THE BOSPHORUS STRAIT: EXCHANGE FLUXES, CURRENTS AND SEA-LEVEL CHANGES

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Abstract.

The flows through the Turkish Straits System (comprised of the Straits of Bosphorus and Dardanelles, and the Sea of Marmara) are subject to a great degree of transient variability, depending on the atmospheric factors and the water budget. As a result of the rapid along-strait variations in the strait geometry, the sharp stratification, and the temporary blocking of the flows in either direction, the maximal exchange regime of controlled flows in the Bosphorus exhibit complex non-linear response to forcing (by the net water budget, pressure and wind setup effects in adjacent basins), resulting in the observed time dependence from daily to interannual time scales. One consequence is the observed controls on sea level changes. Complex relationships exist between the exchange flows, sea level variations, net water budgets and atmospheric pressure variations in the adjacent basins, and are not so easily understood within the full range of time scales. We make a first attempt to discuss these relationships in some detail.

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1. Introduction

The exchange across the Turkish Straits System, with two-layer structure, has direct influence on the environmental problems in the region. The transport of materials between the Black Sea and the Mediterranean, (Polat and Tuğrul, 1995; Özsoy *et al.*, 1995a), and the fate of pollutant discharges, such as wastewater from the city of İstanbul (DAMOC 1971; Gunnerson and Özturgut, 1974; Gunnerson, 1974; Özsoy *et al.*, 1994, 1995b), largely depend on the exchange characteristics. In addition to having a direct impact on local environmental problems, the exchange and mixing in the Turkish Straits System determine the water mass structure and renewal characteristics in the adjacent Black Sea, as well as in the Marmara Sea itself.

In return, the hydrological regimes of adjacent basins establish the long term fluxes across the System and determine the properties of waters in transit. In the Black Sea, the excess of precipitation ($P \simeq 300km^3/yr$) and runoff ($R \simeq 350km^3/yr$) versus evaporation ($E \simeq 350km^3/yr$) is balanced by a net outflow ($Q_b \simeq 300km^3/yr$) through the Bosphorus (Ünlüata *et al.*, 1990). On the other hand, the freshwater runoff, mainly from large rivers in the northwest (Danube, Dnepr and Dnestr) has strong seasonal and interannual natural variability (Sur *et al.*, 1994), influencing the fluxes.

Average fluxes of $Q_1 \simeq 600km^3/yr$ in the upper layer, and $Q_2 \simeq 300km^3/yr$ in the lower layer of the Bosphorus are calculated from mass balance, based on the best estimates of surface fluxes and average salinity in the Turkish Straits (Ünlüata *et al.*, 1990). The long-term average mass budget requires $Q_1/Q_2 = S_2/S_1 \simeq 2$, where Q_1, S_1 and Q_2, S_2 are the fluxes and salinities respectively in the upper and lower layers. The instantaneous fluxes, revealed by ADCP measurements greatly differ from these estimates, as they follow closely the transient meteorological and hydrological forcing in adjacent basins (Latif *et al.*, 1991; Özsoy *et al.*, 1994, 1995a, 1995b, 1996).

It is not clear what drives the exchange through the Turkish Straits. In the long-term, mass balance requires a ratio of ~1/2 between the salinities of the outflowing waters of the Black Sea and the incoming waters of the Mediterranean Sea at the Bosphorus (Ünlüata *et al.*, 1990). On an interannual basis, significant correlations seem to exist between the Danube river water influx and sea level, suggesting efficient control by the Bosphorus. Similarly, meteorological factors, and especially barometric pressure forcing also appear to be important. The roles of these different drivers are not clear, since the nonlinear, controlled exchange flows follow oscillatory but unpredictable seasonal and interannual patterns (Sur *et al.*, 1994; Özsoy *et al.*, 1995a, 1996).

A description of the time dependent exchanges across the strait is given, following Özsoy *et al.* (1995b and 1996), based on experimental results and expected response characteristics. Dependence on time scales is discussed.

2. Bosphorus Strait Hydraulics

Two pertinent features of the Bosphorus Strait geometry, shown in Figure 1, are of great importance for the dynamics of the strait flow: The first is the northern sill, located $\sim 4km$ northeast of the Black Sea entrance, at a depth of $60m$. The sill connects the Bosphorus channel into a bottom feature extending it on towards the Black Sea continental shelf, first in the form of a curved channel, followed by a shallow delta structure on the bottom. The lower layer flow of Mediterranean water is conveyed through these bottom structures after overflowing the sill (Latif *et al.*, 1991). The second important feature is the contraction in the southern Bosphorus, where the width decreases to a minimum of $\sim 600m$, coincident with the maximum depth of $\sim 110m$. Here, the flow in both layers speed up, and surface currents can reach a maximum of up to 2 m/s in the narrow section. The upper layer flow then undergoes dissipation in the southern Bosphorus after the contraction, the shallow topography in this region creates three dimensional circulations, and the surface flow is partly reformed before reaching the southern exit, where it forms a jet issuing into the Marmara Sea. The flow regime of the Bosphorus, based on extensive observations, and confirmed by numerical modelling, has been schematized in Figure 2 (Özsoy *et al.*, 1986; Oğuz *et al.*, 1990). Other complexities of the Bosphorus flows include secondary and eddy circulations induced by the tortuous geometry of the Strait, unsteady effects connected with wind set-up and changes in adjacent basins, along strait density variations resulting from entrainment of fluid from one layer into the other. These effects are important (*e.g.*, Pratt, 1987; Helfrich, 1995), but unconsequential in establishing the 'maximal exchange' regime (Armi and Farmer 1987).

The special setting of the Bosphorus (Figures 1 and 2), with two hydraulic controls, respectively imposed at the sill located offshore of its northern entrance, and at the contraction in the southern part, makes it one of the best examples of the 'maximal exchange' regime (Armi and Farmer, 1986; Farmer and Armi, 1986; Armi and Farmer, 1987). A contraction located between the higher density Marmara Sea and the northern sill, and suitable basin conditions, as considered by Farmer and Armi (1986), allow 'maximal exchange'. The strait displays assymmetrical behaviour with respect to barotropic flow, as a function of the contraction/sill width ratios.

In contrast, 'submaximal exchange' occurs in the Dardanelles Strait (Özsoy *et al.*, 1986, 1988; Oğuz and Sur, 1989; Ünlüata *et al.*, 1990; Oğuz *et al.*, 1990), where a single hydraulic control occurs at the 'narrows', a

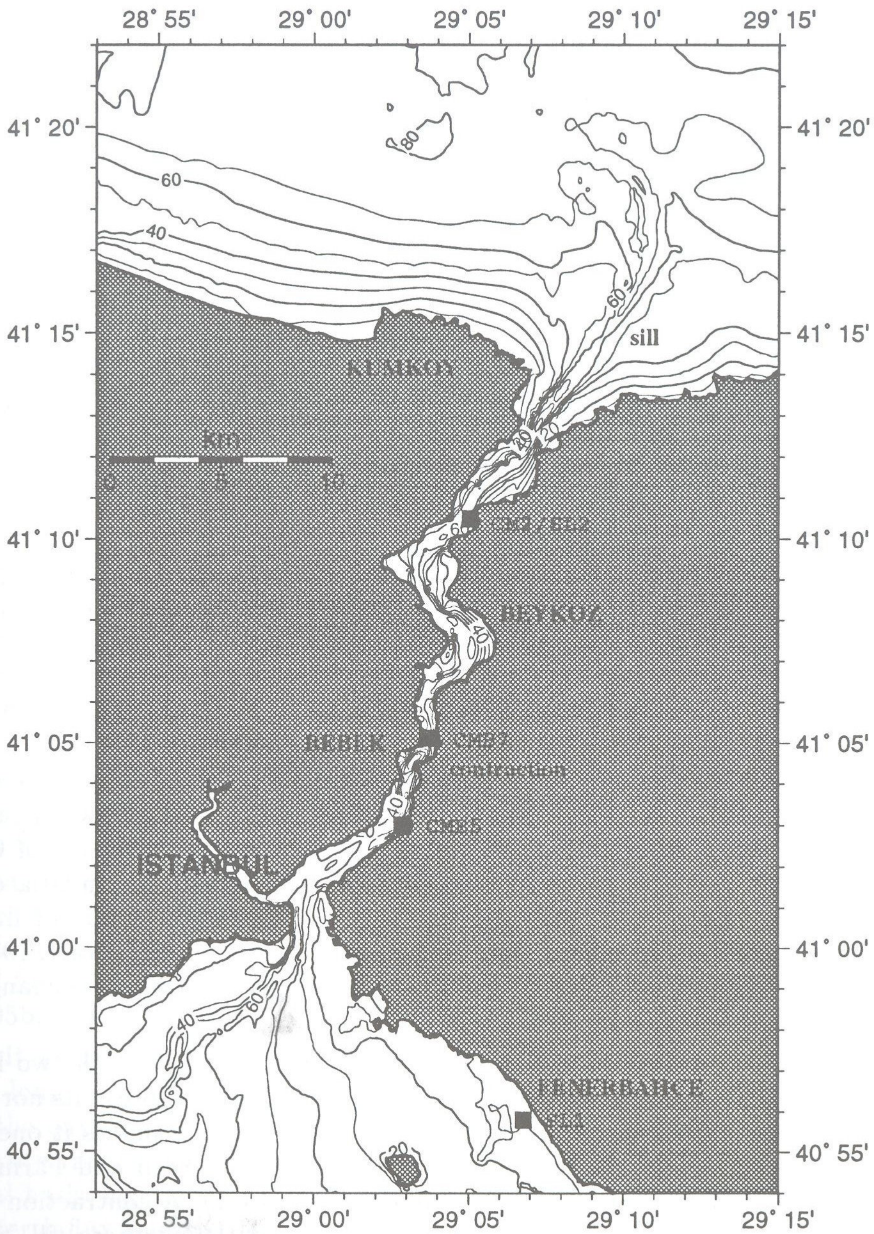

Figure 1. Location map for Bosphorus. CM2 (Anadolukavağı), CMB5 (Station B5), CMB7 (station B7): current meter mooring sites, SL1 (Fenerbahçe) and SL2 (Anadolukavağı) sea level measurement sites.

