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IAG EARTH TIDE COMMISSION ON INTERNET

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This paper is to inform you, that the IAG Earth Tide Commission (President: Hans-Georg Wenzel, Vice-President: Shuzo Takemoto, Secretary: Olivier Francis) has established in December 1996 a home page in the world wide web (www) on INTERNET:

<http://www-gik.bau-verm.uni-karlsruhe.de/~iagetc/>

This home page (see appendix for the cover page) contains

- a photograph of the members of the Earth Tide Commission,
- the terms of reference of the Earth Tide Commission,
- the terms of reference and activity reports of the Working Groups under the Earth Tide Commission,
- a page describing the International Center for Earth Tides (ICET),
- the contents of the most recent Bulletin d'Informations Marées Terrestres published by ICET,
- a description of the electronic software service of the Earth Tide Commission established in 1995 (Wenzel 1995),
- information related to the 13th International Symposium on Earth Tides, Bruxelles July 22 - 25, 1997,
- the list of national delegates to the Earth Tide Commission.

If you have any comment to the home page of the Earth Tide Commission, or to other matters related to the Earth Tide Commission, please send me a message. Keep in contact with the Earth Tide Commission !

References:

Wenzel, H.-G. (1996): Electronic information service of the Earth Tide Commission. Bulletin d'Informations Marés Terrestres, **124**, p. 9393, Bruxelles 1996.

This home page has been established in HTML 3.2 at November 28th 1996 by Hans-Georg Wenzel and is still in progress.

Last update: 1996.12.08.

International Association of Geodesy (IAG)

Commission V: Earth Tide Commission (ETC)

President: Hans-Georg Wenzel (Germany)

Vice-President: Shuzo Takemoto (Japan)

Secretary: Olivier Francis (Belgium)



The above photograph (incolour) shows the participants of the 12th International Symposium on Earth Tides, Beijing 1993. The people on the picture can be considered as the members of the Earth Tide Commission in 1993. The graphic card of your PC should be able to display 16 Million colours in order to have a good impression of the photograph.

1. Program of Activities

The objective of the Commission is to promote international cooperation and coordination of investigations related to the observation, preprocessing, analysis and interpretation of earth tides.

By earth tides, we understand all phenomena related to the variation of the Earth's gravity field and to the deformation of the Earth's body induced by the tide generating forces , i.e. the forces acting on the Earth due to differential gravitation of the celestial bodies as the Moon, the Sun and the nearby planets.

Tidal Gravity Recording at Brussels
with a SCINTREX CG-3M Gravity Meter

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Between June 1995 and February 1996 the SCINTREX gravity meter CG-3M n° 9408265 recorded gravity tides at the Royal Observatory of Belgium (ROB) in Brussels (Uccle) (figure 1). Despite the instrumental drift is very large and the resolution rather low, the tidal parameters are very stable. The main interest of this test was that this instrument has a completely independent calibration.

1. Introduction

As described in the instruction manual (SCINTREX LIMITED, 1992) the SCINTREX model CG-3M is a field gravimeter with an improved resolution of 10nms^{-2} ($1\mu\text{gal}$). As the standard CG3 meter it has a worldwide feedback range and is fully computerized. Basically the system makes a measurement each second and compute an average on a given time interval. To improve the accuracy the analog to digital converter is regularly checked against a stable reference voltage during this procedure. The rate recommended by the maker is once every 12s. In good conditions the standard deviation on a single value is $15\mu\text{gal}$. The m.s.e. error on the displayed reading is thus better than $2\mu\text{gal}$ for an integration time larger than 50s. In any case the digital output is rounded up to the nearest microgal.

For each reading the output record gives the gravity reading and the standard deviation of a single measurement in milligal units, the X and Y tiltmeter signals in arc seconds, the temperature compensation signal in milliKelvin, the tidal correction in milligal units, the number of samples, the number of rejected data and the time in hours minutes and second.

Some of the available options can be very useful for tidal recording.

- In automatic cycling mode it is easy to get a value every minute as a mean of more than 50 readings. The data can be transferred automatically to a PC through an RS232 serial port.
- The system is correcting on line slow tilt changes. This option can be very useful in locations with diurnal ground tiltings. It is probably better not to use the continuous mode but to correct only the mean value. It is also possible to recover tilt data in X and Y directions from the digital output for control.
- In noisy environment it is possible to reject automatically values showing a residual larger than 4 times the standard deviation.

Other options are more or less dedicated to field measurements.

- It is possible to correct on line the tidal gravity changes using a fixed algorithm based only on the geographical position.
- It is possible to subtract a fixed value of gravity from the readings.
- It is possible to reduce the apparent drift of the instrument by an automatic drift correction.

* National Fund for Scientific Research

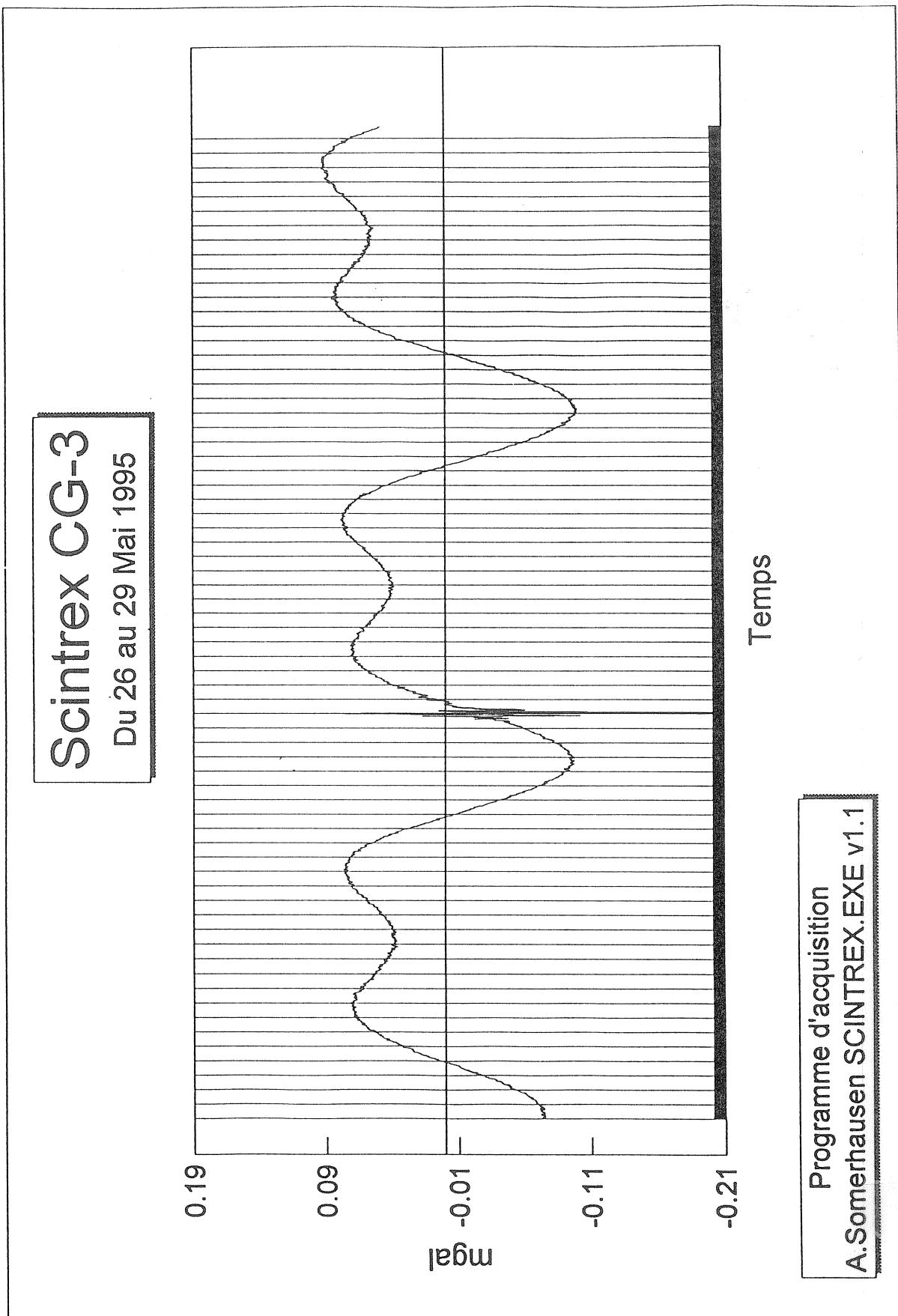


Figure 1 : example of tidal curve; horizontal grid shows the hours.

The temperature compensation signal should be adjusted in each station so that its value is comprised between -1.0mK and $+1.0\text{mK}$. The tilt resolution of $1''$ is excellent as, around the point of minimum of sensitivity to tilting, an error of $1\mu\text{gal}$ corresponds to a tilt of $10''$.

We now describe the procedure of adjustment of the minimum of sensitivity to tilting which is very peculiar as well as the verification of the sensitivity of the tiltmeters.

2. Minimum of Sensitivity to Tilting

As the resolution of the gravity sensor is low, the adjustment of the levels to the position of minimum of sensitivity to tilting is based on large tilts compensated by the automatic correction algorithm.

For an error α_o in radians of the zero reading of one of the tiltmeters, the apparent change of gravity due to tilts in opposite directions will be

$$g_1 = g_o \{ 1 - \frac{1}{2}(\alpha_1 + \alpha_o)^2 \}$$
$$g_2 = g_o \{ 1 - \frac{1}{2}(\alpha_2 + \alpha_o)^2 \}$$

and the difference, neglecting higher order terms in α_o^2

$$g_2 - g_1 = g_o/2((\alpha_1^2 - \alpha_2^2) + 2\alpha_o(\alpha_1 - \alpha_2))$$

The displayed readings will be

$$R_1 = g_1 + C_1$$
$$R_2 = g_2 + C_2$$

where C_1 and C_2 are the computed automatic corrections

$$C_1 = (g_o/2)\alpha_1^2$$
$$C_2 = (g_o/2)\alpha_2^2$$

We thus obtain for the difference of the readings

$$R_2 - R_1 = g_2 - g_1 + (g_o/2)(\alpha_2^2 - \alpha_1^2)$$
$$= g_o\alpha_o/(\alpha_1 - \alpha_2)$$

The unknown zero error of the level adjustment is equal to

$$\alpha_o = 1/g_o(R_2 - R_1)/(\alpha_1 - \alpha_2)$$

Expressing the angles in arc second and g in milligal, we obtain

$$x_o = 4.337 \cdot 10^4 (R_2 - R_1)/(x_1 - x_2)$$

as

$$(206265'')^2/g = 4.337 \cdot 10^4 ''^2/\text{mgal}$$

Expressing the angles in arc second and g in microgal, we obtain

$$x_o = 43.37 (R_2 - R_1)/(x_1 - x_2)$$

To input the correction, the tiltmeter being in the x_2 position, one has only to turn the adjustment potentiometer until the display indicates a new value

$$x'_2 = x_2 + x_0$$

The recommended tilting values are $x_1 = -x_2 = 150''$.
An example of zero adjustment is given in table I

3. Adjustment of the Sensitivity of the Tiltmeters

It is now supposed that the zero of the tiltmeters corresponds to the true vertical position. This procedure allows to determine the calibration factor f of the tiltmeter

$$x = f \cdot V$$

linking the physical units (arc second) to the output voltage of the transducers.

To simplify we convert the tilts expressed in radian α to tilts expressed in arc seconds x by the relation

$$\alpha = x/\beta \quad \text{with } \beta = 206265''$$

Starting with a gravity reading

$$\begin{aligned} R_o &= g_o \\ \text{for} \\ x_o &= 0 \end{aligned}$$

we tilt the instrument by a quantity

$$x_1 = f \cdot V_1$$

and get a reading

$$R_1 = g_1 + g_o \cdot f^2 \cdot V_1^2 / 2\beta^2$$

If the factor is correct $R_1 = R_o$.

If the factor is not correct it is possible to compute the true factor f' such that

$$R_o = g_1 + g_o \cdot f'^2 \cdot V_1^2 / 2\beta^2$$

We then have replacing g_1

$$R_1 = R_o - g_o (f'^2 - f^2) \cdot V_1^2 / 2\beta^2$$

or

$$(g_o / 2\beta^2) (f'^2 - f^2) = (R_o - R_1) / V_1^2 = (g_o / 2\beta^2) \cdot f^2 \cdot (1 - k^2)$$

if we put $k = f'/f$, correcting factor.

Finally we obtain for the correcting factor

$$k = \sqrt{1 + 2(R_o - R_1)\beta^2 / (g_o x_1^2)}$$

Introducing the numerical value of $2\beta^2/g_o$, with g_o in milligal one obtains

$$k = \sqrt{1 + 8.674 \cdot 10^4 (R_o - R_1) / x_1^2}$$

As k is usually close to one we can write directly

$$k = 1 + 4.337 \cdot 10^4 (R_o - R_1) / x_1^2.$$

An example of sensitivity adjustment is given in table 2.

4. Discussion of Errors

If the zero of the tiltmeter was not correctly adjusted we shall have two different values k^+ and k^- corresponding to tilts in opposite directions x^+ and x^- . If $|x^+| = |x^-| = x$, we can write

$$\begin{aligned} k^+ - k^- &= 4.337 \cdot 10^4 (R^- - R^+) / x^2 \\ &= 2x_o / x \end{aligned}$$

We thus have directly a relation linking the error on zero adjustment of the levels to the discrepancy in the correcting factor defined as

$$\epsilon_k = \frac{1}{2}(k^+ - k^-) = x_o / x$$

For a standard $150''$ tilt an offset x_o of $10''$ is thus producing a discrepancy of 0.7% on a single k value.

