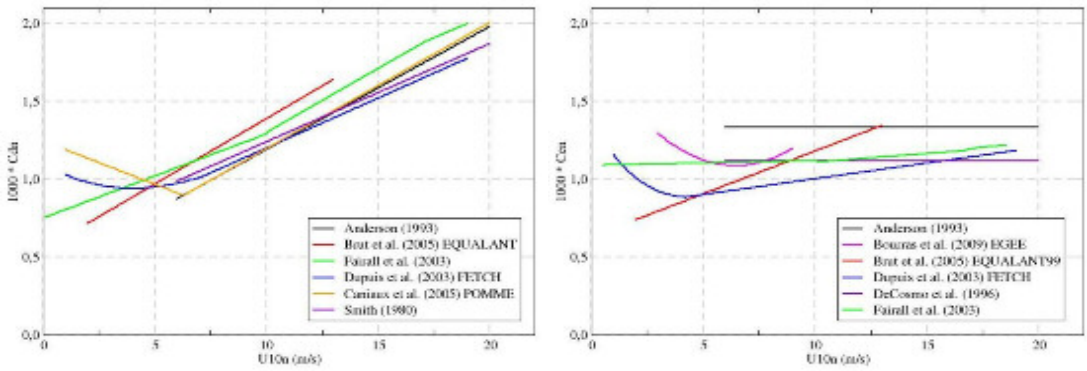

Figure 2: Exchange coefficient (multiplied by 1000) for the wind stress Cdn (left) and for the latent heat flux Cen (right) as a function of the wind velocity (at 10 m and in neutral stratification) for some recent parameterizations and in the range of sampled wind conditions.

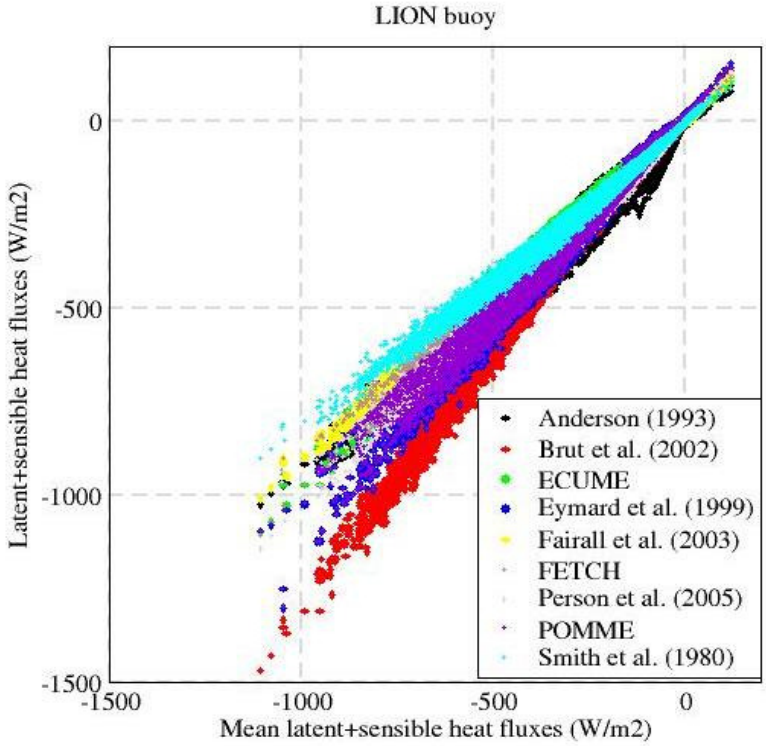

Figure 3: Spread of the sum of the latent and sensible heat fluxes calculated at the LION buoy in the north western Mediterranean Sea (period 2001-2013; N=68722) from nine flux bulk algorithms, as a function of the mean of the nine fluxes. Note that for a mean sensible plus latent heat flux of 1000 $Wm^{-2}$the spread is about 500 Wm^{-2}

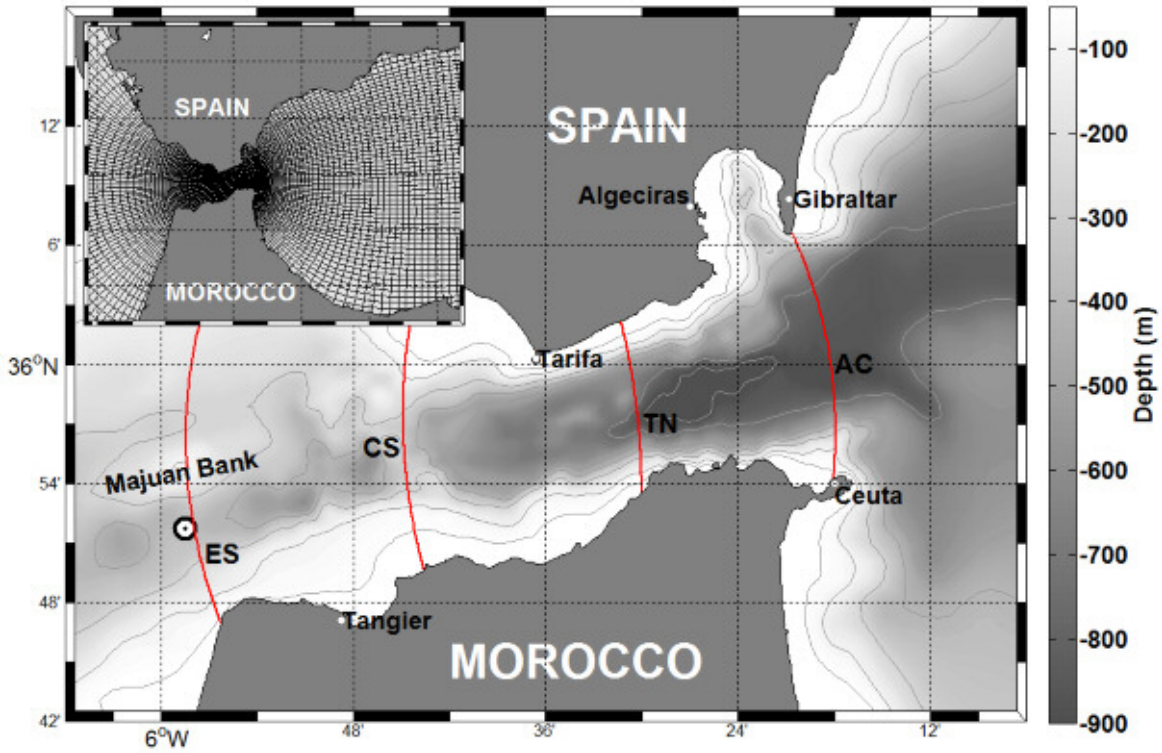

Figure 4. Map of the Strait of Gibraltar showing the bathymetry and the location of the relevant topographic features and sites cited in the text: ES, CS, TN, and AC indicate the position of Espartel sill, Camarinal sill, Tarifa Narrows and Gibraltar-Ceuta cross-strait sections, respectively. The monitoring site at Espartel sill is also indicated. The inset shows the numerical model domain used to accomplish the two hincasts discussed in the text.