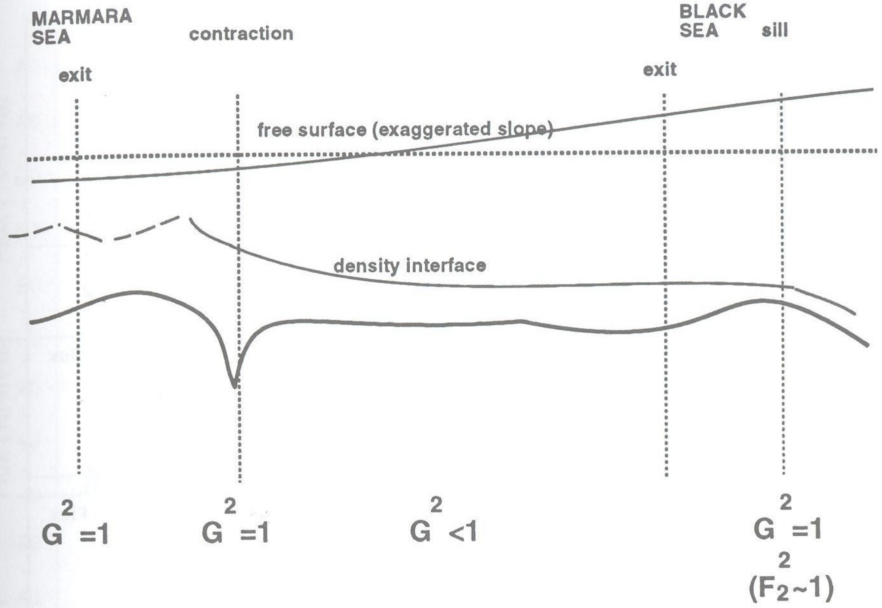

Figure 2. Schematization of the Bosphorus two-layer system.

contraction in the form of a strong bend, in the middle part of the Strait.

The most distinctive characteristic of Bosphorus flows is their transience. Under normal conditions in the Bosphorus, observations indicate two-layer flows with transitions at the control sections. On the other hand, the flow system responds rapidly to changes in driving forces: transient changes in the water budget of the Black Sea or setup by northerly winds, can temporarily cause the lower layer flow to be blocked. Similarly, southerly winds in winter often cause the upper layer to be arrested, and Marmara waters pushed into the Bosphorus, resulting in three-layer stratification (Özsoy *et al.*, 1995a, 1996). The upper-layer blocking is often accompanied by diminishing sea level difference between the two ends of the strait (DAMOC, 1971; Gunnerson and Özturgut, 1974; Arısoy and Akyarlı, 1990), and increased surface salinity is observed in the southern Bosphorus (the so-called *Orkoz* events, Artüz and Uğuz, 1976).

3. Current and Sea Level Measurements in the Bosphorus

ADCP measurements using a vessel-mounted (RD Instruments) unit with 150KHz frequency were obtained in the Bosphorus during the period of 1991-1995 in the course of studies of mixing and dispersion of waste wa-

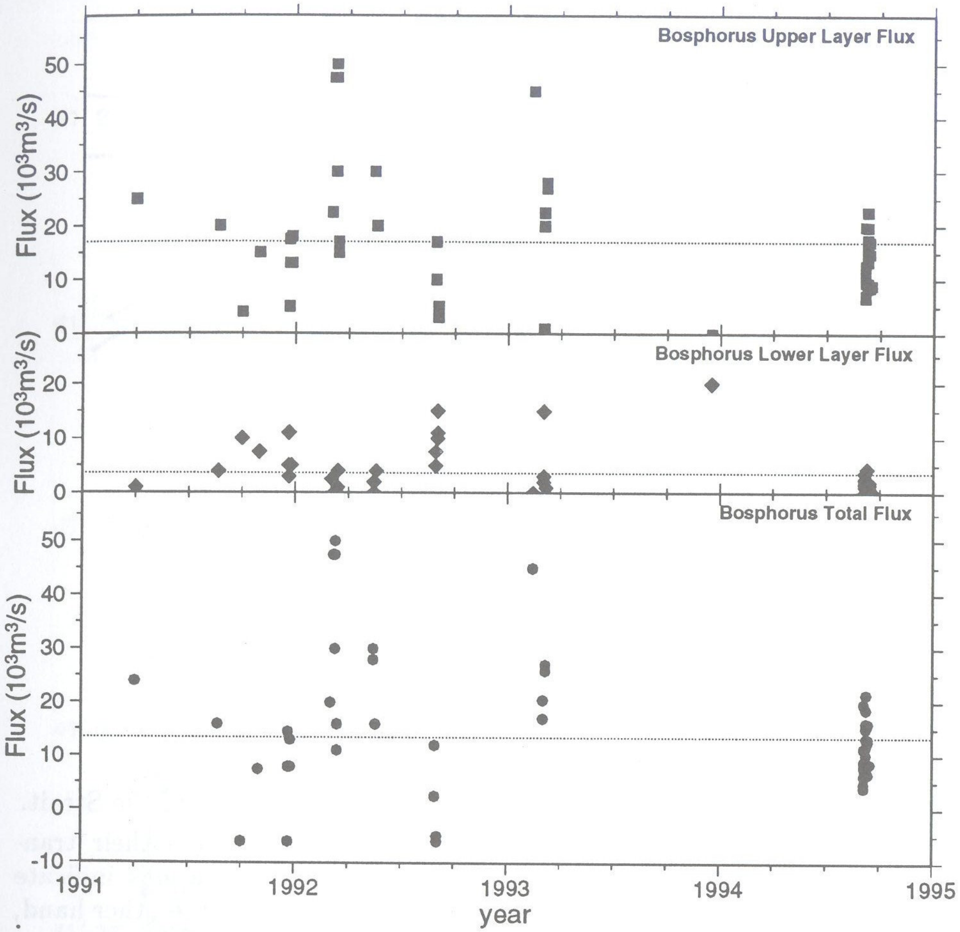

Figure 3. ADCP measurements of upper, lower layer and total volume fluxes in the Bosphorus, during 1991-1994. Dotted lines represent mean values.

ters from the city of İstanbul (Özsoy *et al.*, 1994, 1996). ADCP velocity profiles were obtained by the R/V BİLİM. In addition to data obtained at hydrographic stations, continuous ADCP velocity profiling was made across some sections the Bosphorus at averaging intervals of 30 - 60s along the ship track. CTD measurements obtained with a Seabird 911 system and a JVC echosounder were used to identify the density interface, and hence to compute the discharges in the upper and lower layers.

A summary of ADCP based flux measurements in the Bosphorus is given in Figure 3, and on a seasonal basis in Figure 4. The fluxes estimated from the mean value of these measurements are $Q_1 \simeq 540 km^3/yr = 17,000 m^3/s$ in the upper layer, and $Q_2 \simeq 115 km^3/yr = 3,600 m^3/s$ in the lower layer.

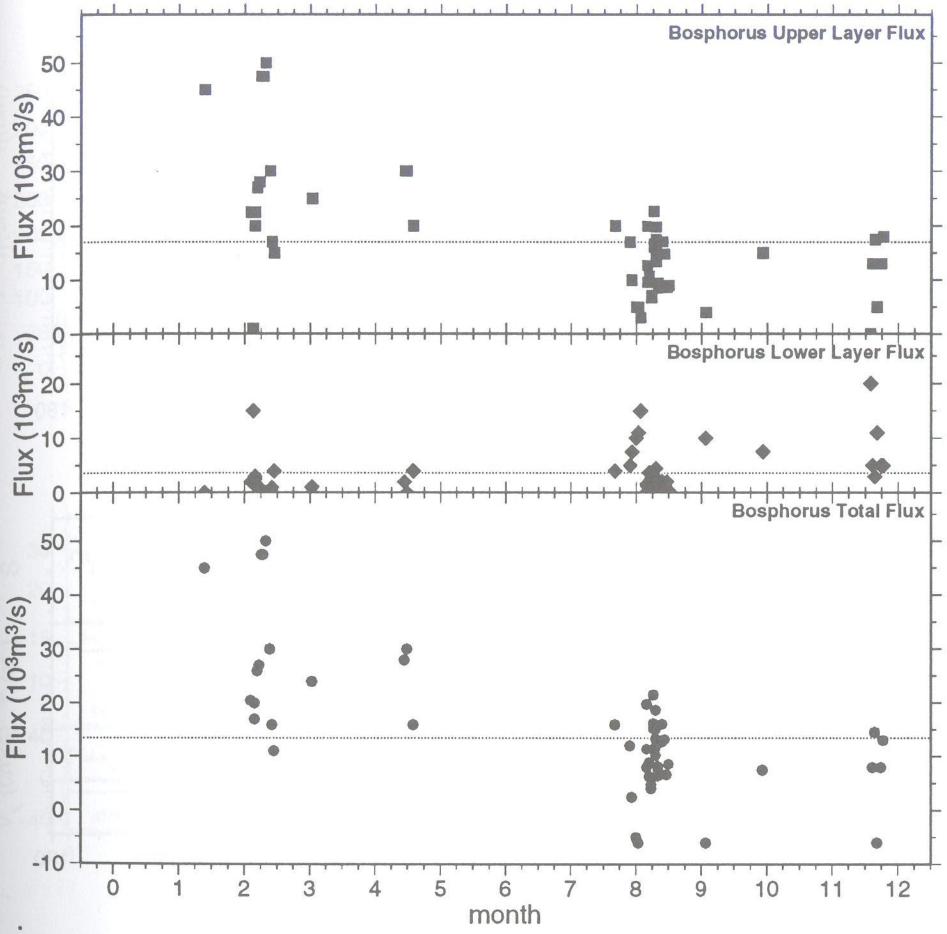

Figure 4. ADCP measurements of upper, lower layer and total volume fluxes in the Bosphorus, plotted on a seasonal basis. Dotted lines represent mean values.