It is clear however that the mean value $k^M = (k^+ + k^-)/2$ is much more accurate as the errors are symmetrical. For example an error of $+10''$ on the zero adjustment produces an error of $-35\mu\text{gal}$ on the positive direction ($160''$ in place of $150''$) and of $+34\mu\text{gal}$ in the negative direction ($-140''$ in place of $-150''$). The difference on the mean is only $5\mu\text{gal}$. The error on the k^M factor is only of the order of 0.1% . To be sure to obtain an accurate value of the correction factor it is necessary to experimentally determine k^+ and k^- .

On the other hand an error on the calibration factor of the tiltmeter has no influence on the determination of the zero of the tiltmeter if the tilts in opposite directions are equal as we can write

$$kx'_o = x_o = (\beta^2/g_o)(R^- - R^+)/x^+ - \frac{1}{2}(k^2 - 1)(x^+ + x^-)$$

where x^+ and x^- are the displayed tilt value.

The result, apparent zero offset, is itself affected by the calibration error. An error of 10% on β corresponds to an error of $1''$ for $x_o = 10''$.

It is possible to define a common procedure for the simultaneous determination of x_o and k^M by successive readings R_o , R^+ , R^- and R_o , R^+ and R^- corresponding to opposite tiltings with a same absolute value x (see table 3). The drift is eliminated by the second measurement R_o . The values of the unknowns are given by

$$x_o = 4.337 \cdot 10^4 (R^- - R^+)/2x$$

$$k^M = 1 + 4.337 \cdot 10^4 (R_o - R^M) / x^2, \quad R^M = (R^+ + R^-) / 2$$

The readings being rounded up to the nearest μgal the determination of x_o is in principle possible with a precision of $0''.2$ and the value of k^M can be appreciated with an error of 0.3% .

However the adjustment of the zero of the tiltmeters is possible only to $1''$.

An example of such a common adjustment is given in table 3.

5. The data acquisition system

The two main features of this new system would have to be: a recording every minute of all the available parameters coming from the meter; and a graphical display of the last 24 hours of the tidal values. This last feature is essential for maintenance purposes, for it allows a direct assessment of the meter's behaviour and also gives, in a glance, an estimation of the noise present on the tide recordings. For earth tide analysis, the correct dating of the sampled values is of utmost importance, therefore the Scintrex internal real time clock must be compared with the PC's central system clock. Being adjustable by an external time source, the computer's real time clock can be used to evaluate the drift of the Scintrex's internal clock.

When using the *Cycle mode*, the meter will transmit a 77 character long ASCII string containing the following information:

- . Gravity Signal
- . Standard Deviation
- . Tilt X Signal
- . Tilt Y Signal
- . Temperature
- . Earth Tide Correction
- . Duration
- . Rejection
- . Time

The program being intended for use on a portable computer with a gray scale VGA display, the screen output only uses monochrome graphics. Portable computers having in general only one serial port, the program is written so that it expects the data from the meter on the COM1: serial port. This may be easily modified in a future revision of the software.

Furthermore, the speed, the stopbits, the parity and the word length have been set to 1200 bauds, 2 stop bits, No parity and 8 bit words. The Scintrex must therefore be set up in this manner (See Scintrex Instruction book).

When using this software, the computer acts like un dumb-mode receiver, the sample rate for the gravity readings and all other recording parameters must be set up manually by the user.

Incomplete or incorrectly formatted lines are simply ignored by the parsing subroutine and will not impede on the working of the program.

The program was entirely written in Borland C++ 3.0.

5.1 File structure and contents.

For security reasons (in case of power line failure) and convenience for data transmissions, a new data file is created every twelve hours (i.e. midnight and noon).

The filenames themselves contain the date and time of the first reading of the file, So for a file started on December 25th, 1995 at noon, we will have have the following code: 95122512.SCG.

Each file starts with a header composed of several comment-like lines, marked by a pound sign (#) as shown below. This header contains relevant information on the instrument, the location, the operator, the revision of program used, etc...:

```
# INFO: Royal Observatory of Belgium
# INFO: for info André Somerhausen (32) 2 373.03.23
# INST: Scintrex gravity meter SCG-3 [9408265]
# DATE: 951225
# TIME: 082534
# EDAS: SCINTREX.EXE VER 3.0
# CHAN: YYYY NDAY1 HH MI SS GRAV ERROR TX TY TEMP TCORR DUR REJ HH MI SS
# INFO: DATE is PC DATE , TIME is CG-3 TIME
```

Thereafter, the recorded data is structured in the following manner:

- . One record per line
- . PC date and time
- . Literal Scintrex CG-3 string

for example:

```
1995 147 12 59 38 48954.027 000.012 002 001 002.12 0000.000 052 001
```

where:

- 1995 year
- 147 day number within the year
- 12 hour
- 59 minute
- 38 second
- 48954.027 tide value (mgal)
- 000.012 Error (mgal)
- 002 Tilt X (seca)
- 001 Tilt Y (seca)
- 002.12 Internal Temperature (mK)
- 0000.000 Earth Tide Correction (mgal)
- 052 Duration
- 001 Rejection

Note that the Scintrex gives the time but not the date; moreover, the time given by the Scintrex is that of the start of the sampling period as opposed to that of transmission time given by the computer.

5.2 Visual output.

Once the program has been started, the main screen appears.

This screen is subdivided into three independant parts. First, the title section which will only appear if the serial port has been successfully initialized. The second part of the screen, the information bar, will start being active after the first complete reading is sent from the meter. This information bar contains all the parameters sent by the Scintrex's internal software. And finally the third section, the tide curve window gives the last twenty four hours tide as recorded by the Scintrex. The full scale represented on the screen automatically adjusts to the minimum and maximum values recorded over the last 1440 minutes.

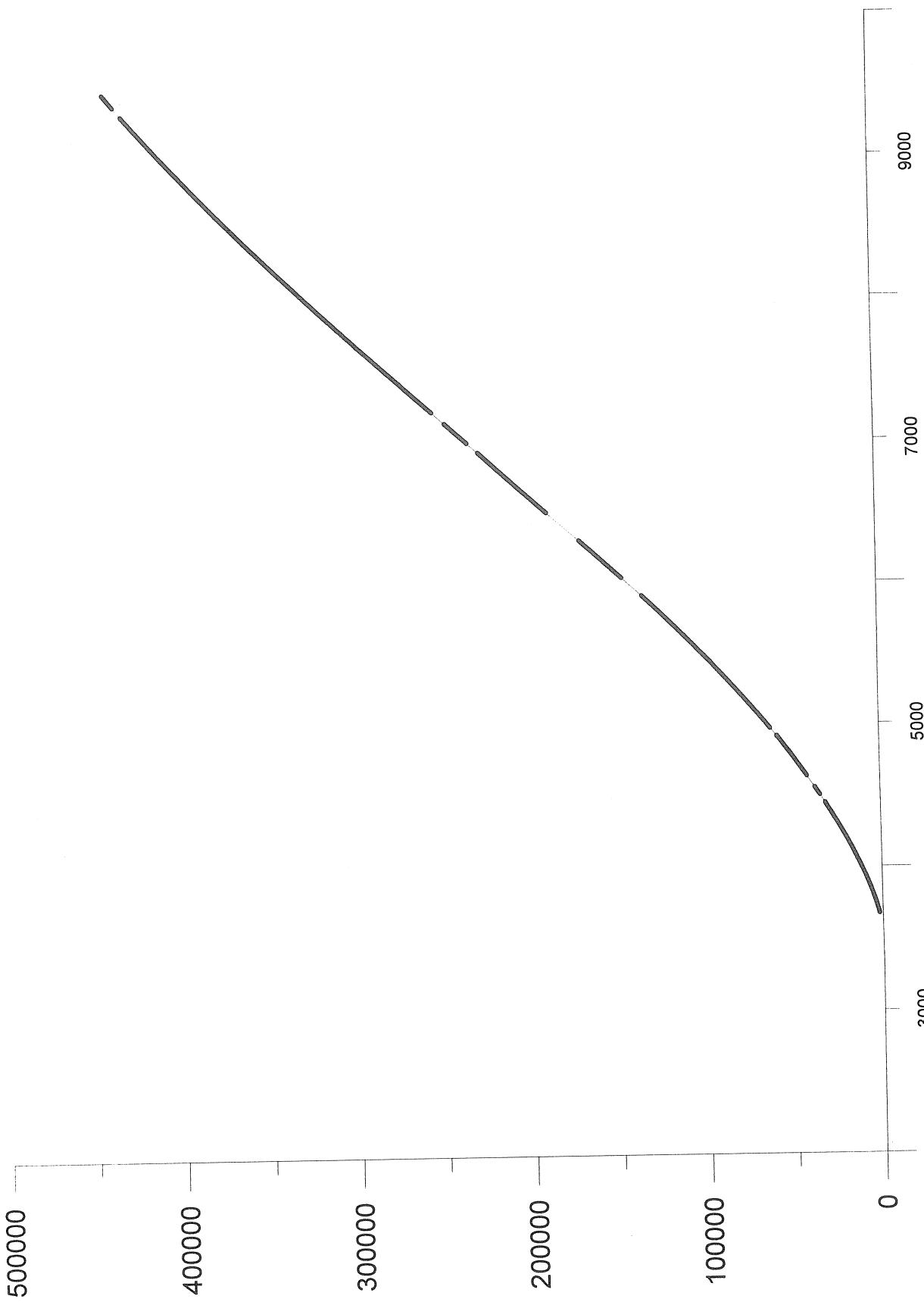


Figure 2 : instrumental drift in nms^{-2} for the period 1995/06/03 to 1996/01/30; horizontal scale is expressed in hours from January 1st, 1995.

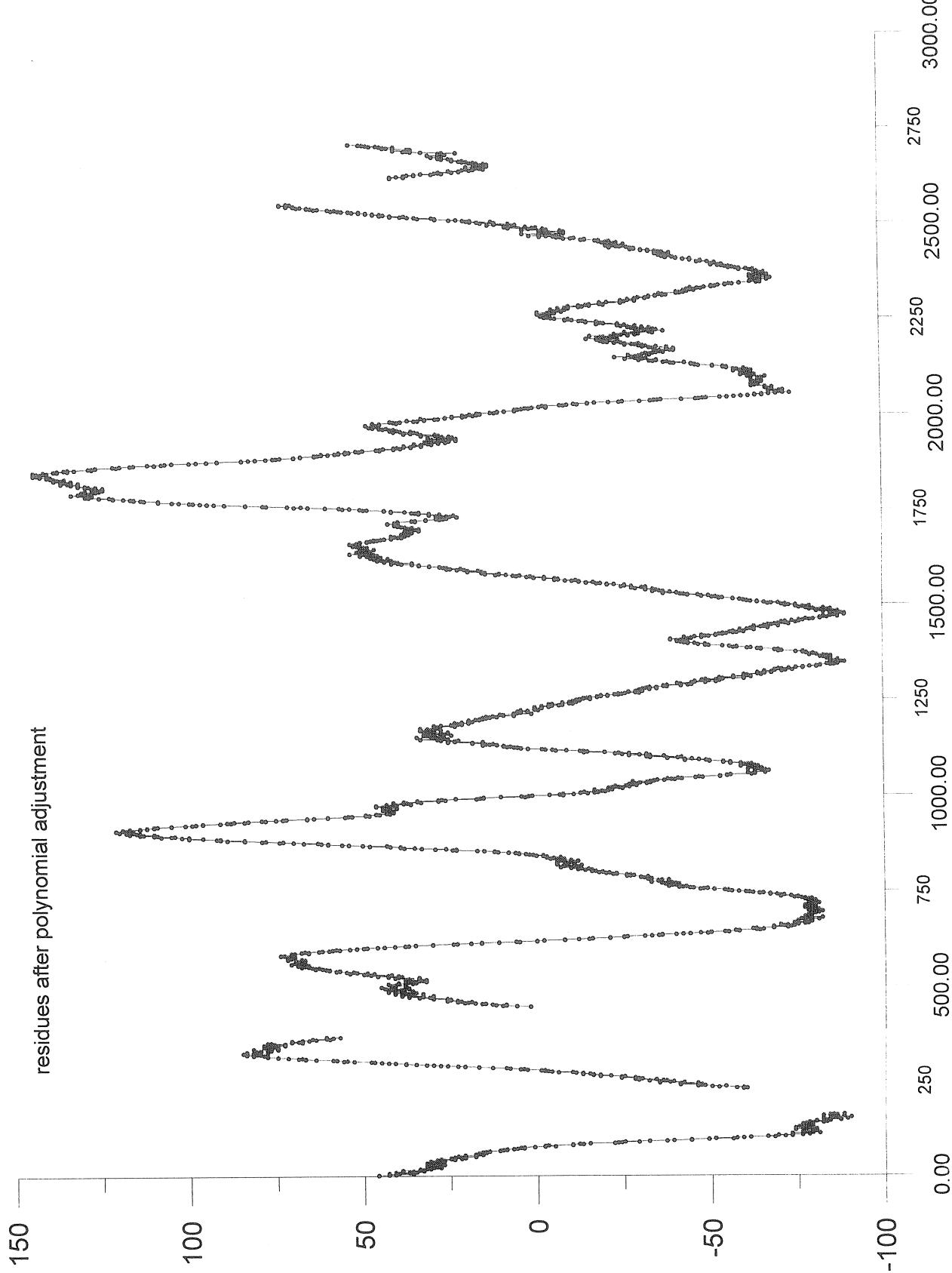


Figure 3 : residues in nms^{-2} after polynomial fit from October 10, 1995 to January 30, 1996.

6. Drift Behaviour

As most of the quartz gravimeters the SCINTREX instruments exhibit a strong drift.

During this initial period of registration this gravimeter showed a parabolic drift becoming quite linear after 3 months (figure 2). In the linear part the drift reached nearly $200 \mu\text{gal}$ per day. The interruption for field measurements after hour 6300 (September 20) did not change the general behaviour. In November an increase of temperature of 5° in the recording room changed the temperature compensation signal of several mK and is reflected in the drift after hour 8000 (November 30). After the elimination of the tidal signal by a Pertsev filter the drift approximation using a polynomial fitting is reliable giving a R.M.S. error on an observation of the order of $5 \mu\text{gal}$. A pressure coefficient was simultaneously computed with a value of $-0.45 \mu\text{gal}/\text{hPa}$ (table V). A graph of the residues after this fit is given in figure 3 for the end of the data.