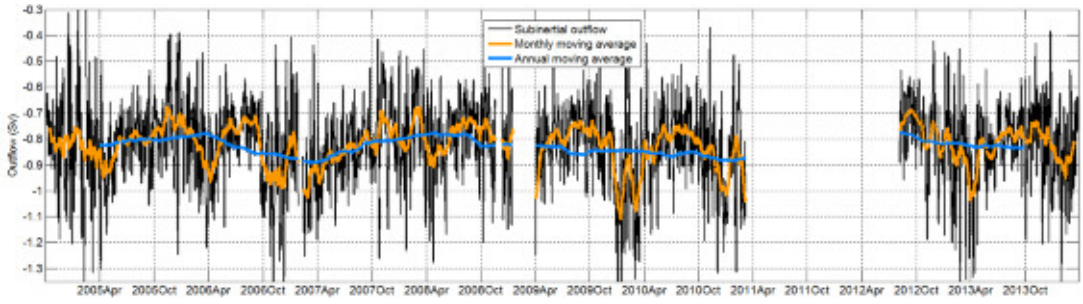

Figure 5.- Time series (black) of the outflow after removing tidal fluctuations by a low-pass 8-th order Butterworth filter, with pass and stop band frequencies of 38^{-1} and 28^{-1} h^{-1}, respectively. Two other versions with different degree of additional filtering are presented to illustrate the annual (orange) and interannual (blue) variability (adapted from Sammartino et al. (2015)).

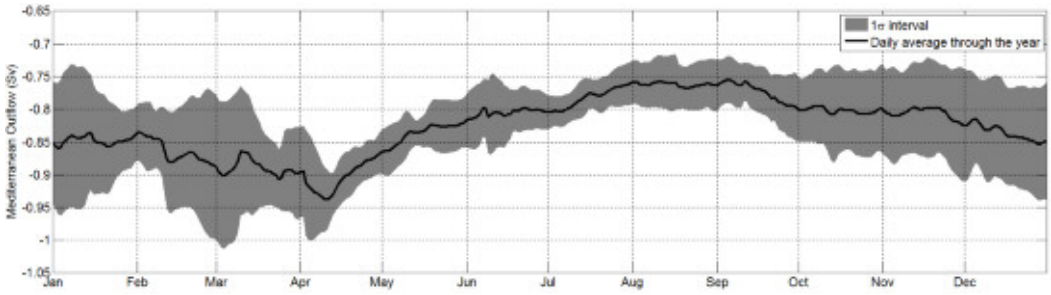

Figure 6.- Seasonal cycle of the outflow at Espartel. The black line is the mean value of the outflow in a given Julian day obtained by averaging the daily value of the outflow the same days of all the sampled years. Grey shade represents the standard deviation of the same statistics (adapted from Sammartino et al. (2015)).

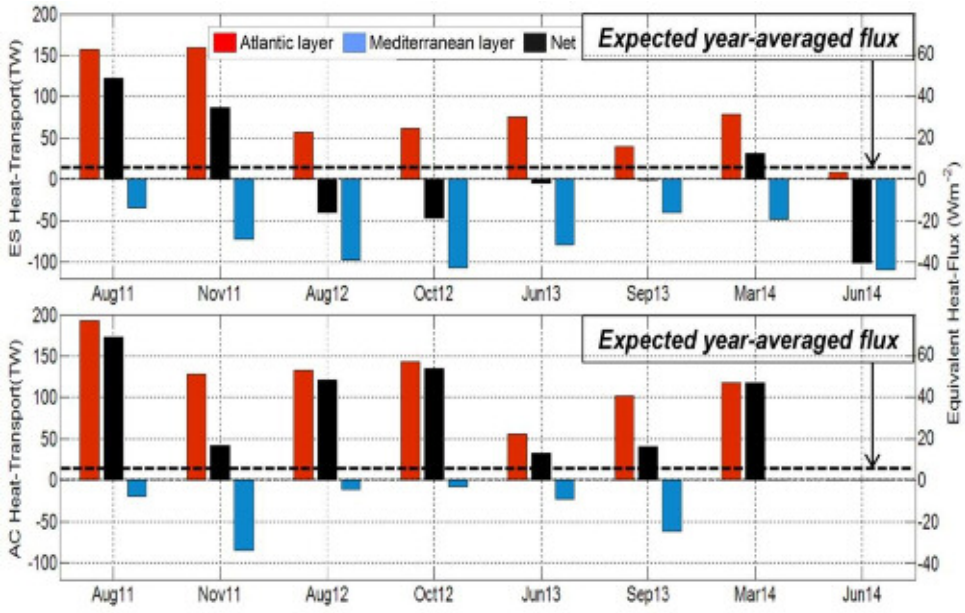

Figure 7. Heat transport computed from observations collected with a CTD probe that had a lowered ADCP incorporated, during different oceanographic cruises at the dates indicated in the x-axis. The top panel shows the transport across ES section and the bottom panel through the easternmost section of the SoG (AC, see Figure 1). The colored bars indicate the heat transport within each layer and their sum (net), being positive towards the Mediterranean Sea and negative towards the Atlantic Ocean. The dashed black line indicates the expected long-term mean according to values in the literature. The scale on the right shows the equivalent heat flux across the Mediterranean Sea surface to balance the heat transport through the SoG, which is computed by dividing the heat transport (left scale) by the area of the Mediterranean Sea.

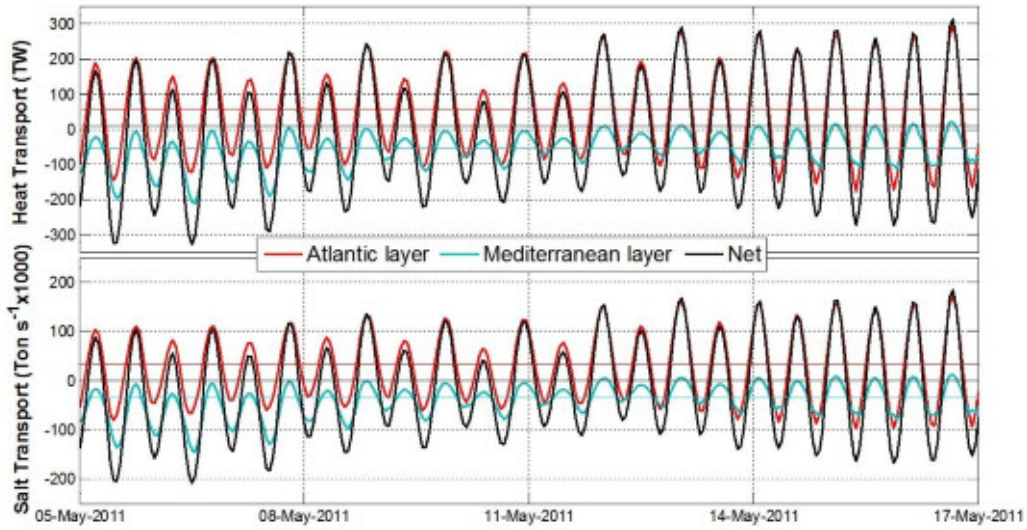

Figure 8.- Short fragment of the time series of heat (top) and salt (bottom) transports computed from the outputs of the numerical model at ES section. The time series include all ranges of variability, although the overwhelming prevalence of the semidiurnal tidal signal masks the variability at other frequencies (subinertial in this case). As in Figure 4, the color code indicate the heat and salt transport within each layer, which must not always be necessarily positive (towards the Mediterranean) in the Atlantic layer or negative (towards the Atlantic) in the Mediterranean layer. Horizontal lines indicate the time-average values of the transports with the same color code (see also Table 2).