The measured mean upper layer flux is close to the calculated value (see above), but the mean lower layer flux appears to be consistently underestimated, compared to the mass budget estimate. This is most probably a result of the loss of resolution and values of ADCP velocity measurements near the bottom. However, we must also note that many values representing blocked or nearly blocked conditions are included in the average. The available current-meter measurements show that in reality such blocking events are mostly short-term, but we can not appropriately represent their true, real-time intermittency in taking a statistical average of the randomly spaced ADCP measurements.

Because of the great transient variability in fluxes, a clear definition of

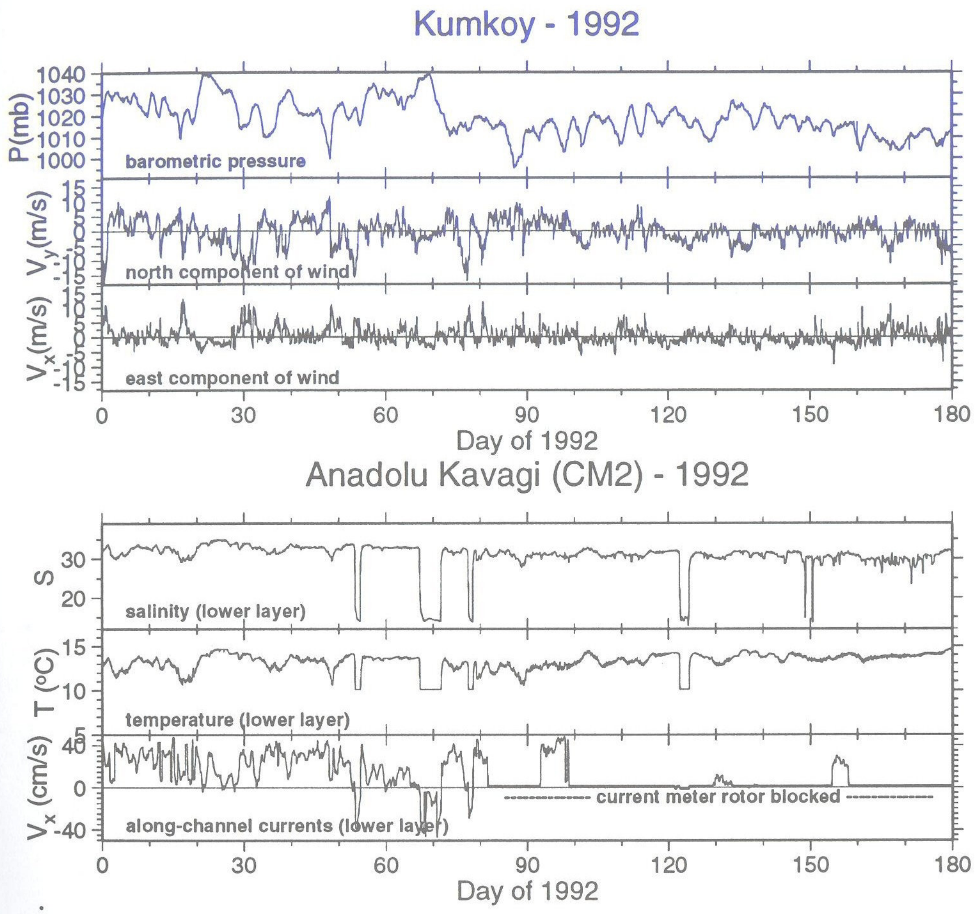

Figure 5. Current meter measurements of the lower layer flow characteristics at Anadolukavağı(CM2), and atmospheric variables at Kumköy during January-March, 1992.

the seasonal signal is not possible. Based on Figure 4, we can only suggest an increased net flow in the late winter and spring and a decrease in the autumn and early winter months, without displaying a statistically significant and well-defined seasonal pattern. Indirect measurements, based on hydrography, suggest an increasing influence of low salinity waters in the Marmara Sea surface layer during the spring and summer months, with a delay appropriate for the residence time of the Marmara basin (Beşiktepe *et al.*, 1994). Therefore we can only make a general statement on seasonal variability based on a combination of different types of measurements.

In addition to the ADCP measurements, a program for long-term mea-

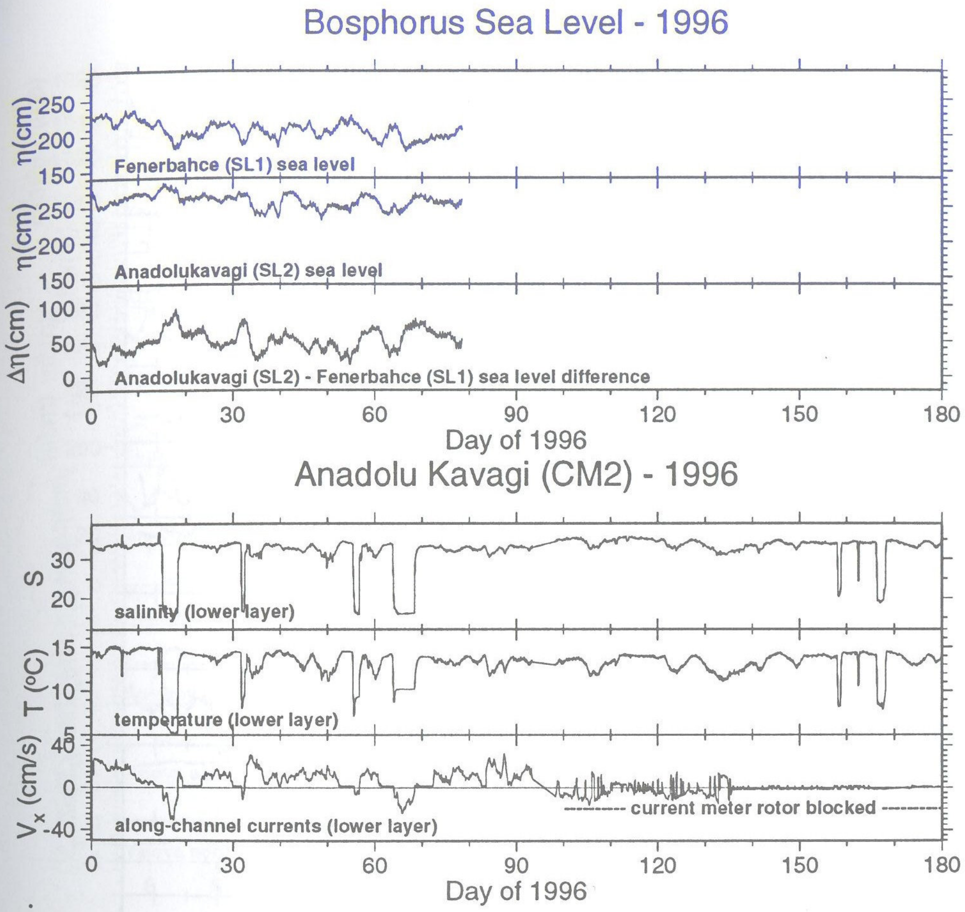

Figure 6. Current meter measurements during September 1994 at stations CMB7 and CMB5 in the Bosphorus, including current, temperature and salinity measurements at two depths. Superposed on top panel is the sea level difference between SL2 and SL1.

surements of currents and sea level in the Bosphorus was carried out in recent years. This program was purely based on opportunities and was not necessarily a part of planned project activities; thus the measurements are not always continuous and concurrent as one would wish them to be. Furthermore, logistical constraints limited these measurements: For example, continuous measurements with current-meters deployed in the Strait are extremely difficult under fast currents and heavy ship traffic, unless special precautions are taken and expensive equipment such as large buoys and acoustic releases are used, such as in the case of September 1994, when we carried out high resolution microstructure, hydrographic, and acoustic

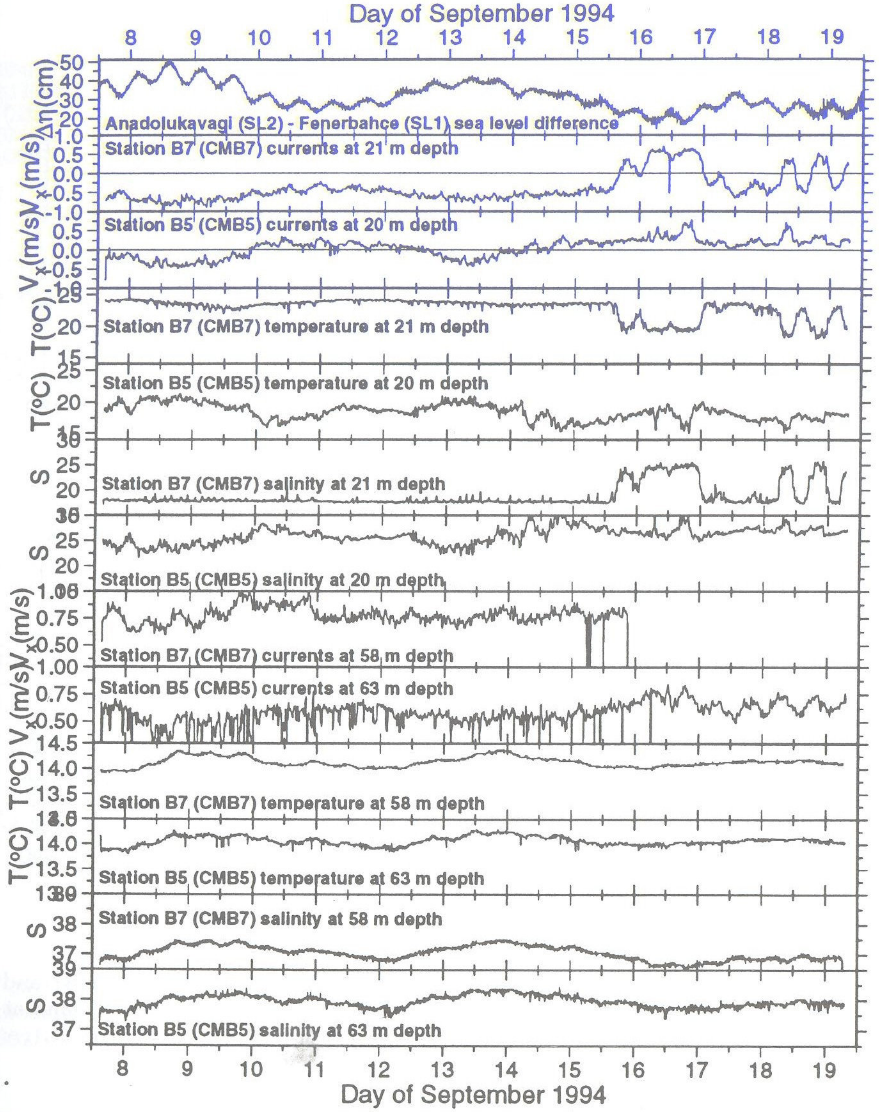

Figure 7. Current meter measurements of the lower layer flow characteristics at Anadolukavağı(CMB7), and Bosphorus sea level measurements at Anadolukavağı(SL2) and Fenerbahçe (Sl1) during January-March, 1992.

