WATER VOLUME CHANGE IN THE LOWER MEKONG FROM SATELLITE ALTIMETRY AND IMAGERY DATA

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Abstract

We have analysed satellite altimetry data from the ERS-2, ENVISAT and Topex/Poseidon satellites to construct water level time series over a 8-year period (from April 1996 to April 2004) over the lower Mekong River basin. The study area includes the Tonle Sap Lake, seasonally inundated areas and several branches of the hydrographic network of the Mekong delta. We found a very strong seasonal signal over the main river north of 13°N, the Tonle Sap Lake and Tonle Sap River, with amplitudes reaching 8-10 meters annually. We also found a clear interannual signal in altimetry-derived water level time-series. For example, year 1999 had weak floods (around 6 m amplitude), contrasting with year 2000 during which strong flood was noticed (around 10 m amplitude). Southward, we also observed large seasonal fluctuations (2-3 m) over inundated floodplains, as identified using satellite imagery data from the SPOT-4 Vegetation instrument. Depending on the location, quite different annual amplitudes were observed, the closer to the Mekong mouth, the smaller the signal (less than 0.5 m seasonal amplitude). Using NDVI (Normalized Difference Vegetation Index) data from the Vegetation instrument, we studied the seasonal extent of flood plains in the delta. Then combining the areal extent of floods with water levels estimated from the ERS-2/ENVISAT data, we computed maps of monthly surface water volume change over six successive years (1998-2003), the period of availability of the NDVI data. Averaged over the lower Mekong basin, this surface water volume change was then compared to the total (i.e., surface plus underground) water volume change inferred from the GRACE satellite. They exhibit in phase fluctuations.

Keywords: radar altimetry, multispectral imagery, water volume change

Abbreviated title: Water volume change in the lower Mekong

1. Introduction

The Mekong river basin is one of the largest of the world, climatically controlled, but with some anthropogenic alterations, especially in the lower part and in the delta (MRC, 2005). The annual recurrence of the monsoon floods in the Cambodian floodplains, from August to November, is of great importance for farming and economic activity. Between 80 % and 90 % of the freshwater use in the Mekong basin is for irrigating rice crops (MRC, 2002). In some years, vast regions are flooded for several weeks, causing serious damage. The year 2000, for instance, was characterized by a particularly severe flood, reported as the most devastating in the last 40 years (Dutta and Takeushi, 2000; MRC, 2005).

Water storage in wetlands and corresponding outflow represent a significant part of the water balance of large river basins (Alsdorf et al., 2001). To better understand the hydrology of large river systems, information about the dynamics of inundation patterns (extent of flooded areas) and water levels of main river channels, tributaries and associated floodplains is required (Alsdorf et al., 2000).

Spatial and temporal patterns of inundation areas have been inferred from multi-temporal satellite visible/infrared images, Synthetic Aperture Radar sensors or microwave radiometers (Mertes et al., 1995 ; Smith, 1997 ; Sippel et al., 1998; Töyrä et al., 2001 ; Hess et al., 2003). In South America, inundation patterns of large floodplains have been derived using the polarization difference at 37 GHz of the Scanning Multichannel Microwave Radiometer (SMMR) passive microwave emission measurements (ground resolution of 0.25°) (Hamilton et al., 2002). Relations between stage and flooded area have been derived using records from in-situ gauge-stations (Sippel et al., 1998).

Hydrological models have been used to determine the extent of flooded areas too. For instance, in the Amazon basin, the extension of the inundated areas was determined using a terrestrial ecosystem model (Integrated BIosphere Simulator - IBIS) and a HYDrological Routing Algorithm (HYDRA) by Coe et al. (2002).

In the lower Mekong basin, the flood extent has been studied using satellite imagery from quick-look mosaics of the SPOT satellite (Chia et al., 2001) and the Special Sensor Microwave / Imager (SSM/I) (Tanaka et al., 2003). The complete hydrological state of the Mekong River and its tributaries was simulated for the period 1994-1998 (Kite et al., 2001). This study provided an estimate of the total outflow of the Mekong basin and area of land flooded by Tonle Sap River.

Recently, a new methodology, based on the combined use of radar altimetry and satellite imagery, was developed to determine volume variations of surface water in the Negro river basin, the main tributary of the Amazon River (Frappart et al., 2005).

In the present study, we apply a similar method to monitor the volume variation of surface water during seasonal floods (August to December) in the lower Mekong basin over a 6-year period (1998-2003), using multispectral imagery from the Vegetation instrument onboard the SPOT 4 satellite and altimetry-derived water levels from the Topex/Poseidon (T/P), ERS-2 and ENVISAT satellites.