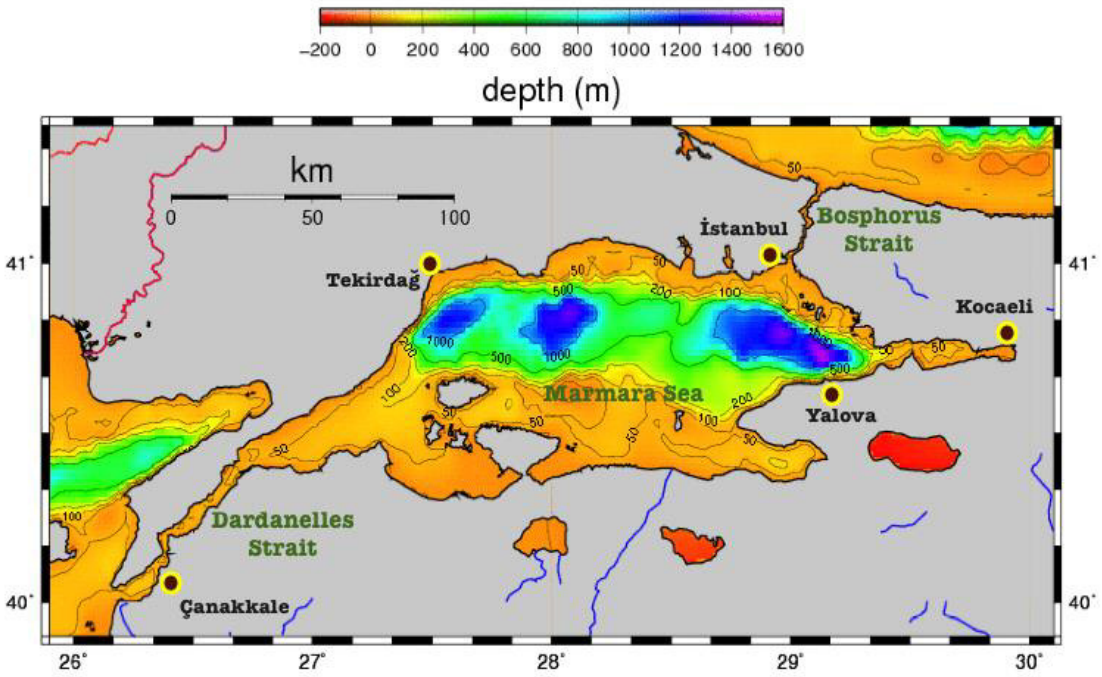

Figure 9. Location and bathymetry of the Turkish Straits System. Adapted from Schroeder et al., 2012.

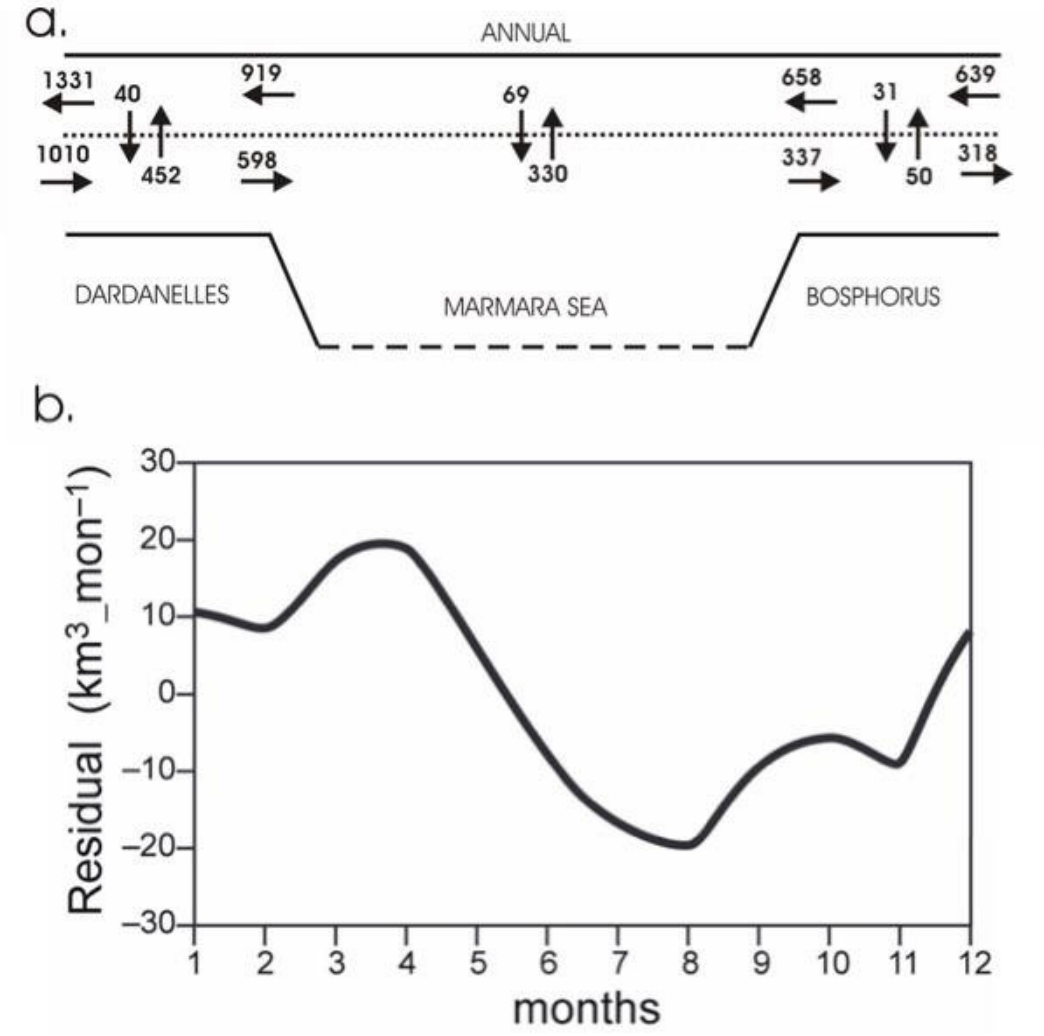

Figure 10: (a) Steady-state annual mean fluxes (km^3/yr) through the TSS and between its individual compartments (after Beşiktepe, 2003); (b) seasonal anomalies of the net transport through the Bosphorus Strait based on data from 1923-1997 (Peneva et al., 2001). Reprinted from *The Climate of the Mediterranean Region - From the past to the future,* Elsevier, 592 p *Lionello P. (ed). 2012*