measurements in the Bosphorus (Gregg, 1995). The long-term measurement strategy was therefore based on deploying Aanderaa current-meters only in the lower layer flow in the northern reaches of the Strait near Anadolu Kavağı (location CM1, Fig. 1), where the instruments were put in an iron cage placed on the bottom (depth $\sim 60-70m$) and secured with a line to the coast. The sea level measurements were obtained at locations SL1 (Anadolu

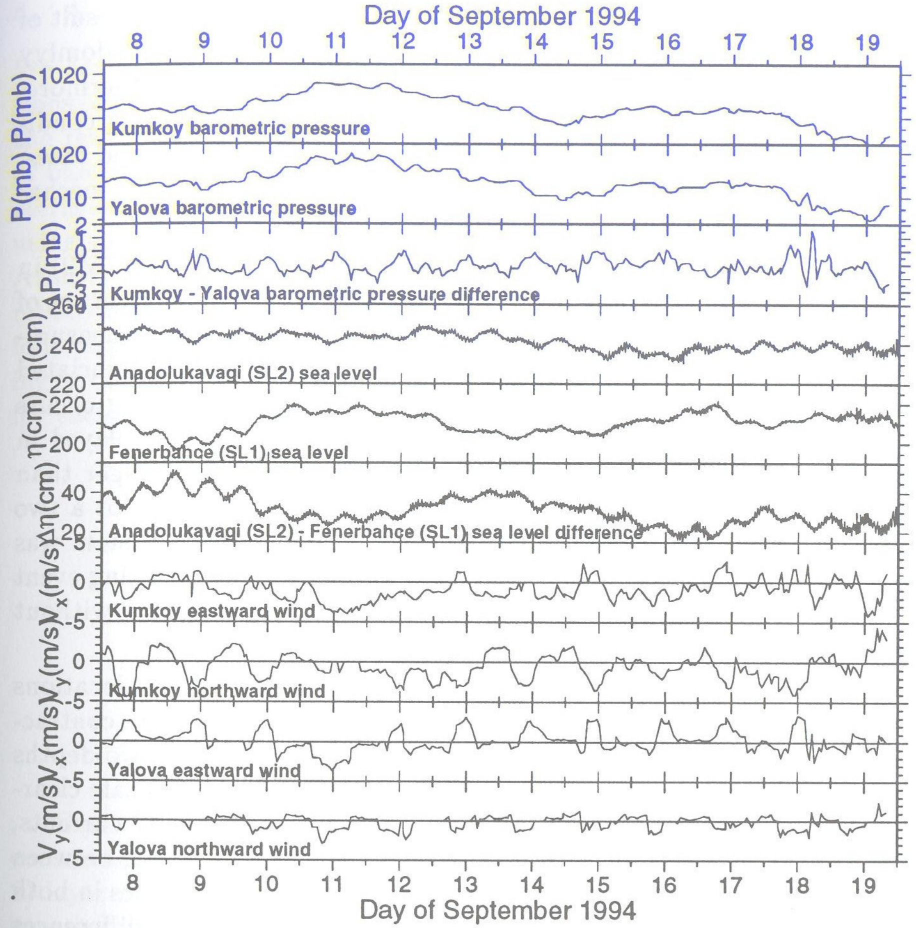

Figure 8. Meteorological parameters at Kumköy (Black Sea) and Yalova (Marmara Sea) sea level at SL2 and SL1, corresponding to the current meter measurements in September 1994.

Kavağı) and SL2 (Fenerbahçe), using Aanderaa sea level recorders.

The current-meter measurements Anadolukavağı (CM2) and meteorological parameters at Kumköy during the first half of 1992 are shown in Figure 5. The along-channel rotated currents were mainly towards the Black Sea at 40 cm/s, but were stopped by blocking events during January-March, and even reversed their direction on several occasions in February and March. A good indicator of blocking, especially in the stronger cases where the lower layer currents reversed, is the dropouts in temperature and salinity time-series, caused by the colder and less saline surface waters reaching the bottom, providing information on blocking after late March, when the current-meter rotor was fouled by debris. An interesting result of these observations is that the blocking conditions occur almost randomlyy in the first six months of the year, without much seasonality. Furthermore, comparison with meteorological parameters show that the blocking conditions can be correlated with increased barometric pressure in some cases, and with increased northerly winds in others, suggesting that the mechanism can not be attributed to a single type of forcing.

In Figure 6, the current-meter measurements Anadolukavağı (CM2) in the first half of 1996 are compared with the only available long-term set of sea level measurements (SL2 and SL1), concurrent with current measurements in the Bosphorus. We observe that the blocking cases are associated with increased sea level differences of about $> 50cm$. The blocking occurs in January-March as well as in June. In all individual events displayed in Figures 5 and 6, blocking of the lower layer does not persist longer than 5 days. However, note that repeated cases of blocking occurred for a two week period in January 1992 (Figure 5) when a high pressure system was probably established in the Black Sea region, but each single blocking event only stopped the flow at the sill, without reversing the currents and without pushing the interface south of Anadolukavağı.

Short-term current-meter measurements were obtained at two locations (CMB5 and CMB7), respectively located south and north of the contraction in the southern Bosphorus, with current-meters placed at two depths below and above the density interface (the upper one placed at a safe clearing distance from ship keels). Figure 7 shows CMB5 and CMB7 currents, temperature and salinity records, together with sea level difference between SL1 and SL2. Harmonious changes in currents and water properties in both layelayers are observed, which generally conform to the pressure differences between the two ends of the Strait. The reversals in the surface currents after September 15 result from the interface becoming shallower than the current-meters, when the sea level difference decreases. In the lower layer, better agreement between currents and sea level occurs at station CMB5, while the same is not true for CMB7, located north of a small uncharted bottom topographic feature (Gregg, 1995). A phase difference of less than one day is noticable between the lower layer currents and sea level difference.

Barometric pressure and difference at Kumköy (Black Sea) and Yalova (southern Marmara Sea), sea level and difference at SL1 and SL2, and wind components at the same meteorology stations are shown in Figure 8. In general, it can be verified that the response of the sea level to changes in barometric pressure is immediate on the Black Sea side, which has a smaller influence on the Marmara side, while the weak summer winds in this case have no appreciable effects.

4. A Two Layer Model of the Bosphorus

We develop simple models of the exchange flows by integrating the equations of motion between critical control sections, to be used for investigating the response of the Turkish Straits to seasonal forcing, allowing for cases of blockage, net through-flow, sea level and pressure changes, and parameterising the effects of internal and boundary friction. Similar approaches, using integral conditions can be found in Welander (1974), Farmer and Armi (1986), and in Sümer and Bakioğlu (1981), Maderich and Efroimson (1986) in the case of Bosphorus. We present the simple hydraulic model following Özsoy *et al.* (1996), starting with the equations of motion governing a two-layer, unidirectional flow, integrated across the channel widths for each layer:

$$b_1\frac{\partial y_1}{\partial t} - \frac{\partial u_1 b_1 y_1}{\partial x} = 0 \tag{1}$$

$$b_2\frac{\partial y_2}{\partial t} + \frac{\partial u_2 b_2 y_2}{\partial x} = 0 \tag{2}$$

$$-\frac{\partial u_1}{\partial t} + u_1\frac{\partial u_1}{\partial x} = -\frac{1}{\rho_1}\frac{\partial P}{\partial x} - g\frac{\partial(y_1+y_2+h)}{\partial x} + \frac{\tau_w - \tau_i}{\rho_1 y_1} \tag{3}$$

$$\frac{\partial u_2}{\partial t} + u_2\frac{\partial u_2}{\partial x} = -\frac{1}{\rho_2}\frac{\partial P}{\partial x} - g\frac{\partial(ry_1+y_2+h)}{\partial x} + \frac{\tau_i - \tau_b}{\rho_2 y_2}. \tag{4}$$

Equations (1,2) and (3,4) respectively are the mass and momentum conservation statements for the upper (subscript 1) and lower (subscript 2) layer unidirectional flows within the strait, where u_j, b_j, h_j, and ρ_j respectively are the velocity, width, depth and density of the flow within each layer ($j = 1,2$). The barometric pressure P is impressed on the surface, and τ_w is the component of wind stress along the strait axis, while τ_i and τ_b are the frictional stresses respectively acting at the two-layer interface and the bottom. The definition $r \equiv \rho_1/\rho_2$ is later replaced with $\varepsilon \equiv \Delta\rho/\rho_2 = 1 - r$, and h is the elevation of the bottom, measured from a horizontal datum. Although it is possible to include effects of the channel cross-sectional geometry (*e.g.* Bryden and Kinder, 1991; Dalziel 1992), for simplicity, we will take the widths b_1 and b_2 to be representative of the average channel cross-section in each layer, in much the same way as we take u_1 and u_2 to be the average velocities in each layer.