2. Hydrological characteristics of the study area

The Mekong basin is the largest river basin in South East Asia with an area of 795,000 km² (IMC, 1988). It is the home for over 65 millions inhabitants (MRC and UNEP, 1997). The Mekong river has its source on the Tibetan Plateau and then flows through six countries (China, Burma, Laos, Thailand, Cambodia and Vietnam) before reaching the south China Sea. With a mean annual flow of 475 000 m^3 and a total length 4800 km, it is the world's tenth greatest and twelfth longest respectively (IMC, 1988). The lower Mekong basin, which drains a total catchment area of 606,000 km² (77 % of the basin), is considered as the most important part of the Mekong basin, both environmentally and economically (IMC, 1988; Hori, 2000). The study area is the lower Mekong River basin and delta. At Phnom Penh (Cambodia), the Mekong River divides into three parts: the Tonle Sap River, the Bassac River, and the lower Mekong. During the flood season, water is drained from the Mekong and Bassac rivers into the Tonle Sap Lake, through the Tonle Sap River, and when the water level decreases in the Mekong River, the water is drained from the Tonle Sap Lake into the Mekong and Bassac rivers. The reversal of the flow of the Tonle Sap river is a unique hydrological feature that is responsible for a higher dry season flow in the delta than if it only received water from the Mekong River (Hoanh et al., 2003). When reaching the Vietnam border, the main river divides into several branches which constitute the Mekong delta. This region is very flat (less than 100 m elevation) and is referred as the Mekong lowlands (MRC, 2003). Figure 1 is a sketch of the study area.

The hydrological regime of the Mekong River is primarily dependent on the climatic conditions of the alternating wet and dry seasons. Climate is governed by the moonson winds that blow northeast or southwest depending on the period of the year. The southwest moonson corresponds to the rainy season which ranges from May to October, with a peak in September. To depict the precipitation patterns over the study area, we have applied Empirical Orthogonal Functions (EOF) decomposition, based on Lanczos orthogonalizations and Singular Value Decompositions (SVD) of large linear systems (e.g., Toumazou and Crétaux, 2001), to precipitation data. The data used here are monthly 1°x1° gridded precipitation of the Global Precipitation Climate Center based on quality-controlled data from 7,000 stations (Rudolf et al., 1994; Rudolf et al., 2003). Figure 2 shows the leading mode of the EOF decomposition of precipitation data over the lower Mekong, over the 8-year period of analysis (April 1996 to April 2004). The temporal curve shows a clear annual cycle, peaking in

September. As indicated by the spatial pattern map, the wettest region is the southern uplands and lowlands, south of the Tonle Sap Lake and northern part of the delta.

Superimposed on the annual cycle, some interannual fluctuations are visible. Fig.2 indicates two particularly rainy years: 1997 and 2000. The lowlands suffer annual floods during the rainy season. During the rainy year 2000, a particularly devasting flood affected a very large area. This event will be discussed below in some detail.

3. Satellite Datasets

In this study, we use different satellite datasets: radar altimetry data from the T/P, ERS-2 and ENVISAT satellites for measuring water level changes, and NDVI data from the Vegetation instrument onboard the SPOT-4 satellite to determine the surface water extent over inundated plains.

3.1 ERS-2, ENVISAT and Topex/Poseidon altimetry data

ERS-2, launched in 1995 by the European Space Agency (ESA), is the successor of ERS-1, designed to study the Earth environment. The satellite carries, among other instruments, a radar altimeter developed for measuring sea surface height. However, retracking of the raw altimeter waveforms (radar echoes) allows the use of the ERS data over continental surface waters (Legrésy, 1995). Here we only use the ERS-2 retracked altimetry data because the ERS-1 retracked data suffer too many gaps. The ERS-2 altimetry time series (20 Hz data – corresponding to an along track ground resolution of about 300 m) used here starts in April 1996. The average intertrack spacing over the lower Mekong basin is 85 km, while the revisit time (orbital cycle) is 35 days. The ENVISAT satellite was launched on February 2002 by