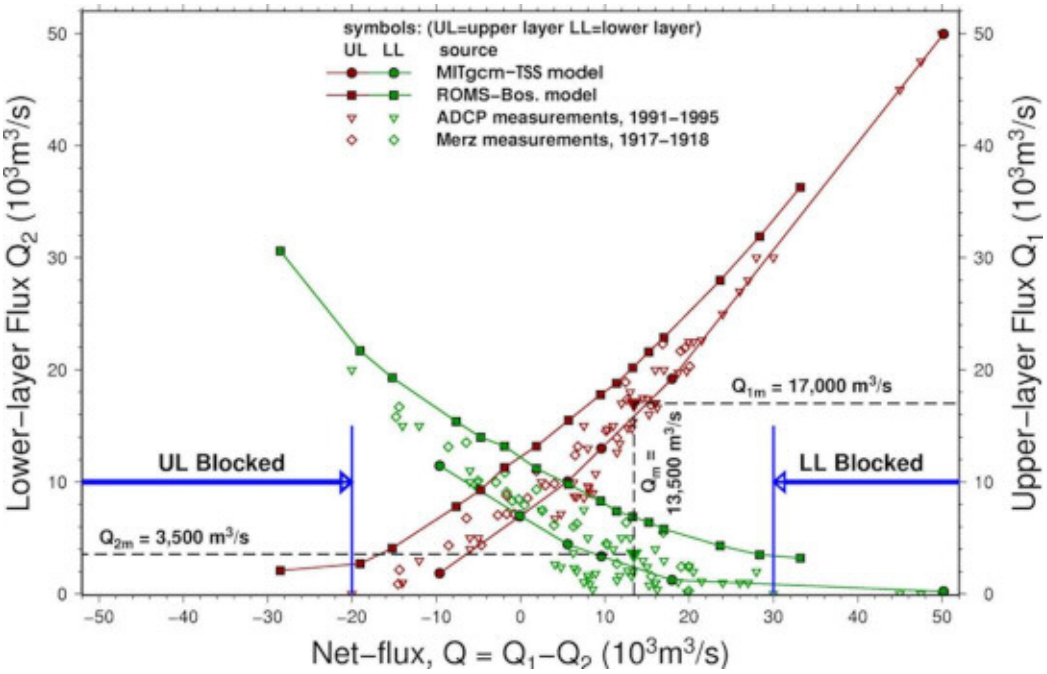

Figure 11. Upper-layer (Q_1) and lower-layer (Q_2) volume fluxes through the Boshorus as a function of the net flux ($Q=Q_1-Q_2$), based on observational data and compared with the results from the Bosphorus model (ROMS) of Sözer (2013) and the TSS model (MITgcm) of Sannino et al. (2015). Data from ADCP measurements on board the R/V BİLİM (Özsoy et al., 1998) and the by Merz and Möller (Möller, 1928) are also shown. The average upper, lower layer and net flux average values obtained from the ADCP measurements are marked (large inverted triangles and dashed lines). The blue arrows show the ranges of net flux where either the upper layer (right) or the lower layer (left) are expected to be blocked.

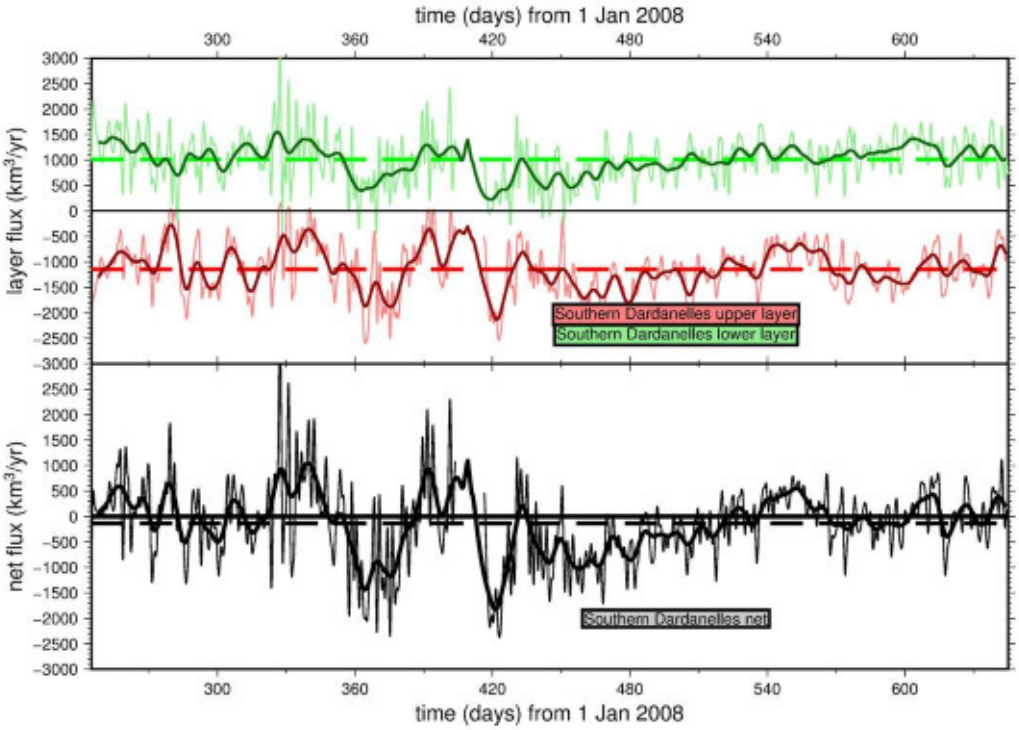

Figure 12. The upper, lower layer and net volume fluxes at the southern end of the Dardanelles Strait, computed from time series of current profiling measurements by Jarosz et al. (2013). The low pass filtered daily time series (black) with a cosine taper filter length of 11-points (~10d) and the mean values of each time series (dashed lines) are also shown.