In constructing a simple model of the Bosphorus Strait exchange, described in Özsoy *et al.* (1996), we make several simplifying assumptions: Firstly, we allow storage in the Black Sea, with uniform sea level in the basin, supplementing a continuity equation for the Black Sea. Next, we consider a quasi-stationary state for the Strait, including seasonal changes in Black Sea mean sea level, but filtering out all high frequency inertial and surface/internal gravity oscillations in the Strait, although the quasi-steady solutions are established by these time dependent motions which should not be neglected in the short-term (Pratt, 1987; Helfrich, 1995). We then make use of the evidence on the existence of the two basic controls at the contraction and northern sill and integrate equations (1-4) between the two control sections. This is a model useful for studying the response of the Bosphorus to seasonal forcing. It is a simple variation of the Farmer and Armi (1986) model, allowing for a finite depth at the contraction, variable width in each layer, and also including barometric pressure differences between the adjoining basins, and wind stress along the strait. A further trivial difference, *i.e.* the presence of a free surface, is kept to allow seasonal storage in the Black Sea basin, although the rigid-lid assumption would be valid in this approximation. We parameterize bottom and interfacial by integrating the frictional terms between the two conrol sections (*e.g.* Sümer and Bakioğlu, 1981). The model equations are:

$$S_b \frac{\partial \eta_b}{\partial t} = Q_f + Q_2 - Q_1 \quad (5)$$

$$\frac{1}{2}\{u_{1s}^2\} = \frac{\Delta_{bs}}{\rho_1} + g(\eta_b - \eta_s) \quad (6)$$

$$\frac{1}{2}\{u_{1c}^2 - u_{1s}^2\} = \frac{\Delta_{sc}}{\rho_1} + g\eta_s - \int_{x_c}^{x_s} \frac{\tau_w}{\rho_1 y_1} dx - \int_{x_c}^{x_s} \frac{\tau_i}{\rho_1 y_1} dx \quad (7)$$

$$\frac{1}{2}\{u_{2c}^2 - u_{2s}^2\} = \frac{\Delta_{sc}}{\rho_2} + g\{\eta_s - \varepsilon(y_{1s} - y_{1c})\} + \int_{x_c}^{x_s} \frac{\tau_i}{\rho_1 y_2} dx + \int_{x_c}^{x_s} \frac{\tau_b}{\rho_2 y_2} dx \quad (8)$$

$$\frac{u_{1c}^2}{\varepsilon g y_{1c}} + \frac{u_{2c}^2}{\varepsilon g y_{2c}} = 1 \quad (9)$$

$$\frac{u_{1s}^2}{\varepsilon g y_{1s}} + \frac{u_{2s}^2}{\varepsilon g y_{2s}} = 1 \quad (10)$$

$$y_{1c} + y_{2c} = y_o \quad (11)$$

$$y_{1s} + y_{2s} + H_s = \eta_s + y_o \quad (12)$$

where subscript s and c respectively denote values at the sill and the contraction, S_b and η_b the surface area and sea level in the Black Sea, y_o the depth at contraction, H_s and η_s the bottom and surface elevation at the sill relative to the contraction, Q_f the net freshwater inflow, $Q_i = u_{ic} b_{ic} y_{ic} = u_{is} b_{is} y_{is}$ ($i = 1, 2$ the layer fluxes, $\Delta_{bs} = P_b - P_s$ and $\Delta_{sc} = P_s - P_c$ the barometric pressure differences between the Black Sea and the sill and the sill and contraction sections. τ_w is the along-strait wind stress, and τ_i and τ_b are the internal and bottom frictional stresses. $\varepsilon \equiv \Delta\rho/\rho_2$ is the density stratification parameter.

(6-8) are Bernouilli equations, and (9,10) are the hydraulic control conditions $\mathbf{G}^2 = \mathbf{F}_1^2 + \mathbf{F}_2^2 = 1$, where $\mathbf{F}_i = u_i/\sqrt{\varepsilon g y_i}$ is the *Froude number* for each layer $i = 1, 2$. Equations (11,12) geometrically relate the thickness of layers with respect to a common datum. The non-dimensional versions of (5-12) are

$$\frac{\partial \zeta_b}{\partial \tau} = q_f + \beta q_2 - \beta^{-1} q_1$$

$$B_1^2 q_1^2 d_{1s}^{-2} = \zeta_b - \zeta_s + \delta_{bs}$$

$$q_1^2\{d_{1c}^{-2} - B_1^2 d_{1s}^{-2}\} = \zeta_s + \delta_{sc} - \{\nu_1 \pi_w + \kappa_i \mu_1 q_1 + \beta^{-1} \kappa_i \mu_2 q_2\}$$

$$q_2^2\{d_{2c}^{-2} - B_2^2 d_{2s}^{-2}\} = \zeta_s - 2(d_{1s} - d_{1c}) + r\delta_{sc} + \{\beta \kappa_i \mu_1 q_1 + (\kappa_i + \kappa_b)\mu_2 q_2\}$$

$$B_1^2 q_1^2 d_{1s}^{-3} + B_2^2 q_2^2 d_{2s}^{-3} = 1$$

$$q_1^2 d_{1c}^{-3} + q_2^2 d_{2c}^{-3} = 1$$

$$d_{1c} + d_{2c} = 1$$

$$d_{1s} + d_{2s} + h_s = \frac{\varepsilon}{2}\zeta_s + 1$$

with the following non-dimensionalizations and definitions:

$$q_i = \frac{Q_i}{Q_{ic}}, \quad Q_{ic}^2 = \varepsilon g y_0^3 b_{ic}^2, \quad i = 1, 2, \quad q_f = \frac{Q_f}{Q_c}, \quad Q_c^2 = Q_{1c} Q_{2c} = \varepsilon g y_o^3 b_{1c} b_{2c}$$

$$\zeta_b = \frac{2}{\varepsilon}\frac{\eta_b}{y_o}, \quad \zeta_s = \frac{2}{\varepsilon}\frac{\eta_s}{y_o}, \quad d_{ic} = \frac{y_{ic}}{y_o}, \quad d_{is} = \frac{y_{is}}{y_o}, \quad i = 1, 2,$$

$$\tau = \frac{2Q_c}{\varepsilon S_b y_o} t, \quad \xi = x/L,$$

$$\delta = \frac{2}{\rho_1 \varepsilon g y_o}\Delta, \quad \pi_w = \frac{2}{\rho_1 \varepsilon g y_o}\frac{L}{y_o}\tau_w, \quad \kappa_i = \frac{2L}{\sqrt{\varepsilon g y_o}} k_i, \quad \kappa_b = \frac{2L}{\sqrt{\varepsilon g y_o}} k_b,$$

$$h_s = \frac{H_s}{y_o}, \quad B_1 = \frac{b_{1c}}{b_{1s}}, \quad B_2 = \frac{b_{2c}}{b_{2s}}, \quad \beta = \frac{b_{2c}}{b_{1c}},$$

$$\nu_j = \int_{\xi_c}^{\xi_s} \frac{1}{d_j} d\xi \simeq \frac{ln(d_{js}/d_{jc})}{d_{js} - d_{jc}}, \quad \mu_j = \int_{\xi_c}^{\xi_s} \frac{b_{jc}}{b_j}\frac{1}{d_j} d\xi \simeq \frac{b_{jc}}{\langle b_j \rangle}\nu_j, \quad j = 1, 2.$$

Linear internal and bottom friction have been assumed and their integrals between the sill and constriction have been approximated between the two control sections, using the length scale for the strait $L = x_s - x_c$, where $\langle b_j \rangle$ is the average width between the two control sections.

Equations (13-20) are solved iteratively to determine the eight unknowns ζ_b, ζ_s, q_1, q_2, d_{1c}, d_{2c}, d_{1s}, d_{2s}, for given values of forcing q_f, δ_{bs}, δ_{sc}, π_w, geometry B_1, B_2, β, h_s, friction κ_i, κ_b, stratification ε, and adjustable

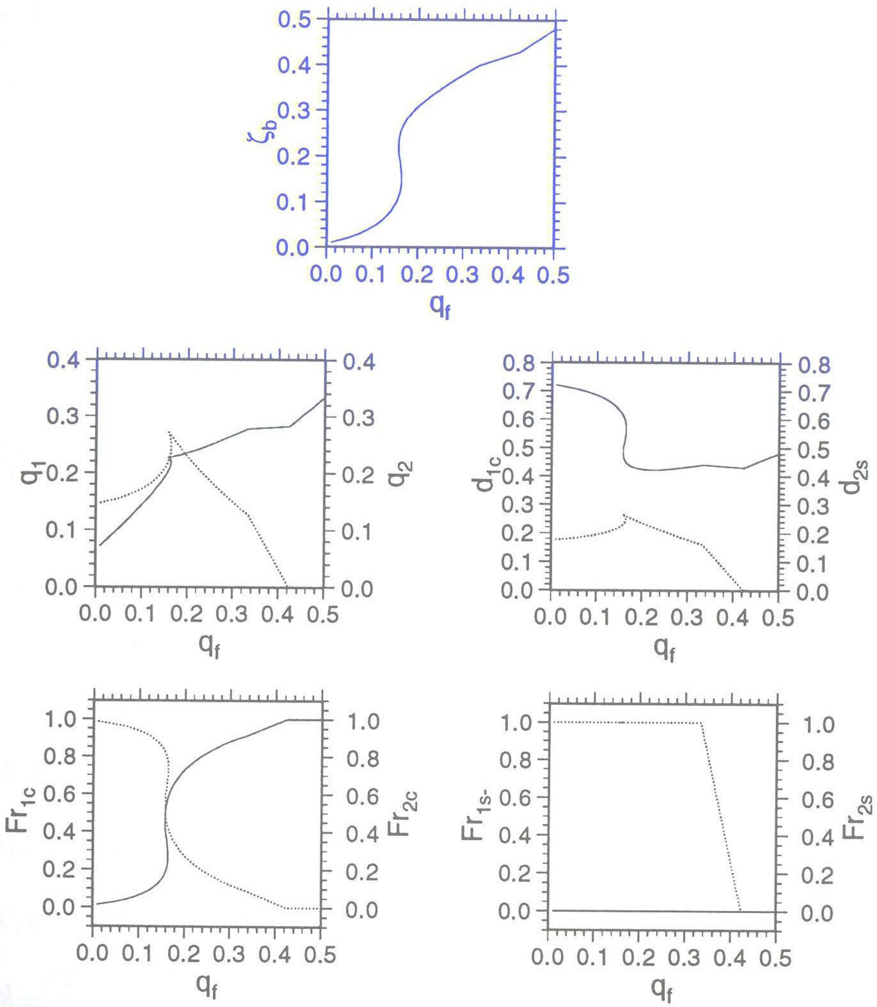

Figure 9. Frictionless, steady-state solutions for values $B_1 = 0.01$, $B_2 = 0.5$, $h_s = 0.15$, $\beta = 0.667$, $\varepsilon = 0.015$ appropriate for the Bosphorus. The Black Sea sea level ζ_b, layer fluxes q_1, q_2, upper layer depth at contraction d_{1c}, lower layer depth at the sill d_{2s}, and Froude numbers $Fr_{ih} = u_{ih}^2/gy_{ih} = q_{ih}^2/d_{ih}^3$ for each layer $i = 1, 2$ at the two control sections $h = c, s$, are plotted as a function net barotropic flux q_f

constants ν_1, μ_1, μ_2. The blocking of the flow in either layer is checked and appropriate equations are used during the iterations.