7. Tidal Gravity Observations

A six month record has been performed on a pillar of the gravimetric room at the Royal Observatory of Belgium. Several interruptions are due to minor failures in the connexion to the PC. From September 22 to 27 recording was interrupted for field measurements.

The stability of the internal clock of the gravimeter is very satisfactory and much better than the stability of the PC clock. All data were recorded in gravimeter time.

As usual we fixed the acquisition cycle to one minute. At the beginning the main problem was that the number of measurements in the mean value plus the internal voltage calibrations exceeded slightly 60s. The time of acquisition was thus drifting away. To avoid that problem we decided later on not to exceed 52 consecutive measurements per minute. It was then impossible to apply standard programs such as PRETERNA on this series and to use the data with drifting acquisition time we had to reinterpolate the series on the second zero using a sliding polynomial fit on 20 consecutive minutes. After that procedure which reduced in a satisfactory way the short period noise on the minute data it was also possible to gain one additional decimal. It was then possible to obtain hourly values expressed in nms^{-2} without additional smoothing.

The first question is to determine the precision of the determination of the tidal parameters. It is well known that it mainly depends on the stability of the sensitivity of the instrument. The resolution of the measurements is only a limiting factor if it is insufficient e.g. less than $0.1 \mu\text{gal}$ on the hourly readings.

As a first test we fit a global tidal factor and phase difference on successive blocks of 48 hours using the best tidal parameters of Brussels. It is the so-called NAKAI procedure (figures 4 and 5). The R.M.S. error of the fit is $1 \mu\text{gal}$. On 113 blocks only one has a R.M.S. error equal to $3 \mu\text{gal}$. The global factor is always close to one. Only 8 blocks give a factor lower than 0.99 or greater than 1.01. The global phase difference is always close to zero. Only 7 blocks deviate of more than ± 0.5 degree. Using the classical Venedikov analysis method we decided to suppress only 2 blocks, one with apparent factor close to 0.97 and one with an internal R.M.S. error of $3 \mu\text{gal}$.

As a conclusion we can state that, even with an higher internal R.M.S. error, the stability of the tidal parameters is comparable with or even better than what is observed with most of the spring gravimeters operated with feedback systems.

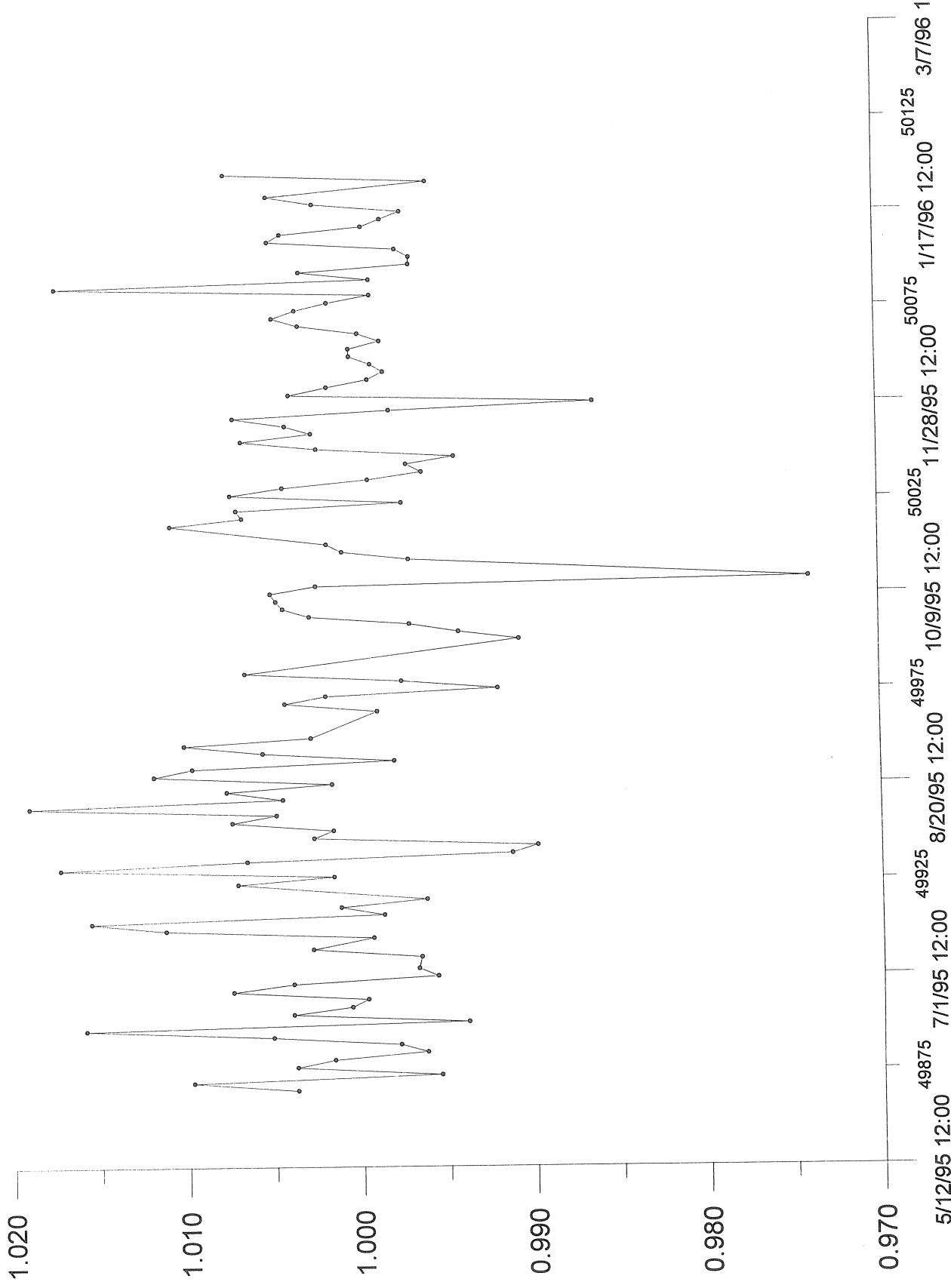


Figure 4 : variation of the global amplitude factors on 48-hour intervals.

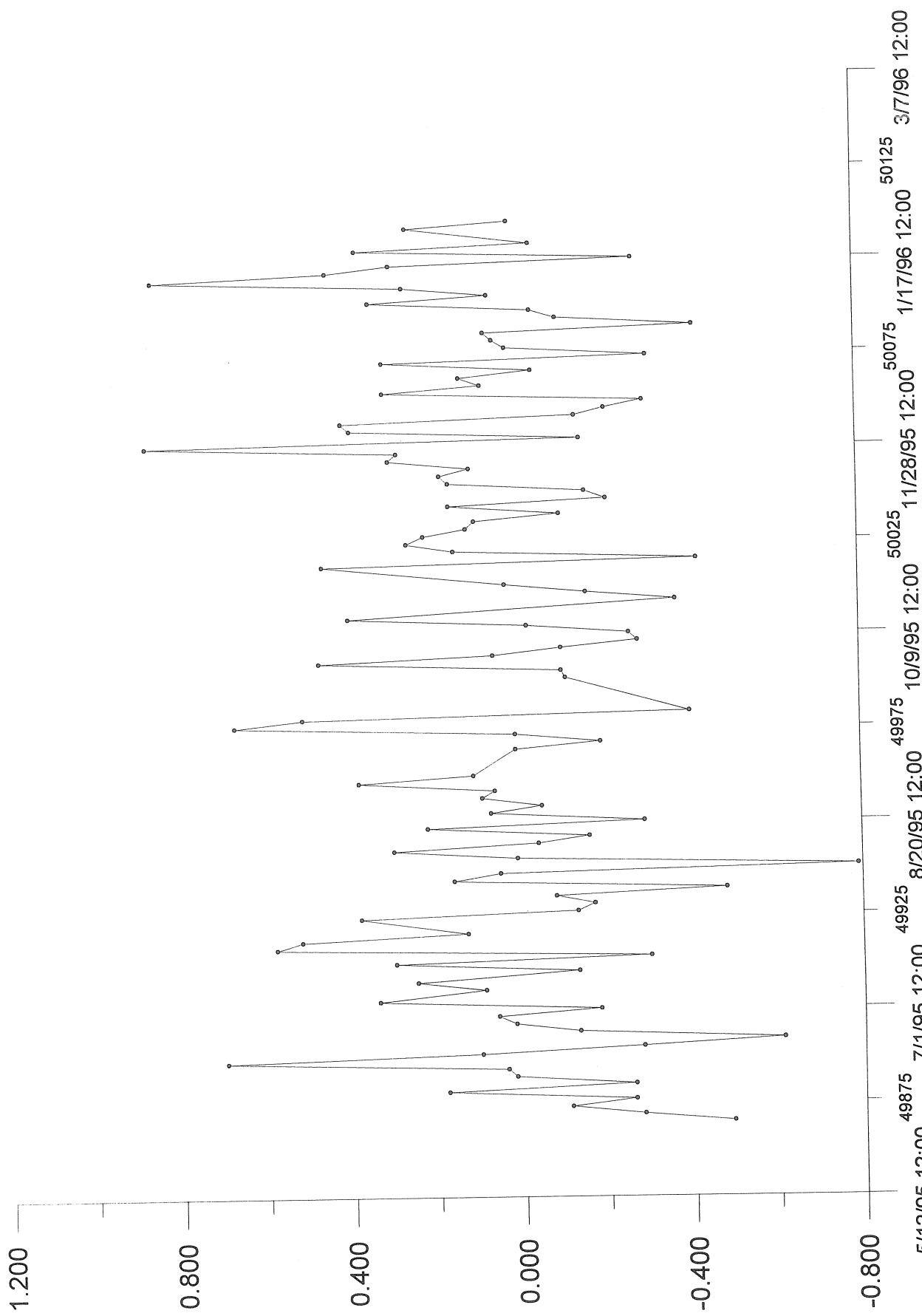


Figure 5 : variations of the global phase differences in degree on 48-hour intervals.

8. Tidal Analysis Results

As a first check we compare the results for O1 and M2 of seven monthly tidal analysis without pressure correction (Table IV). We observe an anomaly on O1 in the fourth analysis. This is confirmed by looking at the results of the global analysis, last line of table IV. As the error on the monthly analysis should be about $\sqrt{7} = 2.5$ times larger than the errors on the global analysis we should have

$$\epsilon_{\delta(O1)} = 0.004, \epsilon_{\alpha(O1)} = 0^\circ.17, \epsilon_{\delta(M2)} = 0.003, \epsilon_{\delta(M2)} = 0^\circ.12.$$

This relation is more or less valid except on the fourth analysis where the errors are twice larger on O1. The computed δ factor on O1 deviate also more than 2.5 times the expected R.M.S. error. This analysis covers the period consecutive to the field measurements from september 22 to 27 1995. On the other hand the change of temperature in the vault in the beginning of november did not affect the tidal factors.

The stability of the results are thus quite satisfactory with very small systematic effects.

An interesting fact is also that the R.M.S. errors in diurnal and semi-diurnal bands are quite similar. In most of the tidal analysis in Brussels there is roughly a factor of two between diurnal and semidiurnal bands. The error in the diurnal band is here greatly reduced and comparable to the error obtained prior to pressure correction with the superconducting gravimeter. It seems that the SCINTREX instrument is less sensitive to noise in the diurnal band.

9. Comparison of Analysis Methods

We give in table V the comparison of the results obtained with the Venedikov VEN66 (Venedikov & Pâquet, 1967) and VEN94 (Venedikov & al., 1995) analysis methods as well as with the ETERNA program (Wenzel, 1996). The main interest of the Venedikov methods based on 48-hour intervals is that it is possible to reject some anomalous data blocks. As already pointed out there are in fact very few bad data in these records.

The VEN94 and ETERNA allow also to determine the transfer function of the pressure. VEN94 gives independent evaluations in the three main tidal bands. The different evaluations are given in table V. The coefficients computed for the ter- and semi-diurnal bands are not significant. It is confirmed by the fact that the pressure correction is only slightly reducing the internal errors in the diurnal band and has no effect on the semi-diurnal one. It seems that below a given limit the noise level of the feedback system becomes the main source of error. The global evaluation is similar to the coefficient in the diurnal band and does not differ significantly from $-0.3\mu\text{gal}/\text{hPa}$.

Table VI gives a comparison of the tidal factors for the main waves. The results obtained by different analysis procedures are coherent within the associated R.M.S. errors. M3 is not significant. It is due to the increased noise level in the ter-diurnal band well detected by ETERNA (table VIII). Here again it is linked to the poor resolution of the system.

The complete analysis results are given for VEN66 (table VII) and ETERNA (table VIII).

10. Comparison between different instruments

Brussels being the reference station of the Transworld Tidal Gravity Profiles (TWP) a complete discussion of the normalisation has been presented in Melchior, 1994 and applied to the data of the superconducting gravimeter GWR T03. In the first column of table IX we reproduce the results of this instrument as published in Melchior & al., 1996. Besides the results of the SCINTREX CG3-M instrument we present a reanalysis of the observations made by Edge R.J. and Jeffries G. in 1985 with the improved LaCoste & Romberg ET13 gravimeter (Baker & al., 1991) and of two LaCoste & Romberg model G gravimeters transformed in zero method by Van Ruymbeke (1989a, 1995). The interest of the comparison is that the calibration systems are different. The CG3-M uses the original maker calibration, the ET13 has been recalibrated on the Hannover vertical baseline (Kangieser & Torge, 1981) and the sensitivity of LCZ3 and LCZ336 were determined on two different versions of the calibration platform developed at ROB (Van Ruymbeke M., 1986 & 1989b).