ESA for environmental objectives (Gardini et al., 1995). The satellite's payload includes a radar altimeter operating at two frequencies. Its orbital characteristics are similar to those of the ERS satellites (i.e., same repeat cycle and ground-track spacing). The ENVISAT altimeter provides radar echoes over ocean, land and ice to measure sea surface height, surface water level variations over river basins and ice surface elevation (Wehr and Attema, 2001). For the ENVISAT mission, four different algorithms are operationally applied to radar echo data to provide altimeter height estimates (Zelli, 1999): each of them has been developed for a specific surface response. One for ocean (ESA, 2002), two for ice sheets (Bamber, 1994 and Legrésy, 1995) and one for sea ice (Laxon, 1994). In this study, we use the ENVISAT 20Hz height measurements contained in the Geophysical Data Records (GDRs) (ESA, 2002) from cycle 12/09/2002 to 10/04/2004 of ENVISAT Mission. For the ENVISAT data, we use the range measurements processed with the Offset Center Of Gravity (or Ice-1) retracking scheme (Wingham et al., 1986; Bamber, 1994) which is the best suited for hydrological applications (Frappart et al., 2006). For the ERS-2 data, we use the 20 Hz height from ERS-2 retracked with Ice-2 algorithm by the OSCAR project (Observations des Surfaces Continentales par Altimétrie Radar or Land Observations by Radar Altimetry) at LEGOS (Laboratoire d'Etudes en Océanographie Spatiale) in Toulouse (France). As of mid-2002, we use the ENVISAT data instead of ERS-2. Figure 3 shows the ERS-2/ENVISAT track coverage over an Orthorectified Landsat Thematic Mapper Mosaic (https://zulu.ssc.nasa.gov/) of the study area. On Figure 3 are also superimposed the few Topex/Poseidon (T/P) tracks its 1992-2002 orbit. From September 2002, a few months after the launch of its successor, Jason-1, T/P moved to a new orbit, midway between the former orbital tracks. We do not consider the latter data because of the too-short time series available. For T/P, we use the Geophysical Data Records (standard ocean data, AVISO data base) which have been shown to be precise enough for continental water studies (e.g., Birkett, 1998, Birkett et al., 2002; Maheu et al., 2003). The 10-year (1992-

2002) hydrological dataset derived from T/P on its former orbit was expected to be extended with data from Jason-1. Unfortunately, Jason-1 provides too few land surface water measurements due to loss of surface lock by the onboard tracker and inaccurate retracking procedure over land surface waters. The T/P orbital cycle is 10 days. Compared to the 35 days orbit cycle of ERS/ENVISAT, this is obviously more favourable to study water level variations. But this is at the expense of the coverage which is very limited. The purpose of using also the T/P data here is mainly to assess the precision of the ERS-2 data at the crossing points between two tracks. In effect, several studies have assessed the T/P-based water level time series by comparing with in situ gauges measurements (see below). Altimetry data of all satellites have been corrected for the classical geophysical and environmental corrections needed over land (instrumental, ionosphere, wet and dry troposphere, solid Earth and pole tide corrections).

3.2 VEGETATION imagery

The VEGETATION (VGT) sensor, onboard SPOT-4 satellite, was designed to observe the vegetation and land surfaces (Arnaud and Leroy, 1991). Since April 1998, this sensor acquired data in four spectral bands: B0 (blue, 430-470 nm), B2 (red, 610-680 nm), B3 (near infrared, 780-890 nm) and SWIR (short-wave infrared, 1580-1750 nm). With a swath width of 2250 km, VGT provides daily coverage at 1-km spatial resolution. Three standard products are delivered to users: VGT-P (physical product), VGT-S1 (daily synthesis product) and VGT-S10 (10-day synthesis product). There are three 10-day composites for a month: days 1-10, days 11-20 and day 21 to the last day of the month. VGT-S10 products were obtained selecting the VGT-S1 value that have the maximum Normalized Difference Vegetation Index (NDVI) within a 10-day period. This product, known as maximum NDVI value composite

(MVC), is used to minimize the effect of cloud cover and variability in atmospheric optical depth (Holben, 1986). The NDVI, defined by Tucker (1979) is computed using the equation:

$$NDVI_{VGT} = \frac{B3 - B2}{B3 + B2} \qquad (1)$$

NDVI is primarily used to derive vegetation parameters for crop and forest monitoring, for the study of mass and biogeochemical cycles, and for assimilation in general circulation or weather forecast models (Duchemin et al., 2002).

In this study, we use the VGT-S10 NDVI products from May 1998 to December 2004, available at http://free.vgt.vito.be/ to develop a method for mapping inundated areas.