The steady solutions for appropriate parameter values for the Bosphorus are shown in Figure 9, displaying, nonlinear dependence of the sea level ζ_b, layer fluxes q_1, q_2 and interface heights d_{1c}, d_{2s} at the control sections on net water flux q_f. Furthermore one can verify that the layers make equal contributions to the composite Froude number at the contarction, but the

lower layer dominates in the case of the sill. Blocking of the lower layer occurs as q_f is increased.

5. Time-Dependent Response and a Barotropic Model

Current-meter measurements and acoustic Doppler current profiler (ADCP) measurements show large transient changes in the Bosphorus fluxes even within a single day (Latif *et al.*, 1991; Özsoy *et al.*, 1994, 1995, 1996) as result of meteorological, tidal and other natural oscillations under time dependent forcing. On seasonal time scales, the net fresh water input into Black Sea, barometric pressure differences, and wind setup are the possible time-dependent forcing mechanisms for the Bosphorus. Both the hydrological cycle and barometric pressure variations are clearly related with the sea level fluctuations in the Black Sea (Sur *et al.*, 1994, Özsoy *et al*, 1996). It is evident from observations that the Bosphorus operates in the full range of weak to strong barotropic forcing in either direction. The response of the Strait to such forcing is non-linear, and in addition, blocking of the flows in either layer occurs when the net flow exceeds critical values, lasting for a few days each time.

The seasonal behaviour of the Bosphorus, excluding the short-term fluctuations, can be studied using the model presented above. In the inviscid, stationary case, the solutions are multiple valued in ζ_b for given q_f, with two stable and one unstable solutions; therefore seasonal solutions fluctuate around two-fixed points and can jump from a low sea level state to a high sea level state. The multiple valued solutions cease to exist when appropriate friction is included, but the basic nonlinear characteristics survive, resulting in multi-annual fluctuations (Özsoy *et al.*, 1996). These characteristics can be detected in the long-term sea level measurements in the following sections.

In order to study the time-dependent response of the coupled strait-basin systems, it is necessary to consider using the full set of nonlinear equations governing the system, since it is important to represent the surface and internal waves serving to establish the hydraulic controls in the straits (Pratt, 1984). On the other hand, the time dependent equations may not be easy to solve, for a system with two straits coupled with adjacent basins, because one needs to consider the geometry of the entire strait rather than the geometry of the control sections (Helfrich, 1995). For quasi-steady solutions to be valid, the time scale T of the motions studied must be much greater than the travel time of interfacial bores through the strait, *i.e.* for $T \gg T_c$ where $T_c L/\sqrt{\varepsilon g y_0}$ (Helfrich, 1995), preferably $T/T_c > 30$. For the Bosphorus, we can estimate $T_c \simeq 1.2d$. We see that this is very close to the period of some observed motions, *e.g.* diurnal and inertial periods, and as we will see below, it is also close to the some natural periods of the coupled system. We can foresee that the quasi-stationary model presented above can only be used for seasonal fluctuations, *i.e.* for motions with periods longer than a month.

As a compromise, we consider a simplified, linear model of the pure barotropic motions in the Straits, following Candela *et al.* (1989) and Le-Traon and Gauzelin (1997), who studied coupled oscillations of the Eastern and Western Basins of the Mediterranean with the Sicily and Gibraltar Straits. We assume that sea level changes are uniform in the Marmara and Black Seas, while unsteady, barotropic momentum equations describe the motions in the interconnecting Bosphorus and Dardanelles Straits. We write the basin continuity and strait momentum equations as follows:

$$S_b \frac{\partial \eta_b}{\partial t} = Q_f - A_b u_b, \tag{13}$$

$$S_m \frac{\partial \eta_m}{\partial t} = A_b u_b - A_d u_d, \tag{14}$$

$$\frac{\partial u_b}{\partial t} = \frac{\Delta_{bm}}{\rho L_b} + \frac{g}{L_b}(\eta_b - \eta_m) - \lambda_b u_b, \tag{15}$$

$$\frac{\partial u_d}{\partial t} = \frac{\Delta_{ma}}{\rho L_d} + \frac{g}{L_d}(\eta_m - \eta_a) - \lambda_d u_d, \tag{16}$$

where S_b and S_m are respectively the surface area of the Black Sea and Marmara Sea, η_b, η_m and η_a are the sea levels, and P_b, P_m and P_a are respectively the barometric pressures in the Black, Marmara and Aegean Seas, $\Delta_{bm} = P_b - P_m$ and $\Delta_{ma} = P_m - P_a$ are their differences, A_b, A_d and L_b, L_d are respectively the the cross-sectional areas and the lengths of the Bosphorus and the Dardanelles Straits, u_b and u_d the barotropic velocities in the Straits, and λ_b and λ_d are the linear friction coefficients. Note that the total hydrostatic pressure p is the sum of barometric pressure and the pressure in water $p = P + \rho g(\eta + z)$ at depth z below the surface.

By eliminating u_b and u_d we can reduce the system to coupled equations for sea level η_b and η_m:

$$\frac{\partial^2 \eta_b}{\partial t^2} + \lambda_b \frac{\partial \eta_b}{\partial t} + \alpha_b \eta_b - \alpha_b \eta_m = \frac{1}{S_b}\left(\frac{\partial Q_f}{\partial t} + \lambda_b Q_f\right) - \beta_b \Delta_{bm} \tag{17}$$

$$\frac{\partial^2 \eta_m}{\partial t^2} + \lambda_d \frac{\partial \eta_m}{\partial t} + (\alpha_d + s\alpha_b)\eta_m - s(\lambda_d - \lambda_b)\frac{\partial \eta_b}{\partial t} - s\alpha_b \eta_b - s\alpha_d \eta_a$$
$$= -(\lambda_d - \lambda_b)\frac{1}{S_m} Q_f - \beta_d \Delta_{ma} + s\beta_b \Delta_{bm} \tag{18}$$

where we have also defined

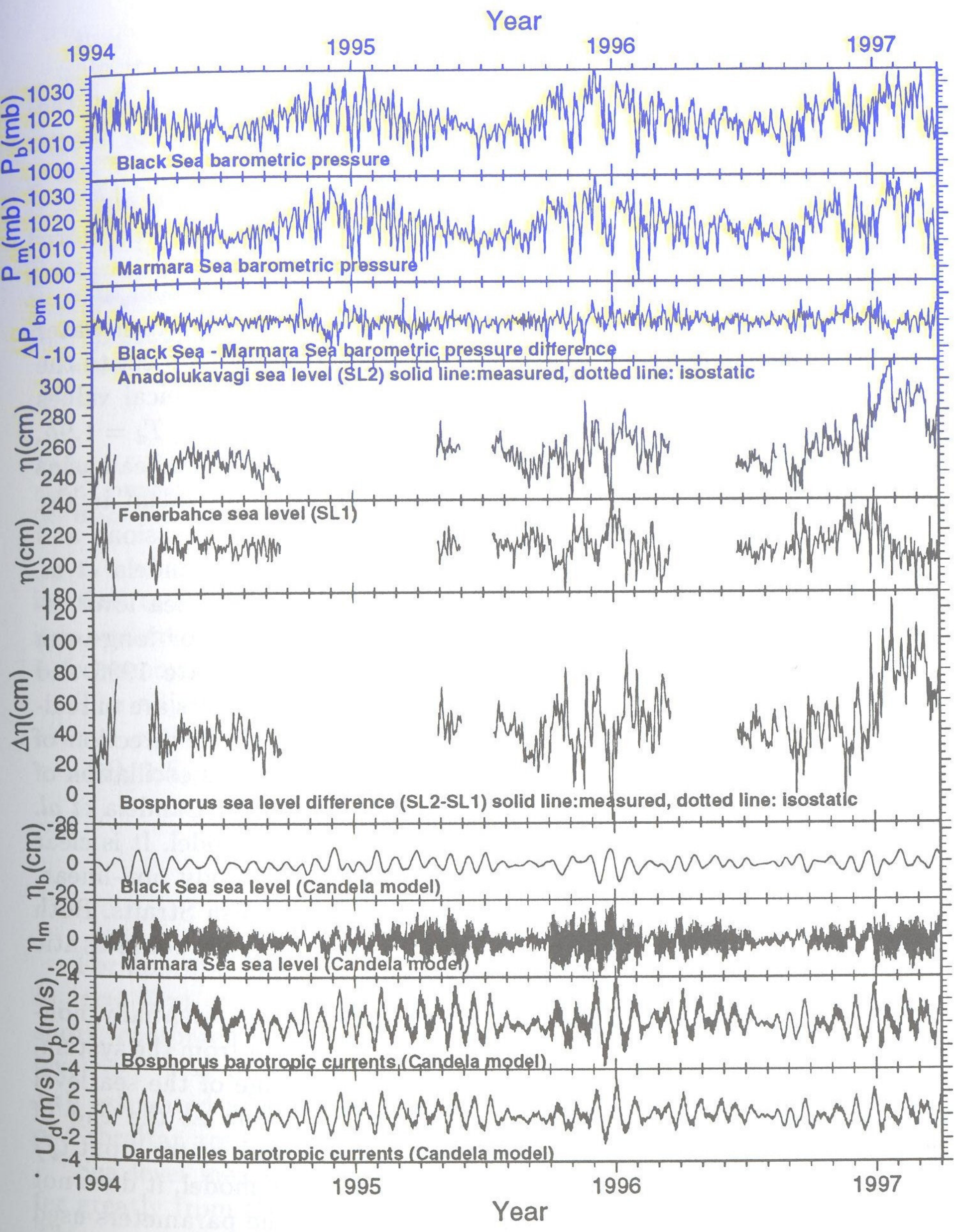

Figure 10. 1994-1997 time series of observed quantities and results from the Candela *et al.* (1989) model. From top to bottom: spatially averaged daily sea level pressure for the Black Sea (Zonguldak, Sinop, Trabzon, Odessa and Kerch), Marmara Sea (Çanakkale, Tekirdağ, Bandırma), pressure difference, sea level measurements at Anadolukavağı (SL2) and Fenerbahçe (SL1), sea level difference, Black Sea and Marmara Sea sea level and Bosphorus and Dardanelles Strait barotropic currents predicted by the Candela *et al.* (1989) model ($A_b = 42 \times 10^3 m^2$, $A_d = 60 \times 10^3 m^2$, $L_b = 40km$, $L_d = 100km$, $S_b = 42 \times 10^4 km^2$, $S_m = 1.15 \times 10^4 km^2$, $\lambda_b = 3 \times 10^{-8}$, $\lambda_d = 0.9 \times 10^{-8}$, $T_b = 14.7h$, $T_d = 1.9h$).