For an easier comparison we give for each wave in table X the ratio between the amplitude factor of T03 and the corresponding factors of the other instruments. Trying to derive a mean factor we computed weighted means based on the wave amplitudes. The R.M.S. error on the unit weight is given under the assumption that the errors on T03 results are an order of magnitude lower than the errors of the other instruments.

The most coherent factors are those of CG3-M and ET13 but there is a systematic difference of 2.10^{-3} between the two instruments in the diurnal as well as in the semi-diurnal band. The two model G gravimeters agree within 0.1% with ET13 but show internal discrepancies. In fact there is a good agreement with CG3-M for O1 and with ET13 for the other waves. Looking to the amplitude ratios in table 9 there is an excellent coherence for M2/K1 but not for M2/O1 and O1/K1. It clearly points out to an anomaly on O1 for LCZ3 and LCZ336. The results of these instruments have thus to be considered as mere indication. We found back here the results already presented in Ducarme & Van Ruymbeke, 1991 and discussed in Melchior, 1994 that the Brussels normalisation had to be reduced by 1.2%. Melchior taking into account other intercomparisons preferred to reduce only this normalisation by 0.8%. The results of the SCINTREX instrument points to a reduction by 1.0%.

As a conclusion we can say that the indetermination on the tidal factors at Brussels is still of the order of 0.2% and that the results of the SCINTREX CG3-M is fitting within this uncertainty.

Concerning the phase determination the CG3-M is very close to ET13. In any case the discrepancies are of the order of magnitude of the R.M.S. errors.

11. Conclusions

The SCINTREX CG3-M gravimeters can be easily adapted for tidal gravity records. The adjustment for minimum of sensitivity to tilting as well as the sensitivity of the built in tiltmeters can be easily checked following the maker's instructions.

The SCINTREX CG3-M gravity meter seems to have several interesting features for tidal records :

- It gives very reliable tidal factors for the diurnal and semi-diurnal waves. The dispersion of the monthly analysis results corresponds to the associated R.M.S. errors. Its sensitivity seems perfectly stable.
- Its calibration fits at least within $\pm 0.2\%$ with the other calibration systems.
- In the diurnal band it has internal errors similar to a tidal LaCoste Romberg instrument. It is slightly less satisfactory in the semi-diurnal band.

On a six month recording period the R.M.S. error is 0.1% on K1 and M2, corresponding to $0.04\mu\text{gal}$ resolution.

The high drift rate being very regular is not a limiting factor for tidal analysis.

Its limitation seems to come from its high short term noise which reaches $2\mu\text{gal}$ on minute samplings. Due to this fact the instrument has a poor resolution, around $0.2\ \mu\text{gal}$ on hourly means. The effect of this high noise is already noticeable in the ter-diurnal band which is unreliable. It is well known that Brussels station has a high level of industrial noise. We are now testing the instrument in the underground station of Membach in almost perfect conditions of stability to see if the precision will be better.

Due to its very good rejection of noise in the diurnal band, its very stable sensitivity and the automatic correction for tilt, this instrument seems to be highly suitable for meteorologically perturbed stations.

We believe that for high resolution measurements it could be usefully associated with a more sensitive instrument to help detecting anomalous records of this last one.

BIBLIOGRAPHY

- Baker T.F., Edge R.J., Jeffries G., 1991.
European Tidal Gravity.
Proc. eleventh Int. Symp. on Earth Tides, Helsinki, July 31 - August 5, 1989. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart : 147-155.
- CG-3/3M Gravity Meter User's guide, 1992.
SCINTREX ltd., 222 Snidercroft Road, Concord, Ontario, L4K-1B5, CDM.
- Kangieser E., Torge W., 1981.
Calibration of LaCoste Romberg gravity meters, model G and D.
Bull. d'Inf. Bur. Grav. Int., 49 : 50-63.
- Ducarme B., Van Ruymbeke M., 1991.
On tidal parameters at Brussels fundamental station.
Proc. eleventh Int. Symp. on Earth Tides, Helsinki, July 31 - August 5, 1989. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart : 157-166.
- Melchior P., 1994.
A new data bank for tidal gravity measurements (DB92).
Physics Earth Planet. Interiors, 82 : 125-155.
- Melchior P., Ducarme B., Francis O., 1996.
The response of the Earth to tidal body forces described by second-and third-degree spherical harmonics as derived from a 12 year series of measurements with the superconducting gravimeter GWR T3 in Brussels.
Physics Earth Planet. Interiors, 93 : 223-238.
- Van Ruymbeke M., 1986.
Calibration of LaCoste Romberg gravimeters by an inertial force resulting from a vertical periodic movement.
Proc. tenth Int. Symp. on Earth Tides, Madrid, sept. 23-28, 1985. Consejo Superior Invest. Cientificas : 35-42.
- Van Ruymbeke M., 1989a.
New developments of feedback electronics.
Working Group on High Precision Tidal Data Processing, Bonn, october 4-6, 1988.
Bull. Inf. Marées Terrestres, 104 : 7292-7298.
- Van Ruymbeke M., 1989b.
Improvement of the vertical oscillating platform for the calibration of gravimeters.
Working Group on High Precision Tidal Data Processing, Bonn, october 4-6, 1988.
Bull. Inf. Marées Terrestres, 104 : 7290-7291.
- Van Ruymbeke M., Somerhausen A., Blanchot G., Claes A., Grammatica N., 1995.
New developments with gravimeters.
Proc. twelfth Int. Symp. on Earth Tides, Beijing, August 4-7, 1993. Science Press, Beijing - New York : 89-101.

Venedikov A.P., Pâquet P., 1967.

Sur l'application d'une méthode d'analyse des marées terrestres à partir d'enregistrements de longueur arbitraire.
Bull. Inf. Marées Terrestres : 48, 2090-2114.

Venedikov A.P., Vieira R., de Toro C., 1995.

The computer program MSV used in Madrid for tidal data processing.
Meeting of the Working Group on High Precision Tidal Data Processing,
Bonn, August 30 - Sept. 2, 1994. Bull. Inf. Marées Terrestres, 121,
9108-9126.

Wenzel H-G., 1996.

The nanogal software : Earth tide data processing package : ETERNA 3.3.
Bull. Inf. Marées Terrestres, 124, 9425-9439.

TABLE I

Tilt sensors zero adjustment

96/02/01 integration on 120 measurements

16h 00m $\chi_1 = 149$ $R_1 = 46.433$
16h 07m $\chi_2 = -149$ $R_2 = 46.395$

$$x_0 = -0.038/298 \times 4.337 \times 10^4 = -5''.5$$

16h 20m $y_1 = 149$ $R_1 = 46.367$
 $y_2 = -149$ $R_2 = 46.457$

$$y_0 = 0.090/298 \times 4.337 \times 10^4 = +13''.1$$

TABLE II

Adjustment of tilt sensitivity

96/02/02 integration on 120 measurements

	x	y	R	R _c
13h 32m	1	1	52.764	52.7640
40m	151	1	52.768	52.7670
48m	1	1	52.766	52.7640
59m	1	151	52.771	52.7675
14h 09m	1	1	52.769	52.7640

R_c are readings corrected for a linear drift

$$\begin{aligned} \text{tilt } x & \quad k = 1 - 0.0058 = 0.9942 \\ \text{tilt } y & \quad k = 1 - 0.0068 = 0.9932 \end{aligned}$$

No adjustment needed.

TABLE III

Common adjustment of zero and sensitivity

96/02/05 integration on 120 measurements

	x	y	R	drift	R _c	ΔR
07h 28m	0	1	53.186	0	53.186	.0005
36m	150	1	53.185		53.1865	.0010
44m	-150	1	53.185		53.1875	-.0015
52m	0	1	53.182	-4	53.186	.0
08h 00m	0	150	53.180		53.186	.0
08m	0	-150	53.178		53.186	.0
16m	0	1	53.176	-10	53.186	

$$\begin{aligned} \text{tilt } X & \quad x_0 = 4.337 \cdot 10^4 \times 0.001/300 = 0.14'' \\ & \quad k^M = 1 - 4.337 \cdot 10^4 \times 0.001/22500 = 0.998 \end{aligned}$$

tilt Y no correction

TABLE IV
Monthly Tidal Analysis *

	01	M.S.E.	M2	M.S.E.		
	δ	α	D	δ	α	SD
1	1.152 ± 0.004	0.45 ± 0.19	1.65	1.178 ± 0.003	3.03 ± 0.16	1.65
2	1.151 ± 0.004	0.24 ± 0.19	1.51	1.181 ± 0.003	2.94 ± 0.13	1.23
3	1.152 ± 0.003	0.30 ± 0.16	1.36	1.184 ± 0.003	2.72 ± 0.12	1.31
4	[1.140] ± 0.007	0.34 ± 0.33	1.49	1.179 ± 0.003	3.00 ± 0.14	0.81
5	1.148 ± 0.003	0.06 ± 0.14	1.12	1.185 ± 0.003	2.85 ± 0.15	1.43
6	1.151 ± 0.004	0.28 ± 0.22	1.93	1.181 ± 0.004	2.82 ± 0.11	1.17
7	1.150 ± 0.003	0.28 ± 0.16	1.43	1.182 ± 0.003	2.61 ± 0.14	1.51
global	1.1498 ± 0.0015	0.292 ± 0.074	1.633	1.1819 ± 0.0009	2.826 ± 0.067	1.303

* VEN66 without air pressure correction.

TABLE V
Evaluation of Pressure Coefficients ($\mu\text{gal}/\text{hPa}$)

Method	ETERNA	VEN94	Drift study
tidal band	- 0.32 ± 0.03		
diurnal		- 0.29 ± 0.05	
semi-diurnal		0.03 ± 0.18	
ter-diurnal		- 0.20 ± 0.19	
long period			- 0.45 ± 0.02

TABLE VI
Comparison of the results obtained with
different analysis methods

a) Amplitude factors

	ETERNA	VEN66*	VEN94+	VEN94\$
O1	1.1514 ± 0.0013	1.1503 ± 0.0014	1.1508 ± 0.0013	1.1510 ± 0.0014
P1	1.1459 ± 0.0023	1.1464 ± 0.0024	1.1473 ± 0.0023	1.1485 ± 0.0025
K1 S1	1.1396 ± 0.0009	1.1393 ± 0.0009	1.1386 ± 0.0009	1.1388 ± 0.0009
N2	1.1647 ± 0.0043	1.1658 ± 0.0051	1.1574 ± 0.0062	1.1593 ± 0.0061
M2	1.1819 ± 0.0008	1.1816 ± 0.0009	1.1820 ± 0.0009	1.1826 ± 0.0009
S2	1.1938 ± 0.0019	1.1915 ± 0.0021	1.1963 ± 0.0032	1.1964 ± 0.0031
M3	1.020 ± 0.084	1.074 ± 0.068	0.983 ± 0.076	1.030 ± 0.083

b) Phase differences

	ETERNA	VEN66*	VEN94+	VEN94\$
O1	0.200 ± 0.072	0.248 ± 0.068	0.251 ± 0.065	0.244 ± 0.068
P1	0.451 ± 0.132	0.374 ± 0.120	0.485 ± 0.115	0.461 ± 0.124
K1 S1	0.311 ± 0.049	0.279 ± 0.045	0.292 ± 0.044	0.266 ± 0.047
N2	3.812 ± 0.249	3.940 ± 0.251	3.942 ± 0.314	3.866 ± 0.310
M2	2.864 ± 0.045	2.828 ± 0.046	2.824 ± 0.045	2.796 ± 0.044
S2	1.248 ± 0.933	1.204 ± 0.104	1.430 ± 0.187	1.464 ± 0.186
M3	2.33 ± 4.81	5.58 ± 3.58	0.84 ± 4.38	4.74 ± 4.53

* 2 bloks rejected, pressure coefficient - 0.3 $\mu\text{gal}/\text{hPa}$

+ weighted intervals

\$ 3 bloks rejected with $\sigma \geq 2.560$

TABLE VII

Tidal Analysis Results with VEN66

STATION 0201 BRUXELLES-UCCLE COMPOSANTE VERTICALE BELGIQUE

50 47 55 N 04 21 29 E H 101 M P 4M D 90KM 981 117 301
BASSIN EOCENE DE BRUXELLES SUR LE CAMBRIEN DU MASSIF DU BRABANT.

SABLES LUTETIENS.