4. Water level time series

4.1 Altimetry-derived water levels

Along the satellite tracks shown on Figure 3, we have considered 80 altimetry stations (called below virtual stations) at which we have computed water level time series from the ERS-2 and ENVISAT altimetry data. These altimetry stations are identified on Figure 3 by the white dots and yellow stars (the white dots correspond to total set of computed water level time series while the yellow stars only correspond to those shown in Figure 6). The virtual stations selected here are those which provide good quality water level time series (according to the data editing and error bars associated to each time series –see below-) and data gaps. Most stations are located over the floodplains in the delta lowlands. Other stations correspond to intersections of the satellite tracks with the river. Finally, two stations are located on the Tonle Sap Lake. To construct a water level time series, we consider all 20 Hz altimetry data along a portion of satellite track. The intersections between satellite tracks and rivers or floodplains are determined using an Orthorectified Landsat Thematic Mapper Mosaic (https://zulu.ssc.nasa.gov/) of 28.5 m resolution. Over rivers, the portion of satellite track considered corresponds to the river-satellite track intersection. Over floodplains, this portion ranges from 1 to ~5 km. Once selected, the data are expressed in terms of water height (water level) above the geoid (for that purpose, the GRACE geoid GGM02C, complete to degree and order 150, has been used; Tapley et al., 2005). For each intersection between the river (or the floodplain) and the satellite ground-track, we define a so-called “virtual station”, represented by a rectangular window. Outliers are deleted using a 3-sigma criterion on the whole time span of analysis. For each 35-day cycle, the water level at a given virtual station is obtained by computing the median of all the high-rate data (20Hz for ENVISAT and ERS-2 and 10Hz for T/P) included in the rectangular window. This process, repeated for each cycle, allows the construction of a water level time series at the virtual station.

The dispersion in L1 norm is given by the estimator known as median absolute deviation:

$$MAD(x) = \frac{1}{N-1}\sum_{i=1}^{N}\left|x_i - x_{med}\right| \qquad (2)$$

where MAD(x): median absolute deviation of the observations, N: number of observations, x_i: i^{th} observation, x_{med}: median of the observations.

4.2 Comparison of ERS-2/ENVISAT measurements with *in-situ* gauge data and T/P time-series

The accuracy of T/P water level time series over river and floodplains has been discussed in several previously published papers (i.e., Birkett, 1998; de Oliveira Campos et al., 2001; Birkett et al., 2002; Maheu et al., 2003). Recently, validation of several tens of T/P-based water level time series over rivers worldwide (through comparisons with *in-situ* gauge data) indicates an average precision of about 30 cm (Cauhopé et al., 2006). Water levels derived from ERS-2/ENVISAT were only used in few studies (Berry et al., 2005; Frappart et al., 2006), indicating an averaged uncertainty of about 20 cm for ERS-2 and 15 cm for ENVISAT.

We have compared the water level time series from ERS-2/ENVISAT with those from T/P at crossover points (intersections between the T/P and ERS-2/ENVISAT tracks). In Figure 4a is presented an example of ERS-2 and T/P time series at a cross over located in the floodplain (see the white open circle in Figure 3). We present an example (Figure 4a) of T/P, ERS-2 and ENVISAT time series at a crossover in the floodplain (see the white ellipse on Figure 3). In Figure 4a, the ERS-2 time series has been completed by the ENVISAT data beyond mid-2002. Error bars are estimated as explained above. Figure 4b presents the T/P-based and ERS-2 (ENVISAT)-based water level time series at the crossover point after removing the mean to each time series. The comparisons shown in Figure 4b indicate that the ERS-2 time series agrees very well with the T/P one, both in amplitude and phase. The difference between the two time-series is also presented in Figure 4b. The corresponding RMS difference is 16 cm. This comparison allows us to quantify the precision of the ERS-derived water levels over the Mekong floodplain.

The water level time series derived from radar altimetry were further compared with measurements from two *in-situ* gauge stations (Moc Hoa and Kompong Luong – hereafter Kg Luong, pronunciate Ky Luong - in Figure 3). The Kg Luong station (104.20°, 12.56°), on the Tonle Sap Lake, is located 6 km upstream from an ERS-2/ENVISAT track and the Moc Hoa station (105.93°, 10.77°) is located 2 km away from a T/P track. The satellite and in situ water level time series are respectively presented in Figure 5a and b. Note only these two *in-situ* gauge data sets were available over the study period in close vicinity of the satellite tracks. The RMS difference between *in-situ* and altimetry-derived water levels are respectively 0.23 m and 0.15 m. The results obtained with ERS-2 on the Tonle Sap Lake are similar to those obtained by Mason (1990) with Geosat, by Birkett and Mason (1995) with T/P and those presented by http://www.pecad.fas.usda.gov/ with T/P and Jason-1.