$$\alpha_b = \frac{gA_b}{L_b S_b}, \quad \alpha_d = \frac{gA_d}{L_d S_m}, \beta_b = \frac{\alpha_b}{\rho g}, \quad \beta_d = \frac{\alpha_d}{\rho g}, s = \frac{S_b}{S_m}.$$

The characteristic periods of oscillation of the coupled system, *i.e.* the natural periods of the Black Sea and Marmara Sea (excluding the effects of friction) are

$$T_b = \frac{2\pi}{\sqrt{\alpha_b}} = \frac{2\pi}{\sqrt{\frac{g}{S_b}\frac{A_b}{L_b}}}, \quad T_m = \frac{2\pi}{\sqrt{\alpha_d + s\alpha_b}} = \frac{2\pi}{\sqrt{\frac{g}{S_b}\frac{A_b}{L_b} + \frac{A_b}{L_b}}}.$$

The model equations driven by the respective pressure and fresh water forcings, Δ_{bm}, Δ_{ma} and Q_f, are solved in Figures 10 and 12, subject to the boundary condition $\eta_a = 0$ on the Aegean Sea exit. With typical values assigned, the characteristic periods are found to be $T_b = 14.7h$, $T_d = 1.9h$, reflecting the differences in the inertia of the Black and Marmara Sea basins coupled through the Straits. Both of these time-scales are evident in the calculations of Figure 10, using characteristic geometric dimensions and realistic friction terms on the same order of magnitude as Candela *et al.* (1989). However, the model fails to reproduce the observed sea level response of both Seas, and differs from the data especially for long-term fluctuations and abrupt changes observed for instance, in late 1995 and early 1997. Furthermore, the barotropic velocities at the Straits are unrealistically large and display bi-weekly fluctuations changing the direction of net currents. These results contrast strongly with the coupled oscillation of the Mediterranean Straits predicted in a much better way by Candela *et al.* (1989) and LeTraon and Gauzelin (1997) using the same model. It is clear that the simplified model is not appropriate for the strongly non-linear, stratified system consisting of the Black Sea and the Turkish Straits. Both observations and model results are far from the smaller amplitude isostatic response $\eta = -(P - P_o)/\rho g$ displayed in Figure 11.

It is evident that the Bosphorus poses a stronger and essentially non-linear resistance to the hydrological excess flow leaking out from the system. This is evident especially if one considers the large range of the sea level difference across the Strait, increasing from about -10cm at the end of 1996 to about 110cm in early 1997. Although the net fresh water input Q_f (mean monthly averages, shown in Figure 12) drives the model, it does not produce appreciable seasonal change of sea level with the parameters used in Figure 10. When the friction parameters are increased several orders of magnitude, realistic seasonal changes in sea level and currents (with net outflows) can be reproduced, as shown in Figure 12, at the cost of damping inertial oscillations of the coupled mass system. This result shows that other types of flow restriction, *e.g.* resulting from the non-linear maximal exchange or non-linear frictional drag of the strait geometry should be

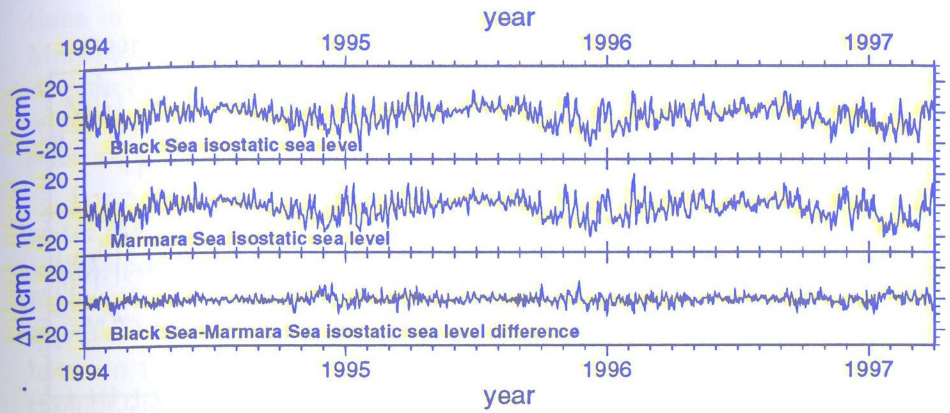

Figure 11. 1994-1997 time series of sea level computed from isostatic response to barometric pressure, $\eta = (P - P_o)/\rho g$ (a constant mean pressure of $P_o = 1015mb$ is used), (a) Black Sea, (b) Marmara Sea and (c) difference Black Sea - Marmara Sea.

invoked to account for the slow leakage from the Black Sea. Furthermore, it is evident that the low frequency flow resistance offered by the Bosphorus is frequency dependent.

The adjustment of Black Sea sea level to forcing in the maximal exchange regime is characterized by the time scale (based on the scaling $\tau = 2Q_c/\varepsilon S_b y_o t$ above)

$$T_a = \frac{\varepsilon S_b y_o}{2Q_c} = \frac{S_b\sqrt{\varepsilon}}{2\sqrt{g y_o} b_{1c} b_{2c}}.$$

For appropriate values $\varepsilon = 0.015$, $S_b = 420 \times 10^3 km^2$, $y_o \simeq 100m$, $b_{1c} \simeq b_{2c} \simeq= 1km$, we find $T_a \simeq 42d$. This adjustment time scale is relatively short, compared to the filling time scale $T_f = S_b\bar{\eta}_b/\bar{Q}_f$ (required for sea level rise of $\bar{\eta}_b$ due to net river inflow of $\bar{Q}_f$ in a closed basin - equation 1a), estimated as $T_f = 240d$. In comparison, the time scales for the Strait of Gibraltar are estimated to be $T_a \simeq 9d$ and $T_f \simeq 25d$.

Sea level measurements at Constanza and Sevastopol (Figure 12) differ greatly from those in the northern Bosphorus during the same period. This comparison shows that the large along-strait sea level gradients in the Bosphorus rapidly decline outside of the Strait, and therefore the sea level at the Anadolukavağı (SL2) station can not be compared with the average sea level in the Black Sea. The difference in sea level between the Bosphorus and the Black Sea should be accountable by the Bernouilli set-down $\Delta\eta_b = u^2/2g$ upon exit from the Strait. For the observed $\Delta\eta_b \simeq 0.5m$, the order of magnitude estimate for barotropic velocity would be $u = 3m/s$.

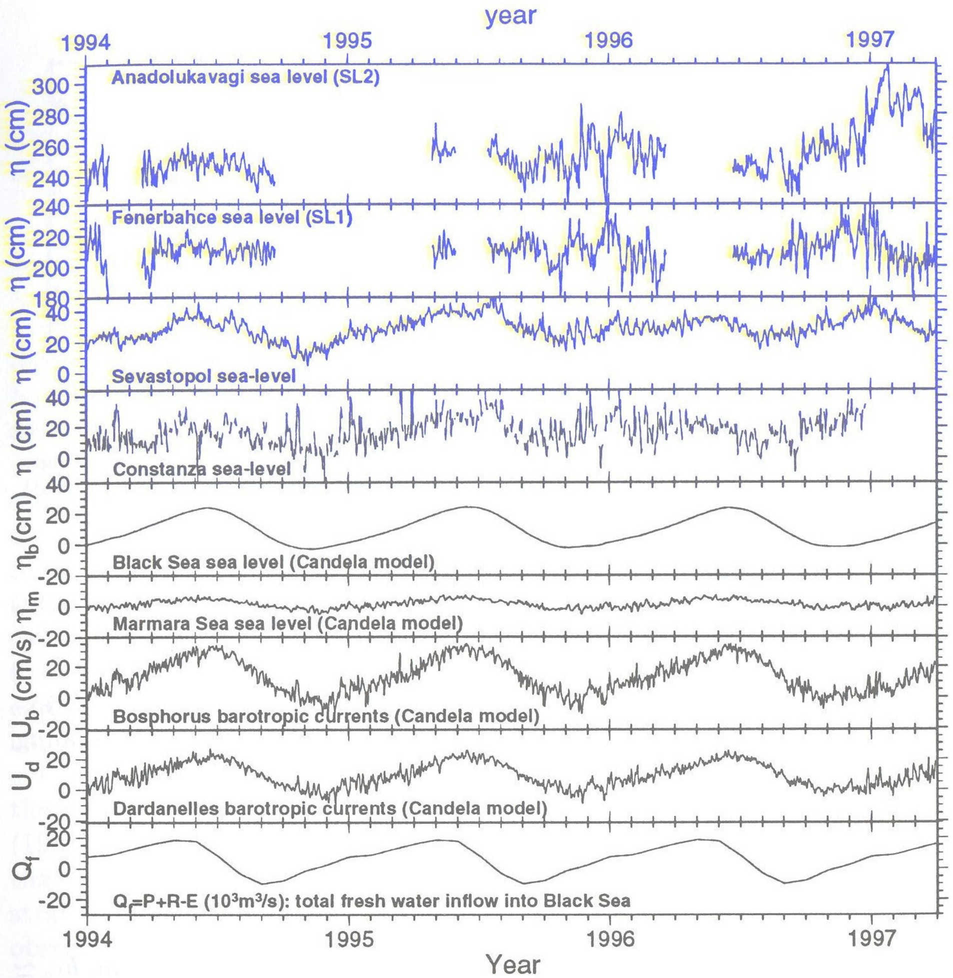

Figure 12. Daily average time series of sea level measurements at Anadolukavağı (SL2) and Fenerbahçe (SL1), Sevastopol and Constanza (measurements at 7:00 am only), Black Sea and Marmara Sea sea level predicted by the Candela *et al.* (1989) model ($\lambda_b = 1.5 \times 10^{-4}$, $\lambda_d = 0.3 \times 10^{-4}$), and monthly average water budget $Q_f = P + R - E$ of the Black Sea

This is rather large for the Helmholtz mode oscillation we have so far envisioned in the Black Sea, and calls to attention other types of possible rotational and divergent topographically trapped motions and dissipation in the Black Sea, driven by the the Strait pressure differences, such as treated by Candela and Lozano (1994).