OBSERVATOIRE ROYAL DE BELGIQUE DEPT.1 R. VERBEIREN

GRAVIMETRE SCINTREX CG3-M 9408265

CORRECTION DE PRESSION -0.3microgal/hPa

CALIBRATION CONSTRUCTEUR

INSTALLATION B.DUCARME, A. SOMERHAUSEN

MAINTENANCE A. SOMERHAUSEN, R. LAURENT

LEAST SQUARE ANALYSIS / VENEDIKOV FILTERS ON 48 HOURS / PROGRAMMING B.DUCARME
POTENTIAL TAMURA COMPLETE DEVELOPMENT
COMPUTING CENTER INTERNATIONAL CENTER FOR EARTH TIDES/FAGS/ BRUSSELS
DATA PROCESSING BY B.DUCARME ON 96/04/18
COMPUTER HP-UX 9000/819
INERTIAL CORRECTION NOT APPLIED

G265 95 6 3/95 7 9 95 713/95 723 95 727/95 9 3 95 9 9/95 919 95 929/951013
G265 951019/951023 951030/96 122 96 127/96 129

TIME INTERVAL 242.0 DAYS 5136 READINGS 8 BLOKS EFFICIENCY 0.88

WAVE GROUP ARGUMENT	ESTIMATED AMPL. N WAVE	AMPL. R.M.S.	AMPL. FACTOR	PHASE R.M.S.	RESIDUE R.M.S.	AMPL. R.M.S.	PHASE
133.-136. 33 Q1	6.6530.043	1.14154	0.00745	-0.7490	0.3738	0.109	-127.35
142.-145. 32 O1	35.0150.041	1.15034	0.00136	0.2482	0.0677	0.170	116.66
152.-155. 30 NO1	2.7210.037	1.13656	0.01551	0.5661	0.7840	0.047	144.78
160.-163. 17 P1	16.2360.034	1.14639	0.00242	0.3735	0.1203	0.107	99.95
164.-169. 39 S1K1	48.7740.039	1.13934	0.00091	0.2789	0.0453	0.379	38.75
174.-177. 25 J1	2.7660.043	1.15549	0.01806	-0.3533	0.8959	0.017	-89.20
184.-187. 23 O01	1.6260.079	1.24170	0.06064	1.5329	2.7947	0.121	21.02
233.-23A. 44 2N2	1.0100.022	1.10062	0.02449	4.3414	1.2755	0.096	127.00
242.-249. 50 N2	6.7010.029	1.16582	0.00511	3.9401	0.2508	0.461	88.07
252.-259. 56 M2	35.4750.028	1.18162	0.00094	2.8275	0.0457	1.847	71.34
264.-268. 32 L2	1.0100.017	1.19053	0.02043	3.8661	0.9865	0.072	71.30
270.-273. 16 S2	16.6430.030	1.19150	0.00212	1.2037	0.1038	0.551	39.41
274.-278. 28 K2	16.6680.140	1.19335	0.00999	1.2238	0.4800	0.575	38.28
315.-3A5. 82*M3	0.4040.025	1.08218	0.06776	5.5765	3.5831	0.039	87.54
STANDARD DEVIATION	D 1.499	SD 1.305	TD	1.176	MICROGAL		

QUALITY FACTORS : Q1= 12.6 Q2= 25.9
O1/K1 1.0097 1-O1/1-K1 1.0789 M2/O1 1.0272
CENTRAL EPOCH TJJ= 2449991.5

TABLE VIII

Tidal Analysis Results with ETERNA

```
# STATION 0201 BRUXELLES-UCCLE      COMPOSANTE VERTICALE      #
# 50 47 55 N   04 21 29 E   H 101 M   P   4M   D   90KM      #
# BASSIN EOCENE DE BRUXELLES SUR LE CAMBRIEN DU MASSIF DU BRABANT      #
# SABLES LUTETIENS.      #
# OBSERVATOIRE ROYAL DE BELGIQUE      DEPT. 1      R. VERBEIREN      #
# GRAVIMETRE SCINTREX CG3-M 9408265      #
# ENREGISTREMENT DIGITAL: MOYENNE SUR 60SEC.      #
# CORRECTION 30S SUR LE TEMPS      #
# CALIBRATION      FABRIQUANT      #
# INSTALLATION      A. SOMERHAUSEN, B. DUCARME      #
# MAINTENANCE      B. DUCARME, R. LAURENT      #
#####
```

Latitude: 50.7990 deg, longitude: 4.3580 deg, azimuth: .000 deg.

Summary of observation data :

19950602	0...19950705230000	19950706120000...19950710110000
19950712	0...19950725110000	19950726 0...19950904230000
19950908120000...19950921110000	19950928 0...19951017110000	
19951018120000...19951025230000	19951027120000...19951230230000	

Initial epoch for tidal force : 1995. 6. 2. 0

Number of recorded days in total : 197.00

TAMURA 1987 tidal potential used.

WAHR-DEHANT-ZSCHAU inelastic Earth model used.

UNITY window used for least squares adjustment.

Numerical filter is PERTSEV 1959 with 51 coefficients.

Estimation of noise by FOURIER-spectrum of residuals

0.1 cpd band	99999.9990 nm/s**2	1.0 cpd band	.5481 nm/s**2
2.0 cpd band	.3983 nm/s**2	3.0 cpd band	.5703 nm/s**2
4.0 cpd band	.4026 nm/s**2	white noise	.3611 nm/s**2

adjusted tidal parameters :

from [cpd]	to [cpd]	wave [nm/s**2]	ampl. ampl.fac.	stdv. [deg]	ph. [deg]	lead [deg]	stdv. [deg]
.721500	.906315	Q1	66.718	1.14481	.00676	-.7304	.3887
.921941	.940487	O1	350.494	1.15147	.00131	.0685	.0757
.958085	.974188	M1	27.685	1.15650	.01489	1.0290	.8537
.989049	1.011099	K1	487.150	1.13797	.00078	.2017	.0444
1.013689	1.044800	J1	27.542	1.15054	.01715	-.2055	.9844
1.064841	1.216397	OO1	15.857	1.21058	.05707	-.1428	3.2711
1.719381	1.872142	2N2	10.393	1.13222	.02063	5.4044	1.1810
1.888387	1.906462	N2	66.946	1.16471	.00430	3.5752	.2467
1.923766	1.942754	M2	354.797	1.18182	.00079	2.6244	.0450
1.958233	1.976926	L2	10.166	1.19806	.01615	2.3886	.9260
1.991787	2.182843	S2	166.695	1.19345	.00184	.9772	.1085
2.753244	3.081254	M3	3.809	1.01980	.08368	1.9759	4.7900

Adjusted meteorological or hydrological parameters:

no.	regr.coeff.	stdv.	parameter	unit
1	-3.41518	.29425	airpress.	nm/s**2 / hPa

Standard deviation of weight unit: 10.230 nm/s**2
degree of freedom: 4301

TABLE IX

Comparison of the results obtained with
different gravimeters

a) Amplitude factors

	GWR T03*	CG3M	ET13	LCZ3	LCZ336
	4489 days	242 days	128 days	372 days	317 days
O1	1.15303 ± 0.00008	1.1503 ± 0.0014	1.1455 ± 0.0014	1.1504 ± 0.0030	1.1496 ± 0.0017
P1	1.15213 ± 0.00019	1.1464 ± 0.0025	-	[1.1660] [± 0.0076]	1.1462 ± 0.0043
K1	1.13981 ± 0.00006	1.1393 ± 0.0009	1.1368 ± 0.0009	1.1326 ± 0.0022	1.1328 ± 0.0013
N2	1.16779 ± 0.00030	1.1658 ± 0.0051	1.1691 ± 0.0040	1.1750 ± 0.0064	1.1564 ± 0.0064
M2	1.18359 ± 0.00006	1.1816 ± 0.0009	1.1772 ± 0.0007	1.1766 ± 0.0012	1.1779 ± 0.0010
S2	1.19603 ± 0.00011	1.1915 ± 0.0021	1.1924 ± 0.0014	1.2133 ± 0.0024	1.1856 ± 0.0021
M2/O1	1.0265	1.027	1.028	1.023	1.025
O1/K1	1.0116	1.010	1.007	1.016	1.015
M2/K1	1.0384	1.037	1.038	1.039	1.040

b) Phase differences

	GWR T03	CG3M	ET13	LCZ3	LCZ336
	4489 days	242 days	128 days	372 days	317 days
O1	0.0689 ± 0.0041	0.248 ± 0.067	0.125 ± 0.068	0.022 ± 0.148	0.052 ± 0.085
P1	0.1633 ± 0.0094	0.374 ± 0.120	-	0.614 ± 0.372	0.639 ± 0.214
K1	0.2529 ± 0.0030	0.279 ± 0.045	0.312 ± 0.044	0.216 ± 0.113	0.497 ± 0.064
N2	3.3001 ± 0.0149	3.940 ± 0.251	3.782 ± 0.195	2.823 ± 0.314	3.659 ± 0.318
M2	2.7411 ± 0.0027	2.828 ± 0.045	2.862 ± 0.034	2.644 ± 0.056	2.636 ± 0.051
S2	1.0835 ± 0.0056	1.204 ± 0.104	1.481 ± 0.071	2.506 ± 0.115	1.220 ± 0.104

* normalisation based on ICET Data Bank DB92 (Melchior 1994)

TABLE X

Amplitude ratio of GWR T03 with respect to
the other gravimeters

Weight		CG3M	ET13	LCZ3	LCZ336
1.0	O1	1.0024	1.0066	1.0023	1.0030
0.5	P1	1.0050			1.0052
1.5	K1	1.0004	1.0027	1.0064	1.0062
3.0(2.5)	D	1.0018	1.0043	1.0048	1.0050
	ϵ_D	0.0015	0.0014	0.0030	0.0017
1.0	M2	1.0017	1.0054	1.0060	1.0048
0.5	S2	1.0038	1.0031	1.0031	1.0089
1.5	SD	1.0024	1.0046	1.0050	1.0062
	ϵ_{SD}	0.0009	0.0007	0.0012	0.0010
4.5(4.0)		1.0020	1.0044	1.0049	1.0054

ϵ : R.M.S. error on unit weight

ON THE DETERMINATION OF THE D AND SD EARTH TIDES GENERATED
BY THE TIDAL POTENTIAL OF THE THIRD ORDER

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A problem of the analysis of Earth tide data is the D and SD waves, which we shall denote W_3 , generated by the tidal potential of order 3. A successful separation and determination of W_3 is given in the paper (Melchior, Ducarme, Francis, 1995), further referred as MDF. The only preceding known attempt to obtain W_3 belongs to Dittfeld (1991).

In the present remark are given the results about W_3 we have obtained using nearly the same data as MDF but a somewhat different way of analysis.

The analysis applied by MDF is based on the formation of tidal groups according to (Venedikov, 1966). We have series of frequency intervals (ω_j, ω_{j+1}) defined by a sequence of frequencies $\omega_1 < \omega_2 < \dots < \omega_n$. All tides with frequencies ω in (ω_j, ω_{j+1}) shape the j th group. It is accepted that the tides of one and the same group have one and the same amplitude factor δ and phase shift κ because ω_j and ω_{j+1} are relatively close.

In MDF the intervals (ω_j, ω_{j+1}) are defined in such a way that there are created several groups composed by waves of type W_3 .

We have used, in principle, the same method of analysis, applied by the program NSV (Venedikov, Vieira, de Toro, 1994). An option has been added which is allowing a more general definition of the tidal groups.

We keep the usual (ω_j, ω_{j+1}) . However, all diurnal and semidiurnal W_3 are taken away from (ω_j, ω_{j+1}) , creating two new groups named 3D (diurnal W_3) and 3SD (semidiurnal W_3). In such a way 3D, as well as 3SD, are not defined by intervals (ω_j, ω_{j+1}) with very close ω_j and ω_{j+1} . Nevertheless, there are not theoretical reasons to expect important variations of δ and κ within 3D or 3SD and we can accept that δ and κ remain constant within these groups. By the way, this supposition is confirmed by MDF.

In Table 1 are given the results obtained by MDF, as well as by our technique called "a new determination".

References

- Dittfeld, H.G., 1991. Analysis of third degree waves with diurnal and semidiurnal frequencies. Bull. Inf. Marées Terrestres, 111: 8053-8061.
- Melchior, P., Ducarme, B. and Francis, O., 1995. The response of the Earth to tidal body forces described by second- and third- degree spherical harmonics as derived from a 12 year series of measurements with the superconducting gravimeter GWR/T3 in Brussels. Phys. Earth Planet. Inter. (under press).
- Venedikov, A.P., 1966. Une méthode d'analyse des Mares terrestres partant d'enregistrements de longueur arbitraires. Acad. Royal Belg., Bull. Cl. Sci., t. LIII, fasc. 3: 463-485.
- Venedikov, A.P., Vieira, R. and de Toro, C., 1995. The computer program NSV used in Madrid for tidal data processing. Bull. Inf. Mares Terrestres, 121: 9108-9126.

Table 1.

	determination of MDF						new determination	
	$3MK_1$	M_1	$3MO_1$	$3MJ_2$	$3MK_2$	$3MO_2$	3D	3SD
wave group								
number waves	(5)	(5)	(6)	(7)	(8)	(6)	(91)	(84)
amplitude factor δ	1.0875	1.0835	1.0879	1.0755	1.0838	1.0724	1.0863	1.0726
mean square deviation	± 90	± 43	± 80	± 106	± 28	± 30	± 39	± 22

Traduction

RELATION DE LA SEISMICITE AVEC LES PHASES
DES ONDES DE MARÉE INDIVIDUELLES.

1994. V.A. Nikolayev

Présenté par l'académicien M.A. Sadovski 11.10.93.

· Edité le 20.10.1993.

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Géophysique.

Il y a très longtemps on avait déjà noté une relation statistique entre les tremblements de terre et les phases des marées terrestres reflétant un ordre interne du milieu [1]. Cette relation réside dans l'effet initiant ou comprimant des tensions de marées sur le milieu au foyer du tremblement de terre en préparation en fonction de la direction des tensions agissantes au foyer.

Notre étude s'oriente sur l'étude expérimentale du lien entre les phases des composantes harmoniques individuelles et les époques d'apparition des tremblements de terre à différents niveaux d'échelle: planétaire et régional. La base est le fait que la dépendance propre aux systèmes instables est la répercussion de la fréquence de l'action (selectivité de fréquence), le phénomène "de la main mise de la fréquence" [2] ce qui apparaît dans les processus les plus différents, par exemple dans l'émission acoustique, l'émission électromagnétique, dans la répercussion géochimique sur l'effet de vibration, effet d'autogénération séismique [3 à 5]. Simultanément on sait que la marée terrestre est composée de plusieurs dizaines d'ondes individuelles d'une amplitude différente, de période de 8 heures jusqu'à quelques mois, ce qui permet d'étudier cet effet et, grâce à de longues séries d'observations, d'identifier même ses faibles apparitions. Nous examinerons ici les variations temporelles de l'état de tension du milieu géophysique dans différents niveaux d'échelle.

L'importance des marées terrestres réside dans le fait que le phénomène est pratiquement unique dans la déformation de la Terre pour laquelle on connaît la force provoquant la déformation et on a la possibilité de la calculer. La méthode réside dans le calcul des composantes de marées théoriques des diverses ondes au moment même où se sont produits des tremblements de terre. Pour des volumes déterminés de la lithosphère et également des intervalles de temps déterminés on fait la somme des nombres de tremblements de terre dans la phase de compression N_+ (tension de marée croissante) et dans la phase de dilatation N_- (la tension de marée diminue) pour les composantes verticales des diverses ondes de marées. On n'analyse que les composantes verticales de marées de diverses ondes car lors de la recherche d'un effet il suffit d'examiner une seule composante.