4.3 Analysis of the water level time-series

We have computed water level time series at each of the 80 virtual stations. Since we cannot show all of them, we present on Figure 6 a sample of these time series corresponding to the virtual stations indicated by the yellow stars (from 1 to 12) in Figure 3. Each time series is artificially shifted by 10 meters and error bars are not shown for clarity. The time series numbered 1 and 2 are located on the main river upstream of the intersection with the Tonle Sap River. Both have a mean annual amplitude of about 10 m. The time series numbered 3 is located on the Tonle Sap River. The two time series 4 and 5 located on the Tonle Sap Lake (near the outlet) have a mean annual amplitude in the range 5 -10 m peak to peak. Time series 6 is almost flat. It is located over the lowlands floodplain in Vietnam. Time series 7, 8, 9 also correspond to the lowlands floodplain. Their annual cycle has an amplitude of ~ 2 m. A similar behaviour is noticed for all virtual stations of this area. The remaining three time series of Figure 6 (numbered 10, 11 and 12) are located in the delta, where the main river divides into several branches. All three have very small annual amplitude. We also note that the seasonal signal decreases into the delta where it almost disappears.

In order to get a synthesis of the water level change over the study area, we applied an EOF decomposition to the 80 water level time series (Figure 7). The principal component (temporal curve) shows a minimum water level in year 1998 and a maximum in year 2000. The highest annual amplitudes (up to 10 m) are observed in the northern part of the studied region, upstream on the main river and in the Tonle Sap area. The lowlands floodplains display an annual water level cycle of about 2 m amplitude. Finally the lower delta does not exhibit any seasonal cycle.

5. Interpolated water level maps

5.1 Delineation of the flood extent using SPOT VGT

Among the numerous vegetation indices developed to monitor vegetation structure and activity, the NDVI is the most widely used. Over vegetation cover, NDVI values range in general from 0.3 to 0.8, depending on the vegetation density and photosynthetic activity. Over open water, the NDVI values are generally close to 0. The values of the pixels corresponding to identified water bodies (such as the Tonle Sap lake on Figure 1) are about 0.1. However, NDVI has seldom been used to identify flooded area over land because of the presence in the pixel of partially submerged vegetation (trees, bushes, and high crops), especially in the km resolution pixels of SPOT VGT or NOAA/AVHRR. In the Mekong delta, during the flood season, open water is found in a vast area of agricultural floodplains, dedicated mostly for rice. As an illustration, in Figure 8a shows the time variations of NDVI for a specific region (105.485°<longitude<105.670° and 10.559°<latitude<10.698°) in the lowland floodplain. The NDVI variations correspond to two main crop cycles in the region, with maximum vegetation activity in April-May and December-January. From August to October 2003, low values of NDVI should indicate the presence of open water over the region. Figure 8b shows the water level variations derived from ENVISAT altimetry. Values of the water level in August to October corroborate with the flooding observed by NDVI. An extensive sensitivity analysis indicates that the flooding conditions correspond to NDVI values between 0 and 0.35. A sensitivity analysis was undertaken on the whole study area, in the flood period from August to November. When the threshold in NDVI value changes between 0.15 and 0.25, the inundated area varies by 5%. We assumed that the pixels that correspond to flooded areas have NDVI values lower than 0.2. The remaining uncertainty due to the NDVI threshold is expected to be not important in the analysis of the relative variation of the flooded area.

A second source of error is the presence of clouds which is likely to affect the delineation of the flooded zones in the Mekong basin. Cloud and poor atmospheric conditions generally depress NDVI values (Arino et al., 1992), and as a consequence, some non inundated pixels are incorrectly considered as flooded. Although most of the cloudy pixels were eliminated by the Maximum Value Composite (MVC) approach (Tarpley et al., 1984), a residual cloud contamination has been observed on some VGT-S10 products (Xiao et al., 2002). To reduce the effects of cloud contamination and atmospheric interference in seasonal NDVI time series from daily AVHRR data, Viovy et al. (1992) developed the Best Index Slope Extraction (BISE) technique. It is based on the two following assumptions: (1) NDVI is depressed by cloud and atmospheric contamination, and (2) rapid and non-persistent increases or decreases in NDVI are inconsistent with natural vegetation growth. Our method to discriminate between flooded and non-flooded but cloudy pixels is very similar to this developed by Viovy et al. (1992). We assume that: (1) NDVI is depressed by cloud and atmospheric contamination, (2) an inundated pixel remains inundated during several 10-day intervals (except at flood peak), and (3) during the flooded period (which occurs after the rainy season), pixels are rarely cloudy for two or more consecutive 10-day intervals. Hence, a pixel with a NDVI value lower than 0.2 during only one 10-day interval, is considered cloudy. This method has two main limitations: (1) if clouds are present on two or more consecutive 10-day intervals, the pixel can be wrongly considered inundated, and (2) some pixels on the boundary of the floodplains, inundated during only one 10-day interval can be discarded.

In each VGT-S10 image, we extract all the pixels whose value is finally lower than our threshold to produce 10-day flood maps. The monthly flood map is obtained as the union of the three 10-day flood maps and corresponds to the maximum extent of the flood during the month.