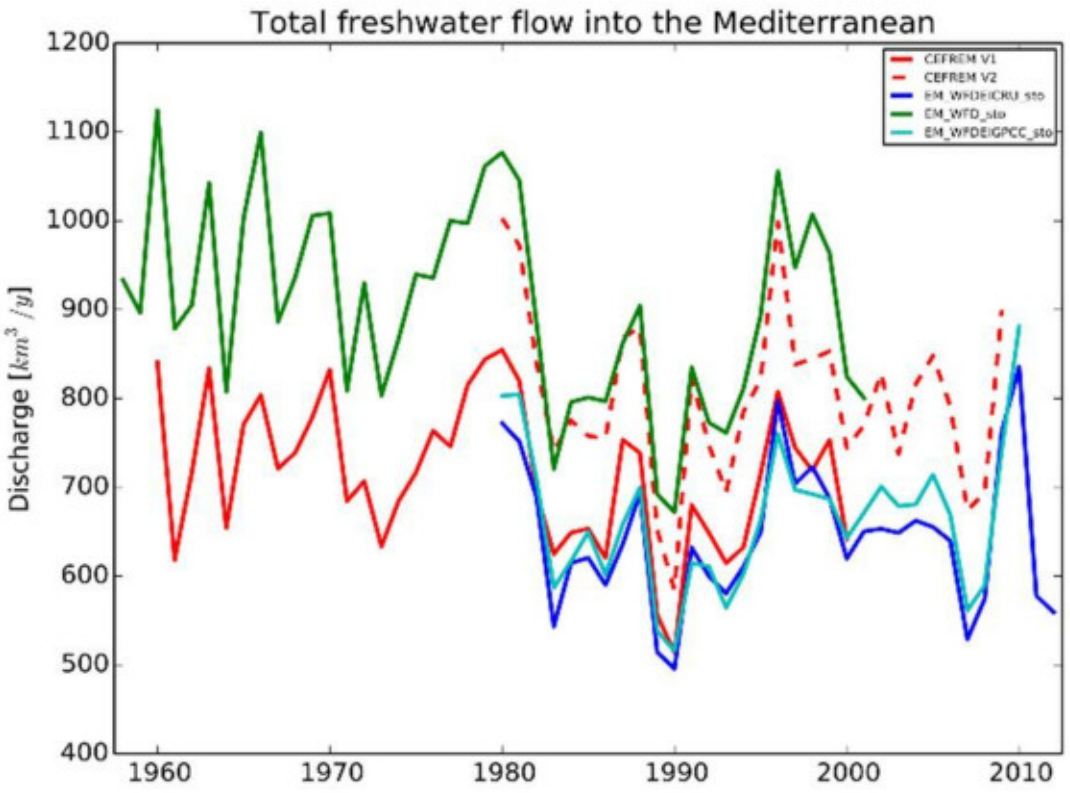

Figure 13 : The evolution of riverine input into the Mediterranean as as provided by CEFREM (Ludwig et al. 2009 for V1 and personal communication for V2) and simulated by the ORCHIDEE model forced with three different atmospheric conditions : 1) WFD (Weedon et al. 2011, labelled EM_WFD_sto), 2) WFDEI, precipitation corrected with CRU (Weedon et al. 2014, labelled EM_WFDEICRU_sto) and WFDEI corrected with GPCC (labelled EM_WFDEIGPCC_sto).

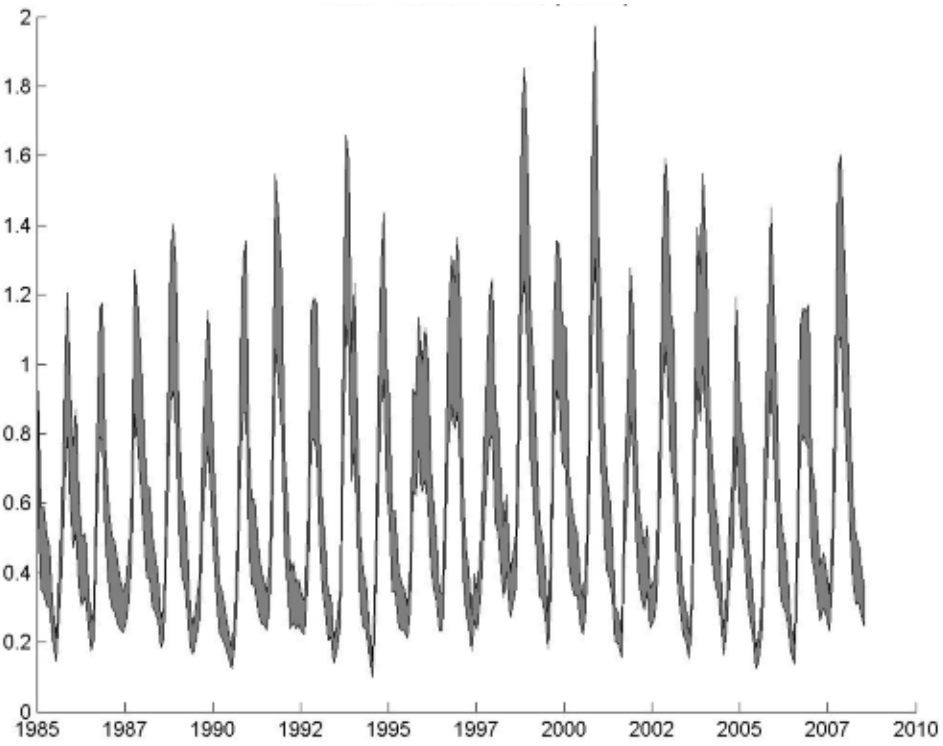

Figure 14. Temporal evolution of the heat flux (in W/m^2) associated to the river runoff. See section 5.5 for details.

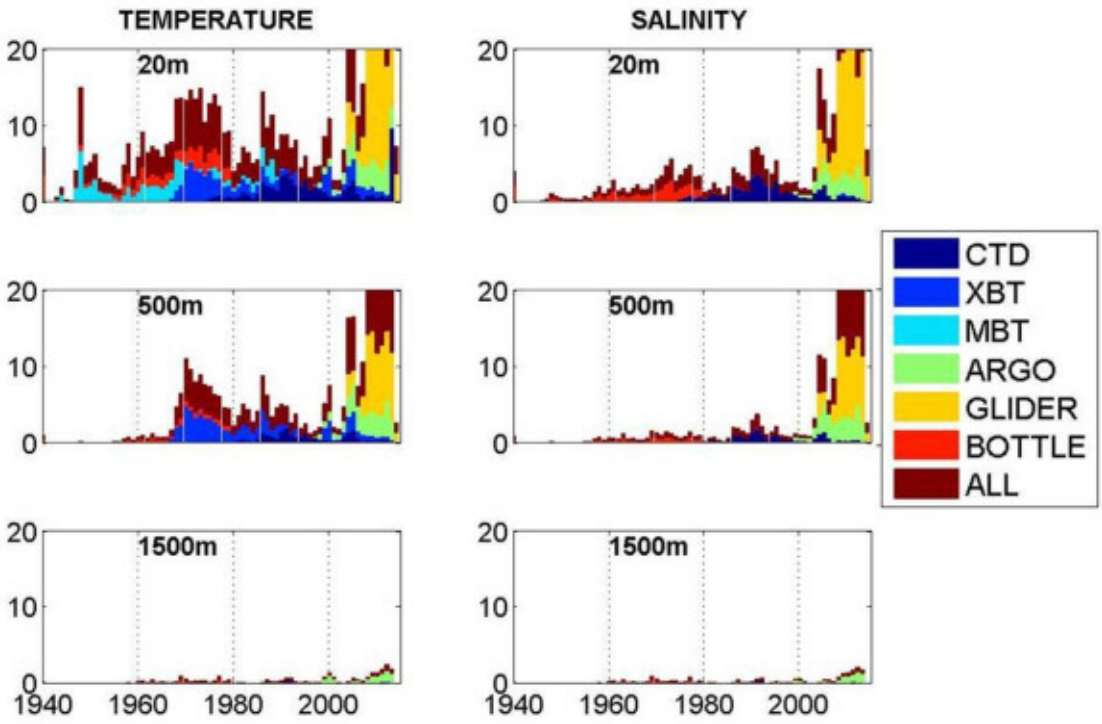

Figure 15. Temporal evolution of the amount of in-situ temperature (left) and salinity (right) data in the Mediterranean Sea from different platforms (vertical axis scaled by 10^3). The results are shown for 20 m (top), 500 m (middle) and 1500 m (bottom).