The larger than isostatic amplitudes of low frequency sea level oscilla-

tions in the Black Sea, especially near the Bosphorus, and in the Sea of Marmara, suggest connection with similar overisostatic oscillations in the Aegean Sea for periods larger than 10 days (Tsimplis and Vlahakis, 1994), which may result from the restrained leakage from the Black Sea of the seasonal fresh water inputs, and the efficient controls at the Bosphorus.

Long-term, monthly time series of sea level at Sevastopol, and net fresh water inflow, with components of total river runoff, estimated precipitation and evaporation in the Black Sea (Simonov and Al'tman, 1991) displayed in Figure 13 verify close relationship between various elements of the hydrological cycle and sea level variations. The most direct relationship is observed between the river inflow and sea level, based on direct measurements. Although the other elements of surface fluxes involve certain assumptions and calculations, the relationship of sea level to total fresh water fluxes survives, especially for large events.

From the above, it is not clear if the long term sea level response is predominantly determined by pressure or fresh water forcing. In Figure 14, monthly time series for Sevastopol sea level is compared with Danube river discharge and barometric pressure at Sinop. The sea level fluctuations are seen to be much larger than the isostatic sea level, and not particularly well-correlated to either the pressure or fresh water effects. It seems that both factors contribute, and furthermore the annual and interannual sea level changes are far greater than isostasy would suggest, even at these low frequencies.

6. Conclusions

The Bosphorus represents an outstanding case of the 'maximal exchange' Strait. While a great deal still needs to be learned about the exact nature of the exchange between two basins possessing the world's most distinct physical characteristics (the Black Sea and the Mediterranean), even a cursory examination reveals interesting possibilities for driving mechanisms and complex response characteristics on a variety of time scales. The sea level variations are far greater than linear theory would predict. The Bosphorus response to fresh water leakage and pressure is frequency dependent, greatly restrictive at low frequencies, and strongly non-linear. Sea level at the northern or southern ends of the Bosphorus greatly differs from the average sea level in the Black Sea. The Marmara Sea is largely driven by the Black Sea and Bosphorus oscillations, while the Dardanelles contributes to high frequency fluctuations with periods of about a few days. The flow restrictions at the Bosphorus, and the slow response in the Black Sea could partly explain the observed overisostatic response in the Aegean Sea and the Eastern Mediterranean. Further analyses of data, and direct measure-

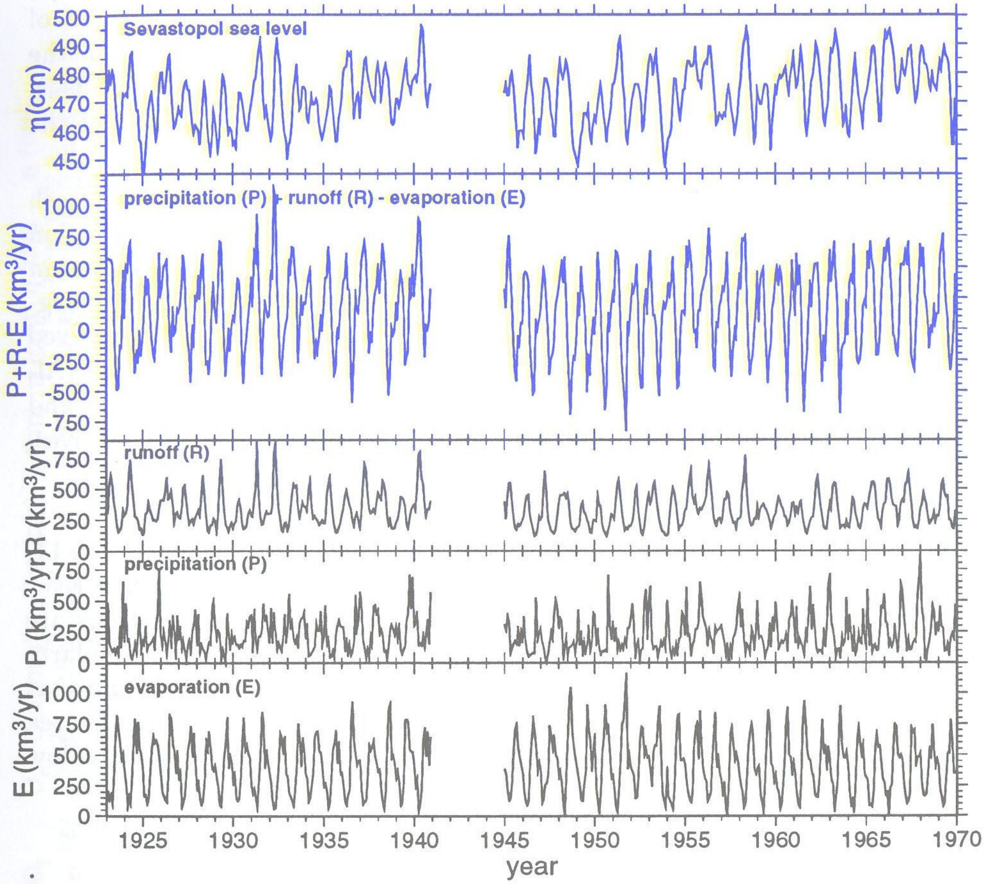

Figure 13. 1923-1970 monthly mean time series for sea level at Sevastopol (η), total freshwater inflow $Q_f = P + R - E$, precipitation (P), river runoff (R), and evaporation (E) in the Black Sea, the hydrological data are after Simonov and Al'tman (1991).

ments of fluxes and sea level, are needed for a better understanding of the Turkish Straits - Black Sea coupled oscillations at periods ranging from a day to several years.

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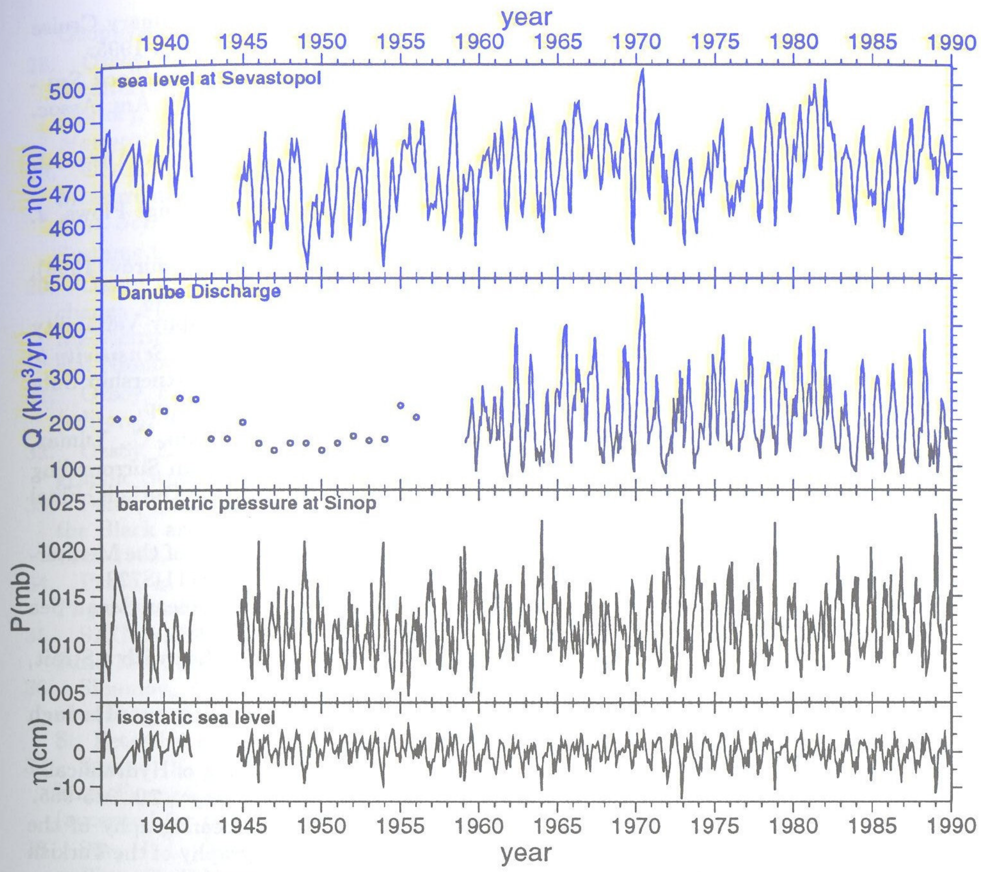

Figure 14. 1936-1990 monthly mean time series for sea level (η) at Sevastopol, discharge Q of the Danube river, barometric pressure (P) at Sinop, and the computed isostatic sea level ($\eta_i = (P - \bar{P})/\rho g$, where $\bar{P}$ is taken as $\bar{P}$=1012mb).

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