5.3 Monthly maps of interpolated surface water levels

Only one value of water level is available at each virtual station for a given month. The date of the water level measurements varies according to the location of the virtual station. For the ERS-2 period, some data are lacking during one month. We interpolate the water level between two consecutive months to complete the dataset. As a consequence only monthly maps of water level can be estimated over the floodplain. At a given month during the flood, the altimetry-based water levels measured at the virtual stations were linearly interpolated over the flooded zones of the lower Mekong basin. Maps of interpolated surface water levels with 1 km-resolution have been constructed for each month of the flood period, between July and December, for 1998-2003. ERS-2-derived surface water levels are used in the interpolation process for the period 1998-2002 and ENVISAT for 2003. As mentioned above, water levels are expressed with respect to the geoid. For the purpose of hydrological interpretation of water volumes, referring to the topography would be best as far as floodplains are concerned. However, available topographic data bases are not necessarily precise enough for the present study. Altitude of the flooded region delineated by SPOT VGT were derived from elevation points network in the 1/250000 reference map of Vietnam (1986). Figure 9 shows 5 sub zones of homogeneous altitude from 2 to 4 m. The lower Mekong is thus mainly composed of large uniform zones (Figure 9) where only month to month difference maps really represent water volume change (see below).

The monthly maps of water level were produced by bilinear interpolation scheme to estimate the water level for each grid point. South of 11° N, the Mekong river flows preferentially East-West. The ERS-2/ENVISAT tracks cross the river nearly perpendicularly, allowing clear separation of the contributions of the river mainstream from those of associated floodplains (see Figure 3). As a consequence, the interpolation in the along-track direction follows the difference of water levels between the mainstream and the floodplain. In the cross-track direction, interpolation over several tens of kilometers will only reflect the mean slope of the river. Over the northern/southern flowing parts of the river network, the ERS-2/ENVISAT tracks run parallel to the river. In these cases, depending on the choice of the geographical coordinates of the virtual station, the time series can be influenced by the elevation variation within the adjacent floodplain (Birkett et al., 2002). The latter authors did not report obvious amplitude or phase differences due to the inclusion of some floodplain. On the central floodplain, the network of virtual stations is sufficiently dense to avoid important interpolation errors.

The monthly flood maps indicate that the flood period generally ranges from September to November. As an example for year 2003, Figure 10 shows maps of interpolated water levels between July and December 2003. In July and August, limited flood is observed in the Mekong basin. In September, the upstream part of the Mekong basin is flooded. Water levels decrease in the upstream part of the Mekong River whereas they still rise in the Tonle Sap River between September and October. This situation corresponds to the flow reversal of the Tonle Sap River: during the flood, the water flows from the Mekong River to the Tonle Sap River whereas it flows from the Tonle Sap to the Mekong River when the water level decreases. The flood reaches the delta in October. The level of the Tonle Sap River decreases between November and October and the downstream part of the basin is inundated in November. The flood has ended in December and the situation is similar to the month of July.

Figure 11 presents maps of water level differences for two couples of consecutive months (October minus September 2003) and (November minus October 2003). The sequence of events discussed above is even more visible in Figure 12. Important variations in the extent of the flooded zone can be observed from one year to another. This can be shown in Figure 12 which shows maps of interpolated water levels from the month of October for 1998-2003. Year 2000 is characterized by an exceptionally long flood period, from August to December, as was reported by Dutta and Takeuchi (2000) and Chia et al. (2001). The maximum extent is reached in 2000, and years 2001 and 2002 present large floods compared with years 1998, 1999 and 2003.

6. Water volume variations

6.1 Surface water volume change

Determining the temporal variation of water volume stored in the floodplains has many applications in hydrology. For the inundated areas permanently or temporarily connected to main channels, the determination of the water volume variation is equivalent to the estimation of water volume potentially stored and/or released by the valley reach during flood stages. The water volume variation in this type of inundated area is an important parameter for the hydrodynamic modeling of the river flow and the determination of its transport capacity. For the inundated areas that never connect to the main channel, the volume variation is essentially a function of the base flow variation, the inputs from the local basin and the rain. Some floodplains present both types of flooding zones. Areas where the river water mixes with the local water are called "perirheic zones" (Mertes, 1997). In all cases, the inundation area is a buffer zone between the river and the upland watershed. The water volume variation represents the flood pulse of floodplains as expressed by Junk et al. (1989). Therefore, it is a key ecological characteristic that cannot be easily measured in the field.