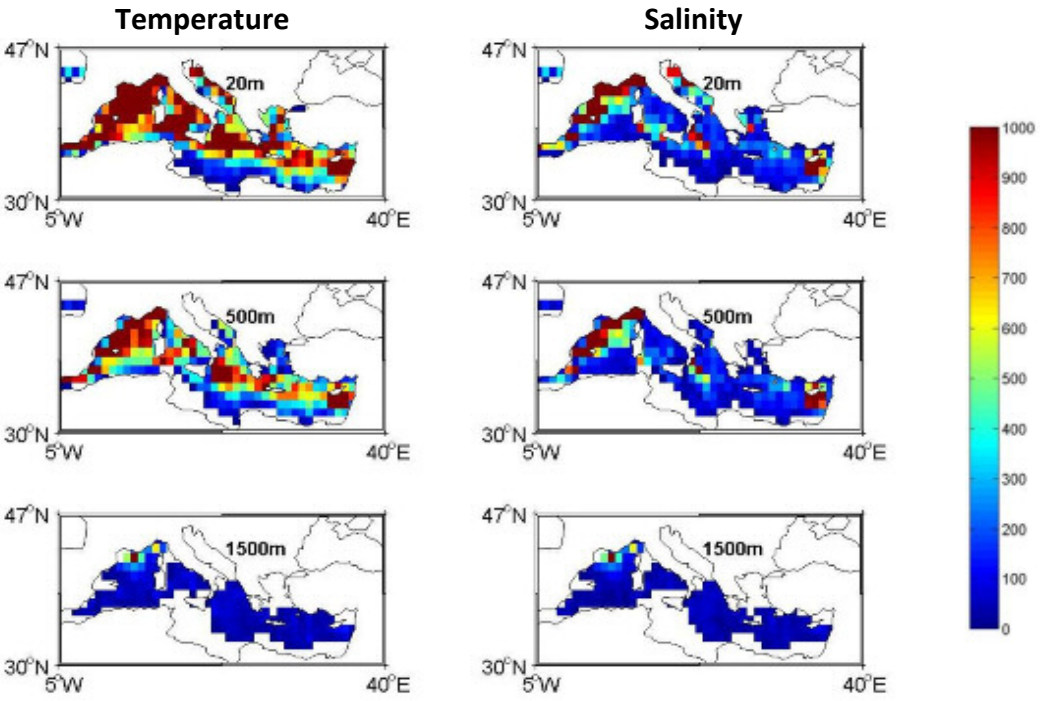

Figure 16. Number of observations of temperature (left column) and salinity (right column) gathered during the period 1940-2014 at different depths. The observations have been binned in 2ºx2º boxes and all kind of observations have been included in the count.

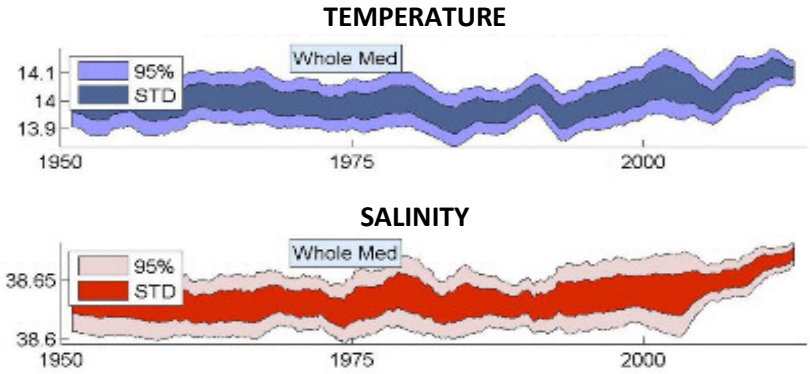

Figure 17. Time evolution of the 3D basin averaged temperature (top) and salinity (bottom) as obtained from Jordà et al. (2016). The dark patch indicates the normal error while the light patch indicates the 95% confidence interval.

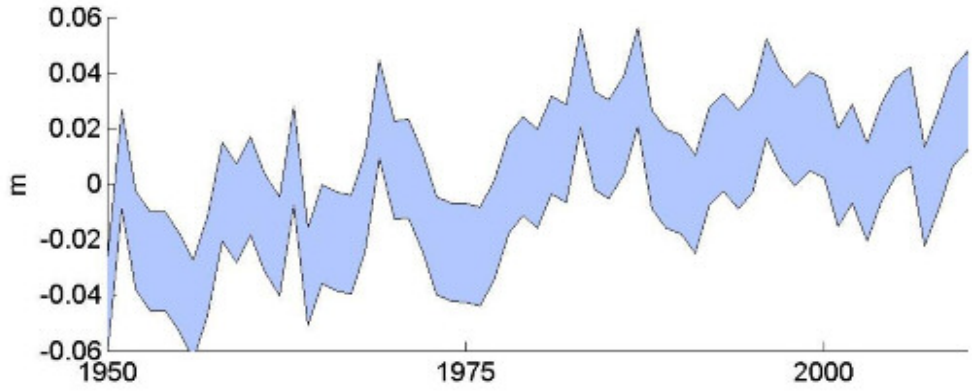

Figure 18. Mediterranean mass water content anomalies (in equivalent water thickness, in m) obtained from the combination of reconstructed basin mean sea level and water expansion from hydrographic datasets. The anomalies are computed with respect to the mean of the time series. The patch indicates the range of uncertainty as described in the text (adapted from Jordà and Gomis, 2013).

Table 1: Comparison of in-situ measurements of daily incoming long-wave and short-wave radiations with satellite estimates, in terms of bias and rms. Ship data collected during the SOFIA (NE Atlantic, Weill et al., 1995), SEMAPHORE (NE Atlantic; Eymard et al., 1996) and CATCH (NW Atlantic, Eymard et al., 1999) experiments; anchored buoy data for the POMME experiment (NE Atlantic; Mémery et al., 2005), while for 2006, several land or buoy pyranometers and pygeometers are operationally used for evaluation. Note the change of statistics between the SEMAPHORE and CATCH experiments in the incoming shortwave radiation due to an improved cloud classification for satellite retrievals. N is the number of days of the series.