La non fortuité de la différence des valeurs N_+ et N_- se caractérisera par leurs écarts à la valeur moyenne $(N_+ + N_-)/2$ correspondant à la même probabilité d'apparition d'un tremblement de terre dans les phases positive et négative de l'onde de marée et se soumettant à la répartition binomiale; ainsi la valeur quadratique moyenne des variations de la moyenne

$$\sigma = 0.5 (N_+ + N_-)^{0.5}$$

Pour comparer les résultats et l'estimation simultanée de la valeur statistique on calcule les différences standardisées

$$t = (N_+ - N_-)/2\sigma$$

Cette valeur caractérise l'écart de la valeur moyenne exprimée en unités de l'écart standard. Si le module t dépasse la valeur 2.0, cela témoigne de la non fortuité de l'écart avec une confiance 0.95.

Le programme de calcul des phases, des amplitudes et de leurs variations pour les diverses ondes de marées et pour les composantes du vecteur de marée a été réalisé sur personal computer avec pour base l'algorithme de Longman [6]. La méthode de calcul se fait comme suit: pour le territoire étudié on fait la sommation du nombre de tremblements de terre qui se sont produits dans la phase de compression de marée et dans la phase de dilatation de marée aussi bien pour les composantes du vecteur de marée que pour les ondes individuelles. Le résultat obtenu est moyenné sur un intervalle de 6 mois et ensuite on en étudie la signification statistique.

Les parties avec accroissement des valeurs des différences se rapporteront aux intervalles de compression et les parties avec diminution de ces valeurs se rapporteront aux intervalles de dilatation. Ensuite on calcule les diagrammes spectro-temporel indiqués sur les figures 1, 2.

Pour étudier les variations temporelles de la relation de la séismicité planétaire et des phases des ondes de marées on a analysé 38193 tremblements de terre provenant du catalogue de Hadson, dont la magnitude est supérieure à 5.0 et qui se sont produits sur toute la Terre de 1962 à 1988. Sur la figure 1 on a représenté le diagramme spectro-temporel de la répartition des différences standardisées du nombre de tremblement de terre dans les phases de compression et dans les phases de dilatation pour les diverses ondes de marées. On note sur l'axe horizontal l'époque correspondant au centre de la fenêtre temporelle glissante sans recouvrement égal à 6 mois. Sur l'axe vertical on note dans la partie inférieure les ondes de marées de périodes de 8 heures à 6 mois et dans la partie supérieure les composantes séparées du vecteur de marée: module, composantes verticale, Nord-Sud et Est-Ouest. Des écarts statistiquement importants pour les composantes et les ondes sont notés sur le diagramme par un trait condensé situé sur les niveaux des ondes correspondantes et des composantes. L'intensité et la localisation de ces zones témoignent de la domination de certaines ondes de marées. Cependant la domination des ondes séparées s'observe sur différents intervalles de temps et a une durée différente. La partie supérieure du diagramme de la figure 1 montre que les tremblements de terre prédominent de façon stable dans les phases de compression dans la direction N.S et dans les phases de dilatation dans la direction E.W.

Il y a des intervalles de réorganisation de la structure spectrale, par exemple de 1969 à 1971 s'exprimant par le remplacement des ondes dominantes par d'autres, par leur conjonction et par leur opposition. De 1966 à 1975 l'onde bimensuelle Mf se détache bien, qui ensuite s'évanouit. La réorganisation principale de la structure spectrale se passe de 1969 à 1971: les ondes significatives Ssa, Mm, Q₁, O₁, L₂ s'évanouissent statistiquement et apparaissent l'onde μ_2 qui domine ensuite 18 ans consécutivement, ainsi que les ondes ψ_1 , ϕ_1 , 2N₂. On peut isoler aussi en 1977 la désagrégation de l'onde ϕ_1 tandis que l'onde J₁ se détache. Les intervalles de réorganisation sont remplacés par un intervalle de stabilité relative en 1972 à 1989. Il convient de noter également la domination des ondes semi-diurnes S₂ et R₂ commençant en 1976 et jusqu'en 1989.

Pour l'analyse des tremblements de terre de magnitudes moyennes de 3.0 à 7.0 dans l'échelle régionale on a examiné 9700 tremblements de terre d'Asie Centrale pour 1962 à 1989. Par la partie supérieure des diagrammes de la figure 2 on constate que depuis 1969 les tremblements de terre dominent dans les phases de dilatation selon la composante Ouest-Est. Depuis 1971 l'onde de 6 mois Ssa se sépare bien pour les phases de dilatation, s'évanouit ensuite

et réapparaît à partir de 1978. Depuis 1972 l'onde mensuelle Mm se sépare en phases de compression, restant significative jusqu'à la fin des observations en 1989. Depuis 1975 le nombre de tremblements de terre croît dans les phases de dilatation de l'onde diurne S₁. Nous noterons parmi les autres ondes à courte période, de 8 heures, remarquable dans les phases de dilatation sur tout l'intervalle des observations. La réorganisation principale de la structure spectrale se passe de 1966 à 1970: les ondes Mf, ψ_1 , ϕ_1 , OO₁, L₂, T₂ statistiquement significatives avant cela s'amortissent, étant apparues pendant 1 à 2 ans et l'onde Ssa se manifeste, domine avec une rupture de trois ans jusqu'en 1989, les ondes Mm et S₁ dominent aussi jusqu'à la fin des observations. On peut séparer aussi l'onde S₁ dont se détachent en 1982 les ondes P₁ et K₁ faiblement exprimées. Les intervalles de réorganisation se remplacent par un intervalle de stabilité relative de 1975 à 1989.

Les diagrammes spectro-temporels pour la séismicité planétaire forte (figure 1) et de la séismicité régionale modérée de l'Asie Centrale (figure 2) indiquent avant tout la prédominance de la dilatation dans la direction Est-Ouest. En même temps ces diagrammes ont des harmoniques dominantes qui en général ne coïncident pas. Il convient de noter le phénomène d'inversion: les forts tremblements de terre planétaires coïncident avec les phases de compression pour l'onde μ_2 , les faibles tremblements de terre d'Asie Centrale se sont produits le plus souvent dans les phases de dilatation pour cette même onde.

La non concordance des ondes dominantes pour les tremblements de terre forts de tout le globe terrestre et régionaux, des tremblements de terre plus faibles de l'Asie Centrale s'explique avant tout par l'apport dominant dans la carte planétaire des tremblements de terre de la ceinture de l'océan Pacifique qui a une structure et des tensions agissantes différent de l'Asie Centrale.

L'instabilité de l'action des ondes séparées sur le processus séismique peut probablement être expliquée par la structure complexe hiérarchique de bloc de l'écorce terrestre. Au cours du temps le milieu géophysique se reforme, les relations existantes sont changées entre les blocs séparés de l'écorce terrestre qui s'expriment en particulier dans la présence des ondes de déformation. La tension dans le milieu se redistribue constamment ce qui amène dans certains intervalles de temps à l'instabilité de l'influence des ondes de marées sur la séismicité.

Le résultat expérimental principal du travail est la mise en évidence d'un phénomène de modulation des tremblements de terre par les composantes périodiques de marées de fréquences différentes.

Les harmoniques de marées dominant dans le processus séismique sont variables dans le temps. Les ondes dominant dans tel intervalle de temps peuvent ensuite s'affaiblir ou disparaître et être remplacées par d'autres.

On a mis en évidence une sélection des ondes de marées dominant à l'échelle planétaire et régionale (Asie Centrale) présentant différents intervalles de temps de stabilité.

Les intervalles de réorganisation de la structure régionale spectrale des ondes de marées dominantes ne sont pas durables et sont remplacés par des intervalles de stabilité plus longs.

Dans les intervalles de temps où le lien entre les composantes du vecteur de marée et de la séismicité n'apparaît pas, il existe un lien statistiquement significatif avec les ondes de marées séparées.

L'auteur remercie A.V. Nicolayev pour son aide dans la réalisation de ce travail.

Ce travail a été réalisé grâce au fond russe des recherches fondamentales.

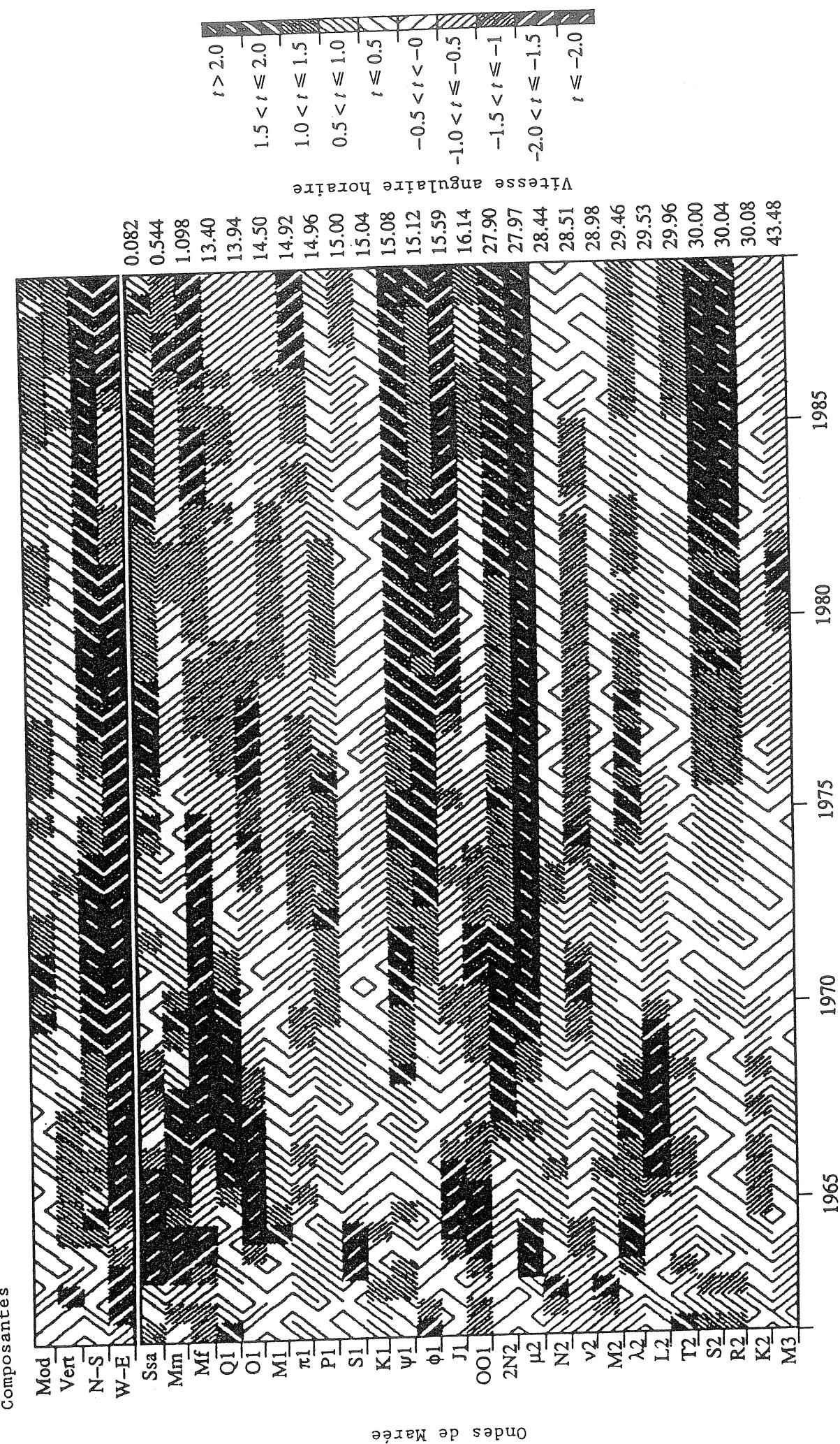


Figure 1 Diagramme spectro-temporel des différences standardisées de la quantité de tremblements de terre qui se sont produits dans la phase de compression de marée et dans la phase de dilatation t pour les tremblements de terre forts, de magnitude $M > 5.0$, pour la période de 1962 à 1988. Relation de la séismicité avec les phases des ondes individuelles de marées.