The variation of water volume corresponds to the surface water levels difference integrated over the inundated surface. These variations $\delta V(t_i,t_{i-1})$, between two consecutive months numbered i and *i-1*, over the floodplain *S* is the sum of the products of the difference of surface water levels $\delta h_j(i,i-1)$, with $j=1, 2, \ldots$ inside *S*, by the elementary surfaces $R_e^2\ \delta\lambda\ \delta\theta\ \sin\theta_j$:

$$\delta V(i,i-1) = R_e^{\,2}\,\delta\lambda\,\delta\theta \sum_{j\in S} \delta h_j(\theta,\lambda,i,i-1)\sin(\theta_j) \qquad (3)$$

where $\delta\lambda$ and $\delta\theta$ are the sampling grid steps along longitude and latitude (~0.01°), respectively and R_e the mean radius of the Earth (~6378 km). The surface water volume variations are expressed in km^3/month.

The error on the method was estimated using:

$$d\delta V = \sum_{i=1}^{n}\left(dS_i\delta h_i + S_i d\delta h_i\right) \qquad (4)$$

where: $d\delta V$ is the error on the water volume variation (δV), S_i is the ith elementary surface, δh_i is the ith elementary water level variation between two consecutive months, dS_i is the error on the ith elementary surface, $d\delta h_i$ is the error on the ith elementary water level variation between two consecutive months.

The maximum error on the volume variation can be estimated as:

$$\Delta(\delta V_{\max}) \le \Delta S \max \delta h_{\max} + S_{\max}\Delta(\delta h_{\max}) \qquad (5)$$

where: $\Delta(\delta V_{max})$ is the maximum error on the water volume variation (δV), S_{max} is the maximum flooded surface, δh_{max} is the maximum water level variation between two consecutive months, ΔS_{max} is the maximum error on the flooded surface, $\Delta(\delta h_{max})$ is the maximum error on the water level variation between two consecutive months.

Using (5), we have estimated the maximum error on the volume change in the lower Mekong basin with the following values :

S_{max} = 45,000 km² in year 2000 (MRC, 2005),

δh_{max} = 2 m,

ΔS_{max} = 5% of 45,000 km²,

$\Delta(\delta h_{max})$ = 0.25 m.

For the whole study zone, we obtained a maximum error of 16 km^3 for the year 2000 for a total positive variation of 32.8 km^3.

Figure 13 presents the differences between two consecutive months of the mean water volume averaged over the study area for 1996-2003. Only months from August to December are considered. Positive variations are obtained between August and October. They are more important for the years 2000, 2001, 2002 than for 1998, 1999 and 2003, with volume variations greater than 15 km^3 and reaching 25 km^3 between September and August 2002. The important flood of 2000 can be related to the exceptional precipitation event (see Figure 2). No similar events can be invoked to explain the great floods of 2001 and 2002.

6.2 Comparison with the total water volume change determined by GRACE

The GRACE mission, launched in March 2002, is devoted to measure spatio-temporal change of the gravity that results mainly from water mass redistribution among the surface fluid envelops (Tapley et al., 2004a). Several recent studies have shown that GRACE data over the continents provide important new information on the total land water storage (surface waters, soil moisture and underground waters, and where appropriate on snow mass) (Tapley et al., 2004b, Wahr et al., 2004, Chen et al., 2005, Schmidt et al., 2005, Ramillien et al., 2005). Wahr et al., (2004) estimated the error on the land water storage to be ~18 mm for a 750-km spatial average GRACE-based land water solutions. Ramillien et al. (2005, 2006) found an error ~15 mm as the final *a posteriori* uncertainties on the land water solutions, with spatial resolution of 660 km. Ramillien et al. (2005) computed monthly change in total water mass volume in several different river basins, including the Mekong. On Figure 10, we present the difference between two consecutive months of the vertically integrated water volume change from GRACE over the whole Mekong basin for the period May 2002 – April 2004. Comparing with monthly differences in surface water volume change computed in the present study, we note that the GRACE-derived total water volume differences display almost in phase seasonal fluctuations. The agreement is particularly good during periods of water falling and around minimum water level. It should be recalled however that GRACE provides the sum of surface waters and soil/underground waters. In addition, due to the current GRACE spatial resolution (~ 600 km), the land water storage has been computed over the whole Mekong basin, an area much larger than that considered in the present study. However this comparison shows that in the future when the GRACE resolution will improve, it could be possible to combine the GRACE result with the water volume change inferred from altimetry data to extract the subsurface water storage change.