Daily data (W/m^2)	**Incoming longwave**		**Incoming shortwave**		**Reference**
	Bias	**RMS**	**Bias**	**RMS**	
SOFIA (1992) ***Ship data***	-2.1 (N=289)	11.3	23.6 (N=149)	75.5	*Eymard et al. (1999)*
SEMAPHORE (1993) ***Ship data***	4.3 (N=640)	15.7	10.6 (N=197)	58.1	*Eymard et al. (1999)*
CATCH (1997) ***Ship data***	0.3 (N=719)	17.0	-1.3 (N=232)	49.9	*Eymard et al. (1999)*
POMME (2001) ***Buoy data***	-1.3 (N=246)	11.1	-1.5 (N=246)	10.6	*Caniaux et al. (2005)*
Year 2006	-1.2	15.6	-1.3	13.7	*LeBorgne et al. (2007)*

Table 2: Errors on the various components of the net heat flux computed with the COARE3.0 bulk algorithm (Fairall et al., 2003) at the LION buoy (period 2011-2013) in the North-western Mediterranean resulting from the observable errors. In parenthesis, the errors when adding the error due to the bulk algorithm. Note that this values have been obtained for a single measurement point and do not represent basin-wide averages.

	Mean values (2011-2013)	**Error**
Air temperature (°C)	16.3	0.2
SST (°C)	18.1	0.1
Relative humidity (%)	0.79	0.03
Wind ($m.s^{-1}$)	7.8	0.3
Latent heat flux (Wm^{-2})	-122	12.2 (19.1)
Sensible heat flux (Wm^{-2})	-14	0.0 (1.7)
Net shortwave radiation (Wm^{-2})	185	3.9
Net longwave radiation (Wm^{-2})	-48	6.9
Net heat flux (Wm^{-2})	+1	14.6 (20.8)
Wind stress (Nm^{-2})	0.15	0.018

Table 3 – Estimations of inflow (Q_A) and outflow (Q_M) and net flow (Q_{net}) through the Strait of Gibraltar published in literature. Gray-shaded rows indicate estimations based on salt ($Q_M S_M + Q_A S_A = 0$) and volume ($Q_A + Q_M = E$) conservation, and, eventually, on the hydraulically controlled flows theory, whereas the remaining values come from observations.

Reference	Q_A (Sv)	Q_M (Sv)	Q_{net}(Sv)
Nielsen [1912]	1.87	-1.78	0.09
Schott [1915]	1.74	-1.64	0.10
Sverdrup et al. [1942]	1.75	-1.68	0.07
Carter [1956]	0.95	-0.91	0.04
Bethoux [1979]	1.68	-1.60	0.08
Lacombe and Richez [1982]	1.21	-1.15	0.06
Bryden and Stommel [1984]	1.67	-1.59	0.08
Bryden and Kinder [1991]	0.92	-0.88	0.04
Bryden et al. [1994]	0.73	-0.68	0.05
García Lafuente et al. [2000]	0.92	-0.87	0.05
Tsimplis and Bryden [2000]	0.78	-0.67	0.11
Baschek et al. [2001]	0.81	-0.76	0.05
Candela [2001]	1.01	-0.97	0.04
García Lafuente et al. [2002]	0.96	-0.84	0.12
Vargas et al. [2006]	0.89	-0.82	0.07
Sánchez Román et al. [2009]	---	-0.78	---
García Lafuente et al. [2009]	---	-0.79	---
Soto Navarro et al. [2010]	0.81	-0.78	0.03
Sammartino et al., [2015]	---	-0.85	---
Mean $\pm$ std from observational estimates	0.86$\pm$0.10	-0.80$\pm$0.08	0.065$\pm$0.033

Table 4.- Time averaged exchange properties at the three notable sections of the SoG, estimated from different runs of the MIT-gcm numerical model applied to the domain showed in the inset of Figure 1 during the periods of hincast I (March 12 to June 24, 2011, upper box) and II (September 1 to November 27, 2011, lower box). The three rows correspond to Espartel Sill (ES), Camarinal Sill (CS) and Tarifa Narrows (TN) sections. Negative sign means transport towards the Atlantic Ocean. Numbers in italic within brackets indicate the associated uncertainty of the time-averaged value, which has been estimated as half the standard deviation of the subinertial series, according to the results by Sammartino *et al* (2015). Column $\Delta S_{effective} = S_M - S_A$ is the difference, expressed in units of the practical salinity scale, between the weighted Atlantic inflow and Mediterranean outflow salinity, which have been computed as and , respectively, overbars representing time-averaged values. Last column is the Outflow Salinity Transport ($Q_M \, \Delta S_{effective}$)

PERIOD	**SECTION**	**Volume Transport,** Q(Sv)			**Salt Transport,** QS(Ton s^{-1}×1000)			**Heat Transport,** QH(TW)			$\Delta S_{effective}$ ("PSU")	***OST*** (Sv×PSU)
		Q_A	Q_M	Q_{net}	QS_A	QS_M	QS_{net}	QH_A	QH_M	QT_{net}		
(I)	**ES**	0.959 *(0.083)*	-0.929 *(0.102)*	0.029 *(0.173)*	35.51 *(3.09)*	-36.34 *(4.01)*	-0.83 *(6.61)*	62.72 *(5.43)*	-51.06 *(5.78)*	11.66 *(10.21)*	1.953 *(0.128)*	-1.814 *(0.080)*
	CS	0.836 *(0.098)*	-0.817 *(0.088)*	0.019 *(0.174)*	31.04 *(3.65)*	-32.29 *(3.48)*	-1.26 *(6.66)*	55.21 *(6.34)*	-43.88 *(4.77)*	11.34 *(10.23)*	2.251 *(0.109)*	-1.839 *(0.109)*
	TN	0.845 *(0.105)*	-0.824 *(0.089)*	0.021 *(0.174)*	31.51 *(3.92)*	-32.61 *(3.51)*	-1.10 *(6.64)*	55.33 *(6.72)*	-44.04 *(4.75)*	11.29 *(10.23)*	2.154 *(0.084)*	-1.775 *(0.123)*
(II)	**ES**	0.918 *(0.096)*	-0.914 *(0.094)*	0.004 *(0.167)*	33.96 *(3.56)*	-35.73 *(3.70)*	-1.78 *(6.37)*	60.76 *(6.78)*	-49.97 *(5.32)*	10.78 *(10.21)*	1.992 *(0.125)*	-1.821 *(0.080)*
	CS	0.800 *(0.106)*	-0.805 *(0.084)*	-0.005 *(0.169)*	29.66 *(3.93)*	-31.84 *(3.33)*	-2.18 *(6.42)*	53.69 *(7.24)*	-43.31 *(4.56)*	10.38 *(10.24)*	2.292 *(0.101)*	-1.846 *(0.112)*
	TN	0.802 *(0.112)*	-0.807 *(0.088)*	-0.005 *(0.169)*	29.84 *(4.15)*	-31.94 *(3.50)*	-2.10 *(6.42)*	53.42 *(7.54)*	-43.23 *(4.73)*	10.18 *(10.18)*	2.205 *(0.087)*	-1.780 *(0.124)*

Table 5.- Contribution of the tidally-driven eddy fluxes to the time averaged exchange properties at the three notable sections of the SoG, estimated from hindcast II (September 1 to November 27, 2011) of the MIT-gcm numerical model applied to the domain showed in the inset of Figure 1. The rows correspond to Espartel Sill (ES), Camarinal Sill (CS) and Tarifa Narrows (TN) sections. Negative sign means transport towards the Atlantic Ocean. Percentage of the eddy fluxes contribution is in italic.