Composantes

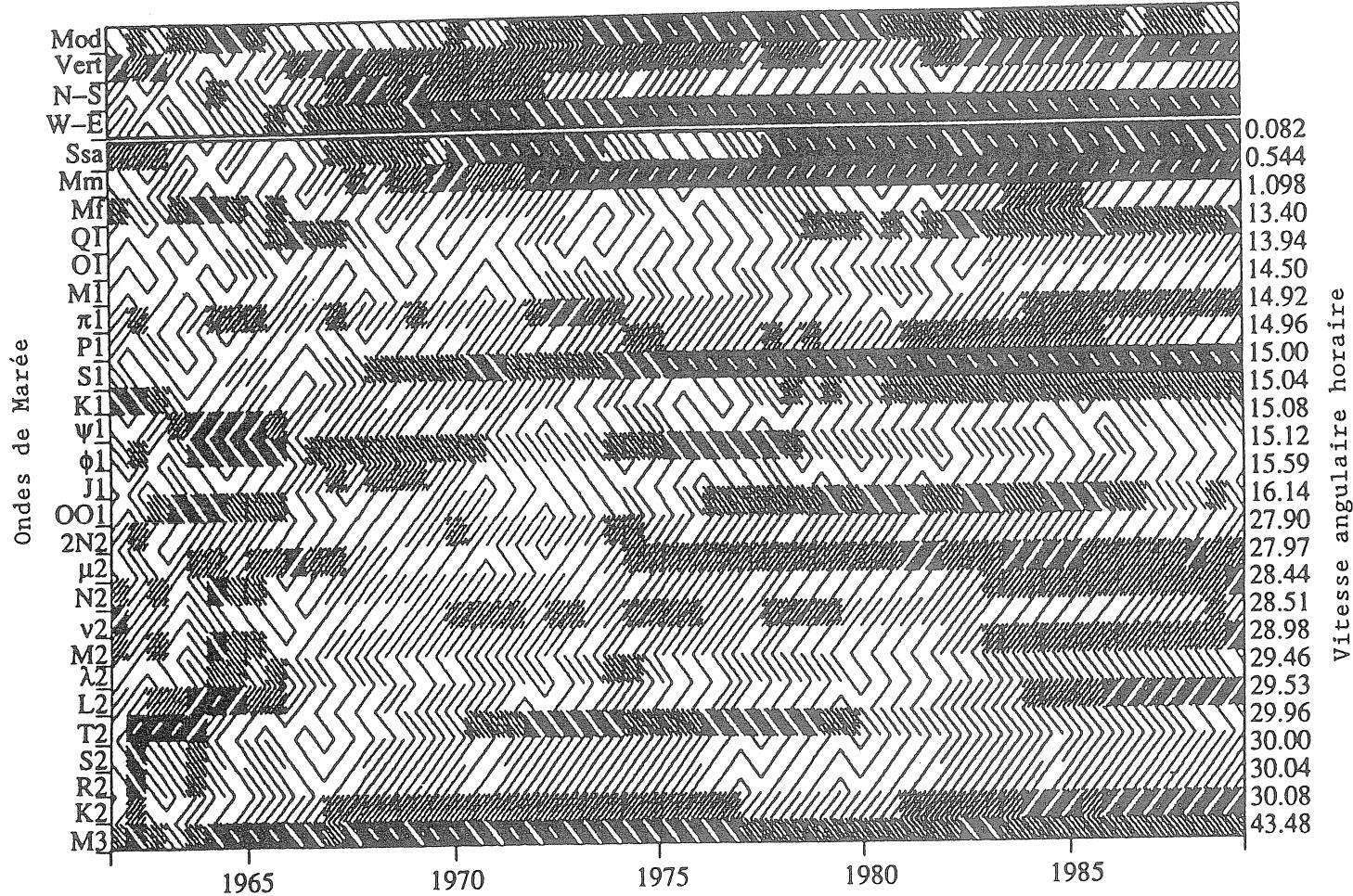


Figure 2 Diagramme spectro-temporel des différences standardisées de la quantité de tremblements de terre qui ont eu lieu dans la phase de compression de marée et dans la phase de dilatation t pour les tremblements de terre régionaux d'Asie Centrale pour la période de 1962 à 1988 ayant une magnitude $M > 3.0$. Mêmes notations que sur la figure 1.

BIBLIOGRAPHIE

1. Николаев А.В., Николаев В.А. // ДАН. 1993. Т. 330. № 2. С. 261 - 266.
2. Каррыев Б.С., Курбанов М.К., Николаев А.В. и др. // ДАН. 1986. Т. 290. № 1. С. 67 - 71.
3. Николаев А.В., Войтов Г.И., Рудаков В.П., Ишанкулиев Д.И. // ДАН. 1993. Т. 330. № 3. С. 376 - 378.
4. Гамбурцев А.Г., Гамбурцева Н.Г., Николаев А.В. и др. // ДАН. 1991. Т. 319. № 2. С. 315 - 318.
5. Гамбурцев А.Г. Сейсмический мониторинг литосферы. М.: Наука, 1992. 196 с.
6. Longman I.M. // J. Geophys. Res. 1959. V. 64. № 12. P. 2351 - 2355.

Traduction

IMPLICATION DES PHASES DE MAREES TERRESTRES
POUR LES SEISMES FORTS.

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On examine ici les problèmes de la corrélation des tremblements de terre forts ayant une magnitude M de plus de 5.0 avec les phases des marées du module du vecteur de marée et de ses différentes composantes : Verticale, Nord-Sud et Ouest-Est. A l'échelle planétaire, dans des fenêtres de $10^\circ \times 10^\circ$, on a calculé la quantité de séismes entre 1962 et 1988 dans les phases de contraction et de dilatation de marée. On a opéré une estimation statistique de la corrélation des phases de marées avec les époques d'apparition de séismes forts, on a séparé les zones de la lithosphère ayant une importance différente dans cette corrélation. On a indiqué les zones de la lithosphère présentant une stabilité différente de la réaction aux tensions de marées pour un intervalle de temps donné. A l'échelle planétaire on a mis en évidence des zones de la lithosphère ayant une haute sensibilité des séismes forts aux phases des différentes composantes de la marée terrestre. Les champs obtenus de la réaction des séismes forts ($M > 5.0$) par les effets de marées interprétés comme des champs de tensions, ont un caractère spatialement réglé avec des anomalies séparées, s'étendant sur 7000 à 8000 km. On a montré une non corrélation de la réaction des séismes par les différentes composantes de marées. On a examiné les problèmes de relation réciproque de la répartition spatiale des structures tectoniques globales et des champs de tensions obtenus.

Par les recherches de ces dernières années on a établi les propriétés physiques des roches et des processus lithosphériques, processus qui ont une haute tenso-vibro sensibilité.

La tenso-sensibilité apparaît dans la modulation caractéristique de l'émission séismique par les tensions de marées [1, 2]; dans les variations de la vitesse de propagation des ondes séismiques en corrélation avec les phases de la Lune et également provoquées par les processus tectoniques fondamentaux.

La vibro-sensibilité apparaît dans les variations de l'émission séismique induites par les microséismes de tempête [2] par le fonctionnement du vibrateur séismique [3]; dans l'influence des explosions souterraines nucléaires et des séismes forts sur la structure fine du processus séismique apparaissant sur des distances d'un millier et demi de kilomètres [4].

On a établi que les régions de haute séismicité, en particulier Pamir-Indu-Kush, zone de séismes de foyer profond [4], ont une sensibilité particulièrement élevée par rapport aux effets séismiques externes. Il est naturel que la haute vibro et tenso sensibilité soit propre aux régions du milieu lithosphérique se trouvant dans un état non stable, à proximité de dislocation et de percolation [5]; aux zones de foyer de séismes forts en préparation. On peut émettre l'hypothèse que les tensions de marées doivent influencer de façon perceptible la structure fine du processus de séismicité. Ce problème a une histoire assez longue; la prévision des séismes en fonction des phases de la Lune et des planètes remonte aux siècles éloignés [6]. Les recherches récentes de l'effet de trigger des marées ont été entreprises par de nombreux chercheurs. Certains ont trouvé une relation statistiquement importante des

marées et des séismes [7 à 13] le plus souvent pour de petites régions séparées, d'autres chercheurs n'ont pas découvert une telle relation [14 à 17]. La plupart de ces résultats ont un caractère disparate mais en partie aussi contradictoire et n'examinent pratiquement pas les variations temporelles de la relation de la séismicité et des marées apparaissant comme un moniteur propre de l'état stressé de la lithosphère.

Des chercheurs précédents ont établi que l'effet trigger de la marée terrestre sur les afterchocs de Spitack (1988) et Ratchinskii (1991) des séismes a un caractère spatialement en mosaïque et après avoir varié dans le temps [18]; la séismicité globale et régionale présente également un caractère spatialement variable de la dépendance de la phase de la déformation de marée et en outre complexe, variable dans le temps et lié d'une certaine façon aux différentes harmoniques de marées [19].

Notre travail est consacré à l'étude de l'influence des marées terrestres sur la structure du processus séismique planétaire.

L'importance des marées terrestres réside dans le fait que ce phénomène de déformation de la Terre est pratiquement le seul pour lequel on connaît la force provoquant la déformation et aussi une possibilité de la calculer. Les tensions de marées créent un petit accroissement variable connu au champ des tensions dont la variation dans les régions séismiquement actives a déterminé pour beaucoup la formation de séismes forts, par le développement de la séismicité fondamentale.

En fonction de la phase les forces de marées peuvent agir aussi bien dans la direction du développement du processus géodynamique qu'à l'inverse de celle-ci en provoquant respectivement une action initiatique ou réprimante. Dans le stade final de préparation du séisme quand un choc extérieur est si faible qu'on peut l'amorcer, ce choc peut se produire par une tension de marée appliquée dans la direction de la tension agissant au foyer du séisme en préparation.

La tension de marée ne dépasse pas 0.3 bar ce qui est 2 à 3 ordres plus petit que la tension mesurée lors d'un séisme fort; cependant, en valeur, la vitesse d'accumulation des tensions tectoniques au foyer préparant un séisme fort est habituellement égale à 0.001 bar/6 heures tandis que la vitesse de variation des tensions de marées est égale à 0.1 bar/6 heures soit 2 ordres plus élevé [7]. Cette relation explique la base physique de l'effet trigger de marée. D'où on peut supposer que le caractère des tensions de marées au moment du séisme correspond au caractère des tensions tectoniques au foyer. La corrélation entre l'orientation de la tension aux foyers des séismes et la direction des tensions de marées est confirmée dans le travail [7] pour 107 séismes forts ($M > 5.0$) et peu profonds ($H < 30$ km) ayant des plans de faille à incidence rapide. C'est pourquoi, en étudiant statistiquement les phases de la marée terrestre aux moments d'apparition des séismes on peut en principe résoudre le problème inverse: trouver les caractéristiques du champ de tension sur les intervalles de temps déterminés et dans les volumes déterminés du milieu [18]. La carte de qualité décrite ici ne tient pas compte de beaucoup de facteurs provoquant également une action initiatrice sur le foyer du séisme. C'est pourquoi le lien du champ des tensions tectoniques avec le champ des tensions de marées doit avoir un caractère caché, statistique.

Méthode de recherche

La méthode repose sur le calcul des composantes de marées théoriques à la surface de la Terre aux endroits des épicentres des séismes et aux époques où ces séismes ont eu lieu. Plus simplement nous supposerons qu'ils caractérisent les forces de marées aussi à la profondeur où se produisent les séismes.

Ensuite on prend le catalogue des séismes pour lequel on a calculé les composantes de marées aux sites correspondants et aux époques correspondantes. Pour des volumes déterminés de la lithosphère et également des intervalles de temps déterminés on additionne les nombres de séismes dans la phase de compression N_+ (la tension de marée croît) et dans la phase de dilatation N_- (la tension de marée décroît) aussi bien pour les différentes composantes du vecteur de marée que pour les différentes ondes de marées.

La non fortuité de la différence des valeurs N_+ et N_- sera caractérisée par leurs écarts à la valeur moyenne correspondant à l'apparition égale des séismes dans les phases positives et négatives de la marée et soumise à une répartition binomiale avec un écart standard

$$\sigma = 0.5 (N_+ + N_-)^{0.5}$$

Pour la comparaison des résultats et l'évaluation simultanée de la signification statistique on calcule les différences normalisées :

$$t = (N_+ - N_-)/2 \sigma$$

Cette valeur caractérise l'écart à la valeur moyenne exprimée en unités de l'écart standard. Si le module t dépasse la valeur 2 cela témoigne de la non fortuité de l'écart avec une probabilité 0.955. Le programme de calcul des phases, des amplitudes des marées et de leurs variations a été réalisé sur un personal computer sur base de l'algorithme de Longman [20]. Le programme permet également de calculer les diverses composantes de la marée: N-S, W-E, verticale et également le module du vecteur de marée. La méthode de calcul se ramène à ce qui suit: le territoire étudié pour lequel on a choisi le catalogue des séismes est partagé en différentes fenêtres, dont la dimension est un compromis entre le détail et la précision dans les limites desquelles on fait la sommation respectivement de la quantité des séismes se produisant dans la phase de reflux et dans la phase de marée. Le résultat obtenu est étudié quant à sa valeur statistique. A la lumière des hypothèses faites plus haut les parties avec prédominance du nombre de séismes dans la phase de compression, se rapporteront aux régions de compression et les parties avec prédominance du nombre des séismes dans la phase de dilatation respectivement aux régions de dilatation. La réaction aux phases du module du vecteur de marée caractérise la structure de tension générale, la réaction aux phases des composantes verticale et horizontale caractérise la compression ou la dilatation respectivement dans les plans vertical ou horizontal.

Etude de la relation de la structure globale de la séismicité avec les marées terrestres

Pour étudier la loi globale du rapport des séismes et des marées terrestres on a utilisé le catalogue de séismes forts de Hadson pour 1962 à 1988 incluant 38.193 séismes avec une magnitude de plus de 5.0, sans limitation en profondeur. La dimension de la fenêtre dans les limites de laquelle on a fait la somme du nombre de séismes dans les mêmes phases, est $10^\circ \times 10^\circ$. Les domaines avec une quantité de séismes inférieure à 4 sont exclus de l'examen ultérieur étant donné leur faible représentativité. Nous nous arrêterons plus en détail sur l'examen de la relation des séismes et des phases de la composante de marée verticale comme la caractéristique le plus souvent examinée de l'effet de marée sur la couche séismo-active. Nous donnons sur la figure 1 la répartition de la relation quantitative des séismes qui se sont produits suivant les différentes phases de la composante verticale. Dans 20 cas d'un total

de 273 cet écart dépasse $2\sigma(7\%)$ ce qui dépasse l'obtention de ce même écart pour le cas stochastique binomial. Pour la quantité totale on observe une égalité exemplaire : $N_+ = 19085$, $N_- = 19101$. Il convient de noter la loi statistique suivante - plus haute est l'activité séismique, plus forte est la réaction aux phases de marées. Pour étudier la stabilité dans le temps des résultats obtenus (l'écart de la quantité moyenne des séismes dans différentes phases de la composante verticale exprimé en unités de σ) on a fait la comparaison des résultats pour deux intervalles: 1962 à 1975 et 1976 à 1988. Une réaction non stable s'exprimant par le changement de signe s'observe dans une grande partie des crêtes médio-océaniques, de la partie Est de la ceinture Alpes-Himalaya, de la fosse du Chili et de la partie Ouest de la ceinture de l'Océan Pacifique. Une réaction stable des séismes aux déformations de marées est caractéristique pour une grande partie de la ceinture Alpes-Himalaya et de la partie Est de la ceinture de l'Océan Pacifique.