7. Conclusion

This study shows that the rather dense coverage of ERS-2/ENVISAT satellites (compared to Topex/Poseidon) over the Mekong floodplains allows us to describe the geographical variations in seasonal high/low water levels. Accuracies of 15 of 23 cm were found for T/P and ERS-2 respectively by comparisons between altimeter-derived water levels and *in-situ* gauge measurements. In particular we noticed a strong difference between the northern part of the Lower Mekong (upstream of the intersection between the main river and the Tonle Sap

River, as well as the Tonle Sap region) and the southern part which includes the main floodplain and delta. The latter exhibits very small seasonal fluctuations (less than 2 m) compared to the former (up to 10 m). This may possibly result from the many canals and dams built during the last two decades in the floodplain area.

This study also demonstrates that combining dense altimetry-based water levels with satellite imagery (here NDVI data from the SPOT-4/Vegetation instrument) provides a new method for remotely measuring surface water volumes over large floodplains. This combination provides valuable information on the dynamics of the inundation of river floodplains. Important interannual variations were observed both in extent and volume over the 1998-2003 period. Knowledge of surface water volumes has several potential applications, for example flood monitoring and forecasting, or sediment transport assessment. In the near future, when GRACE observations will improve in terms of geographical resolution, it will be possible to estimate change in water volumes stored in soil and underground reservoirs by using in synergy GRACE, altimetry and imagery data. Separation of deep and surface land water contributions can also be of interest of global/regional land surface model validation.

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Figure Captions

Figure 1: Map of the Mekong basin. The black triangles represent the *in-situ* gauge stations.

Figure 2: Mode 1 of the EOF decomposition of precipitation from GPCC (Rudolf et al., 1994; 2003) over the lower Mekong, between 1996 and 2003. The upper panel represents the spatial mode whereas the lower panel represents the temporal variation of the precipitation. The percentage of variance explained by the first mode is 82.5%.

Figure 3: Spatial coverage of radar altimeters over the Mekong basin. The red lines correspond to ERS-2/ENVISAT tracks, the light blue dotted lines correspond to T/P tracks. The 80 virtual stations are represented using white dots, the yellow stars correspond to the ones presented on Figure 6. The red circles with a white border represent the *in-situ* gauge stations. The white ellipse corresponds to the intersection between T/P and ERS-2/ENVISAT tracks.

Figure 4: Comparison between T/P and ERS-2/ENVISAT derived water level time-series. Upper panel: water level time series and dispersion derived from T/P measurements, middle panel: water level time series and dispersion derived from ERS-2/ENVISAT measurements (black and grey respectively), lower panel: comparisons of the two water level anomalies (T/P in blue, ERS-2 in black) and residual difference (red curve). The continuous lines are simply joining the dots between the data points. The RMS difference between *in-situ* and altimetry-derived water levels are 0.23 m at Moc Hoa for ERS-2/ENVISAT and 0.15 m at Kg Luong for T/P.

Figure 5: Comparison between altimetry-based and *in-situ* water levels. Upper panel: T/P (red) vs. Moc Hoa gauge station (blue) water levels. Lower panel: ERS-2 (red) vs. Kg Luong gauge station (blue) water levels.

Figure 6: Examples of time-series of water level derived from ERS-2 in the lower Mekong basin (locations of sites 01 trough 12 are shown on Figure 3). Sites 01-02 are main river, 03 is Tonle Sap River, 04-05 are Lake Tonle Sap, 06-09 are floodplain and 10-12 delta.

Figure 7: Mode 1 of the EOF decomposition of water levels derived from radar altimetry over the lower Mekong, between 1996 and 2003. The upper panel represents the spatial mode whereas the lower panel represents the temporal variations of the water levels derived from radar altimetry.

Figure 8: NDVI variations (top) between January 2003 and January 2004 of the region delimited by 105.485E-105.670E and 10.559N-10.698N, and available water level measurements at the virtual station within the region (bottom). The horizontal dashed line on the upper panel represents the threshold on NDVI we used to discriminate between flooded and non-flooded areas. The vertical dashed lines represent the corresponding time span.

Figure 9: Location of the homogeneous altitude zones over the lower Mekong basin (black circles).

Figure 10: Altimetry-based water level maps (wrt GRACE geoid) from July to December 2003.

Figure 11: Altimetry-based water level difference maps: October minus September 2003 and November minus October 2003 (m). The flood reaches the delta in October. The level of the Tonle Sap River decreases between November and October and the downstream part of the basin is inundated in November. The flood ended in December and the situation is similar to the month of July.

Figure 12: Altimetry-based water level maps (wrt GRACE geoid) for October 1998 to 2003 at the maximum of the flood.

Figure 13: Variation of surface water volume from ERS-2/ENVISAT altimetry and SPOT-VGT imagery (green) and total land water volume from GRACE (black).