	SECTION	**Volume Transport,** *Q*(Sv)			**Salt Transport,** *QS*(Ton s^{-1}×1000)			**Heat Transport,** *QH*(TW)		
		Net	Eddy	*(%)*	Net	Eddy	*(%)*	Net	Eddy	*(%)*
INFLOW	**ES**	0.918	0.066	*7.2*	33.95	2.41	*7.1*	60.76	1.22	*2.0*
	CS	0.800	0.281	*35.1*	29.66	10.40	*35.1*	53.69	18.66	*34.8*
	TN	0.801	0.014	*1.7*	29.84	0.43	*1.4*	53.42	-0.60	*1.1*
OUTFLOW	**ES**	-0.914	-0.064	*7.0*	-35.73	-2.53	*7.1*	-49.97	-2.93	*5.9*
	CS	-0.805	-0.282	*35.0*	-31.84	-11.24	*35.3*	-43.31	-16.59	*38.3*
	TN	-0.807	-0.014	*1.8*	-31.94	-0.56	*1.8*	-43.24	-0.45	*1.1*

Table 6. Annual mean of water fluxes in the upper and lower layer at the southern limit of the Dardanelles Strait from different authors (adapted from Jarosz et al., 2013). Units are km^3/year and in brackets the equivalence in Mediterranean sea level change (mm/yr). Negative values mean transports towards the Mediterranean.

km^3/year (mm/year)	**Upper layer**	**Lower layer**	**Net**
Ünlüata et al. (1990)	-1258 (-503)	967 (387)	-291 (-116)
Beşiktepe et al. (1994)	-1218 (-487)	918 (367)	-300 (-120)
Özsoy and Ünlüata (1997)	-1181 (-472)	880 (352)	-300 (-120)
Tuğrul et al. (2002)	-1331 (-533)	1010 (404)	-321 (-128)
Kanarska and Maderich (2008)	-1225 (-490)	947 (379)	-278 (-111)
Jarosz et al. (2013)	-1158 (-463)	999 (400)	-158 (-63)
MEAN	-1228 (-491)	954 (382)	-275 (-110)
STD	61 (25)	49 (20)	59 (24)

Table 7 Projections of the change (in %) of basin average value at the end of the XXI century of different components of the Mediterranean water budget from different studies and scenarios.

All values are expressed as %	**Precipitation**	**Evaporation**	**Prec. – Evap.**	**River Runoff**	**Dardanelles**	**Strait of Gibraltar**	**Sum of sources**
Mariotti et al. (2015) Global CMIP5 ; RCP4.5 ; 2071-2098	-8.5	+7	-13.9	-14.8	N/D	N/D	N/D
Adloff et al. (2015) RCM ; A2 /A1B / B1 ; 2070-2099	[-2; -12]	[+8; +12.5]	[-16; -33]	[-4.5;-32]	[-25;-61]	[+31;+60]	[-31;-60]
Sanchez-Gomez et al. (2009) RCM ; A1B ; 2070-2099	[-4;-28]	[-6;+18]	[-15;-35]	[-5;-43]	[+25;-102]	N/D	[-20;-60]
Mariotti et al. (2008) CMIP3 ; A1B ;2070-2099	[+2;-23]	[+6;+9]	[-11;-25]	[-56;-87]	[-63;-85]	N/D	[-34;-42]

Table 8 Same that Table 7 for the heat fluxes through the sea surface and the Strait of Gibraltar.

All values are expressed as %	**Surface heat flux**	**Gibraltar heat flux**
Adloff et al (2015) RCM ; A2 /A1B / B1 ; 2070-2099	[+25;+75]	[0;+40]
Dubois et al (2012) / Gualdi et al. (2013) RCM ; A1B ; 2021-2050	[+60;+118]	[-11;+3]
Somot et al. (2008) RCM ; A2 ; 2070-2099	[+69]	N/D
Somot et al. (2006) RCM ; A2 ; 2070-2099	[+71]	N/D

Table 9 Projected changes of the basin average temperature, salinity and density at the end of the XXI century with respect to the values at the end of the XX century from different studies.

	T3D (°C)	**S3D (psu)**
Adloff et al. (2015) RCM ; A2 /A1B / B1 ; 2070-2099 vs. 1961-1990	[+0.93;+1.35]	[+0.28;+0.52]
Macías et al. (2015) RCM ; RCP4.5/RCP8.5 ; 2096-2099 vs. 2014-2017	[+1.30 ;+2.50]	[+0.6 ;+0.9]
Carillo et al. (2012) RCM ; A1B ; 2001-2050 vs. 1951-2000	[+0.38;+0.42]	[-0.08;+0.03]
Somot et al. (2006) RCM ; A2 ; 2099 vs CTRL	[+1.5]	[+0.23]

Table 10: Best estimates of the different components of the Mediterranean Sea heat budget (see eq. 1) with their associated uncertainties for the long term average and for the period 2005-2010. In brackets we include the indirect estimate in that case where it provides lower uncertainty than the direct estimate. The indirect estimate is obtained assuming that the budget is closed and using the information of the other components and its uncertainty has been obtained considering that the errors in the different components are uncorrelated. See the text for further details.

HEAT FLUX (W/m^2)	***1980-2010***	***2005-2010***
SURFACE FLUXES	-3.0 $\pm$ 8.0 [-4.9 $\pm$ 4.3]	-3.0 $\pm$ 8.0 [-1.7 $\pm$ 5.1]
GIBRALTAR	4.5 $\pm$ 4.1	4.5 $\pm$ 4.1
DARDANELLES	0.1 $\pm$ 0.2	0.1 $\pm$ 0.2
RIVERS	0.6 $\pm$ 0.4	0.6 $\pm$ 0.4
HEAT CONTENT CHANGE	0.3 $\pm$ 0.3	3.5 $\pm$ 3.0

Table 11: As Table 10 but for the water mass budget.

WATER FLUX (m/yr)	***1980-2010***	***2005-2010***
SURFACE FLUXES	-0.8 $\pm$ 0.2	-0.9 $\pm$ 0.2
GIBRALTAR	0.8 $\pm$ 0.4 [0.5 $\pm$ 0.2]	0.85 $\pm$ 0.4 {0.6 $\pm$ 0.2]
DARDANELLES	0.1 $\pm$ 0.02	0.1 $\pm$ 0.02
RIVERS	0.2 $\pm$ 0.01	0.2 $\pm$ 0.01
WATER CONTENT CHANGE	$5 \cdot 10^{-4} \pm 10^{-3}$	$1 \cdot 10^{-3} \pm 3 \cdot 10^{-3}$

Table 12: As Table 10 but for the salt mass budget.

SALT FLUX (10^6 kg/s)	***1980-2010***	***2005-2010***
GIBRALTAR	-1.5 $\pm$ 6.5 [-0.1 $\pm$ 0.2]	-1.5 $\pm$ 6.5 [0.1 $\pm$ 0.3]
DARDANELLES	0.2 $\pm$ 0.2	0.2 $\pm$ 0.2
SALT CONTENT CHANGE	0.1 $\pm$ 0.1	0.3 $\pm$ 0.2