La répartition spatiale des domaines avec prédominance des séismes dans les différentes phases de la composante verticale du vecteur de marée est indiquée sur la figure 2. Il faut noter une régularité spatiale déterminée: ainsi les parties avec prédominance d'une des phases en une série de cas s'étendent par une bande continue de 5 à 6000 km (du Kazakhstan au Kamtchatka). Cette régularité peut servir d'indication complémentaire sur le caractère non accidentel du champ étudié. Les régions avec prédominance de séismes dans les phases de compression verticale sont observées dans une grande partie de la ceinture de l'Océan Pacifique; dans la partie Sud-Est (Himalaya) de la ceinture des Alpes-Himalaya; le bassin des Caraïbes; dans une série de crêtes océaniques: la partie Sud de la crête mi-Atlantique et dans l'élévation Est Océan Pacifique. Les régions avec prédominance de séismes dans les phases de dilatation verticale s'observent dans la région de la fosse de Java, dans les parties centrales et Ouest de la ceinture Alpes-Himalaya, dans le rift Est Africain, dans des parties séparées des crêtes médio-océaniques : Africano-Atlantique.

La répartition spatiale des régions avec prédominance des séismes dans les différentes phases de la composante Nord-Sud du vecteur de marée est indiquée sur la figure 3 : c'est une grande partie de l'Asie Centrale, du bassin des Caraïbes et une grande partie de la ceinture de l'Océan Pacifique. Il convient de noter la plus grande régularité spatiale : les parties avec prédominance des séismes dans les phases de dilatation dans une série de cas s'étendent en une ceinture continue jusqu'à 8000 km (crête médiane Arabo-Indienne).

Les régions avec prédominance de séismes dans la phase de dilatation en direction Nord-Sud s'observent dans la Méditerranée, la partie Sud de la ceinture de l'Océan Pacifique, dans la partie Sud Est de la ceinture Alpine-Himalayenne, dans la série des crêtes medianes océaniques et l'élévation du Chili.

Les régions avec prédominance des séismes dans la phase de compression en direction Est-Ouest (fig. 4) s'observent en Asie Centrale, la fosse du Pérou, dans le rift du Baïkal, dans les diverses parties des crêtes médio-océaniques. Les régions avec prédominance de séismes dans la phase de dilatation en direction Est-Ouest s'observent dans la ceinture de l'Océan Pacifique, le bassin des Caraïbes, la Méditerranée et le rift Est Africain.

Il faut probablement interpréter la réaction des tremblements de terre aux phases du module du vecteur de marée (fig. 5) comme l'état général de tension de la lithosphère. Les régions avec prédominance des tremblements de terre dans les phases de compression s'observent dans une grande partie de la ceinture de l'Océan Pacifique outre les Rocky Mountains et la fosse du Pérou; dans la partie Sud-Est de la ceinture Alpine-Himalayenne, dans la série des crêtes médio-océaniques: Africaine, Atlantique, Australo Atlantique, Reykyanes

et la partie centrale de la chaîne médio-Atlantique. Les régions avec prédominance de séismes dans les phases de dilatation s'observent dans la région des fosses du Pérou et de Java; dans la partie Ouest de la chaîne Alpine-Himalayenne, dans le Caucase, les Balkans, dans le rift Est Africain; dans les diverses parties des crêtes médio-océaniques.

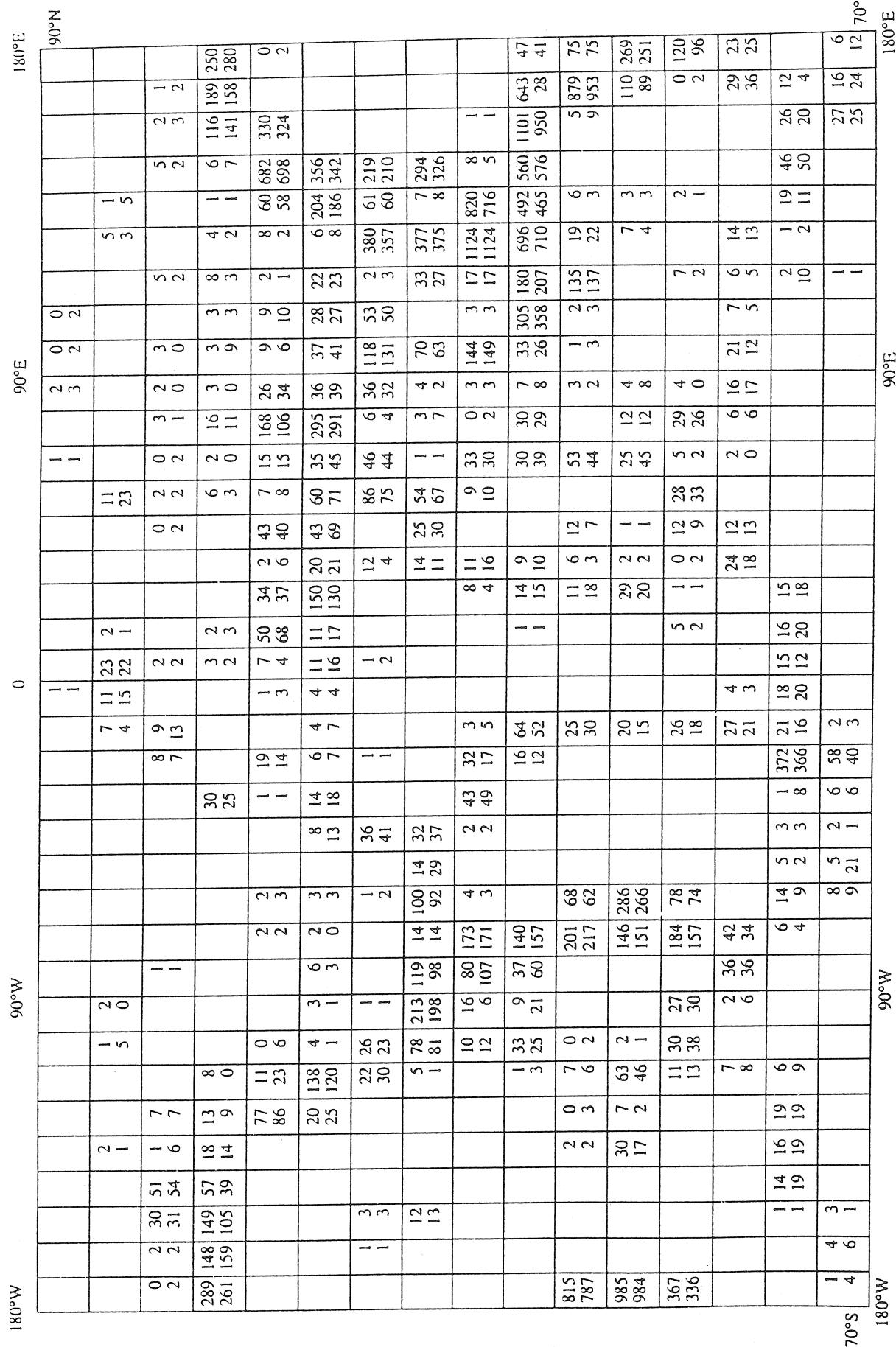
Nous donnons à la figure 6 les directions prédominantes des tensions de marées horizontales au moment où s'est produite la majorité des séismes. Dans la plupart des régions séismiques actives les directions des tensions de marées horizontales prédominantes sont perpendiculaires aux structures tectoniques fondamentales ce qui constitue une confirmation précise de la validité de l'utilisation de la corrélation des phases de marées et des tremblements de terre forts pour l'analyse de l'état de tension de la lithosphère. Ainsi, par exemple pour 80% des parties des crêtes médio-océaniques les tensions horizontales prédominantes sont perpendiculaires aux axes de ces crêtes. Les tensions horizontales dans la région du rift Est-Africain sont également perpendiculaires aux axes des rifts. Pour la série des fosses profondes, sauf le Chili, la prédominance des tensions horizontales perpendiculaires aux axes des fosses est également sensible. Pour la ceinture plissée Alpine Himalayenne la carte de répartition des tensions horizontales est plus complexe et hétérogène.

Une particularité importante des résultats donnés sur les figures 2 à 5 est le fait que les zones pour lesquelles l'effet trigger des phases des composantes de marées est sûr avec une probabilité de 0.955 ne concordent pas pour les différentes composantes. Ce n'est qu'en un seul endroit, en Mésopotamie, que l'écart dépasse 2σ pour les trois composantes et le module, aussi dans deux régions - les monts Altaï et Rockies - l'écart à la moyenne dépasse 2σ pour trois composantes à la fois; dans 8 régions simultanément pour les deux composantes et dans 46 régions cet écart ne s'observe que pour une seule composante. Cette répartition témoigne avant tout de la non corrélation des champs indiqués sur les figures de 1 à 4 et également de l'information indépendante, presque la même, des différentes composantes du vecteur de marée pour le processus séismique global. En tout cas un effet trigger (0.955) statistiquement sûr s'observe dans 57 régions sur 273 c'est à dire dans 20% des cas.

Conclusions principales

A l'échelle planétaire on a mis en évidence des régions de la lithosphère avec une haute sensibilité aux phases des différentes composantes de la marée terrestre. Les champs obtenus de la réponse des séismes forts pour lesquels $M > 5.0$ aux effets de marées interprétés comme champs de tensions ont un caractère spatialement régularisé sur 7000 à 8000 km. On a montré la non corrélation de la réaction des séismes aux différentes composantes de marées.

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Répartition de la quantité de séismes dans la phase de contraction de marée (chiffre supérieur) et dans la phase de dilatation de marée (chiffre inférieur) pour la composante verticale du vecteur de marée.

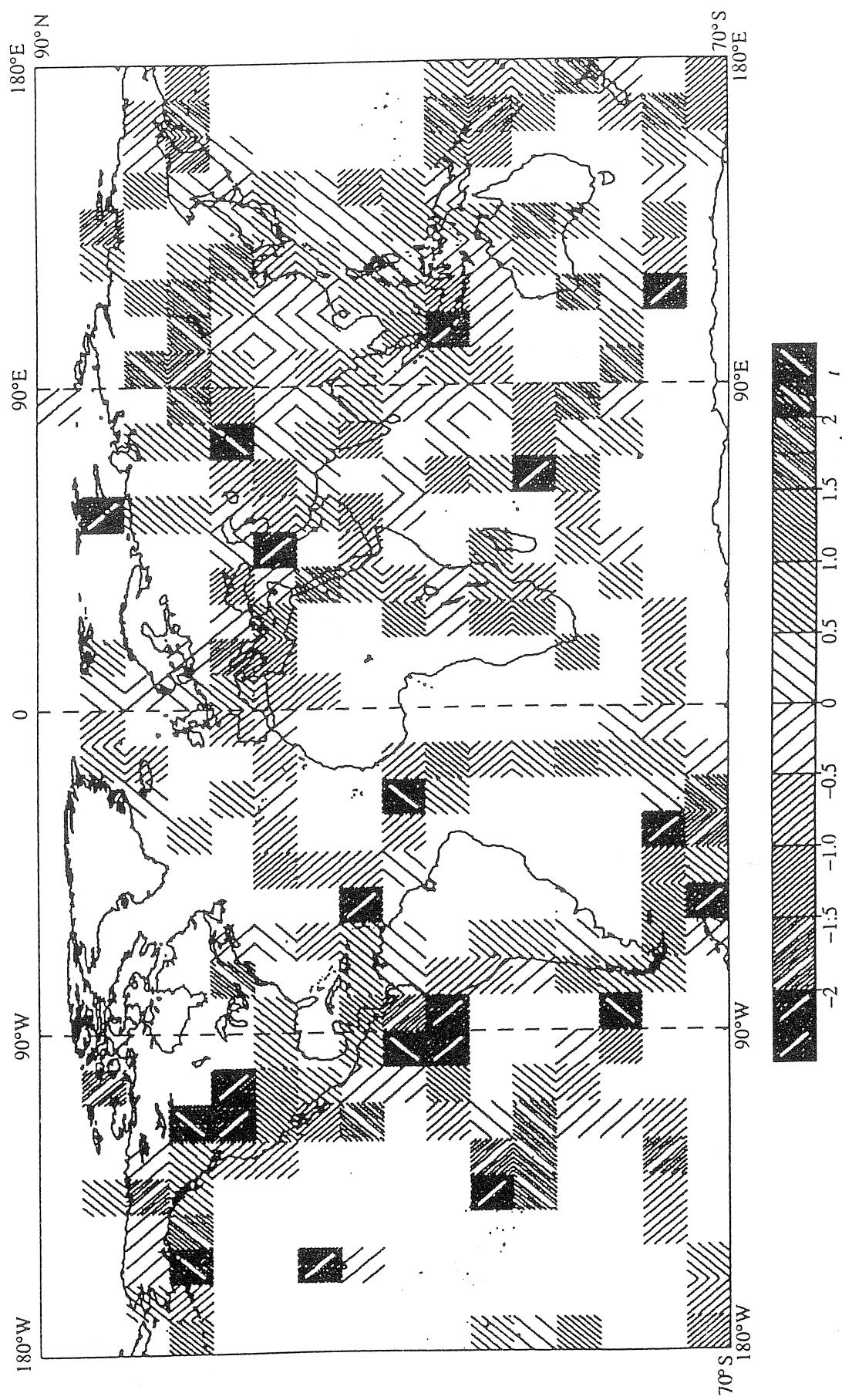


Figure 2 Répartition spatiale des régions avec prédominance des séismes dans les phases de dilatation ($t < 0$) et dans les phases de contraction ($t > 0$) de la composante verticale du vecteur de marée.

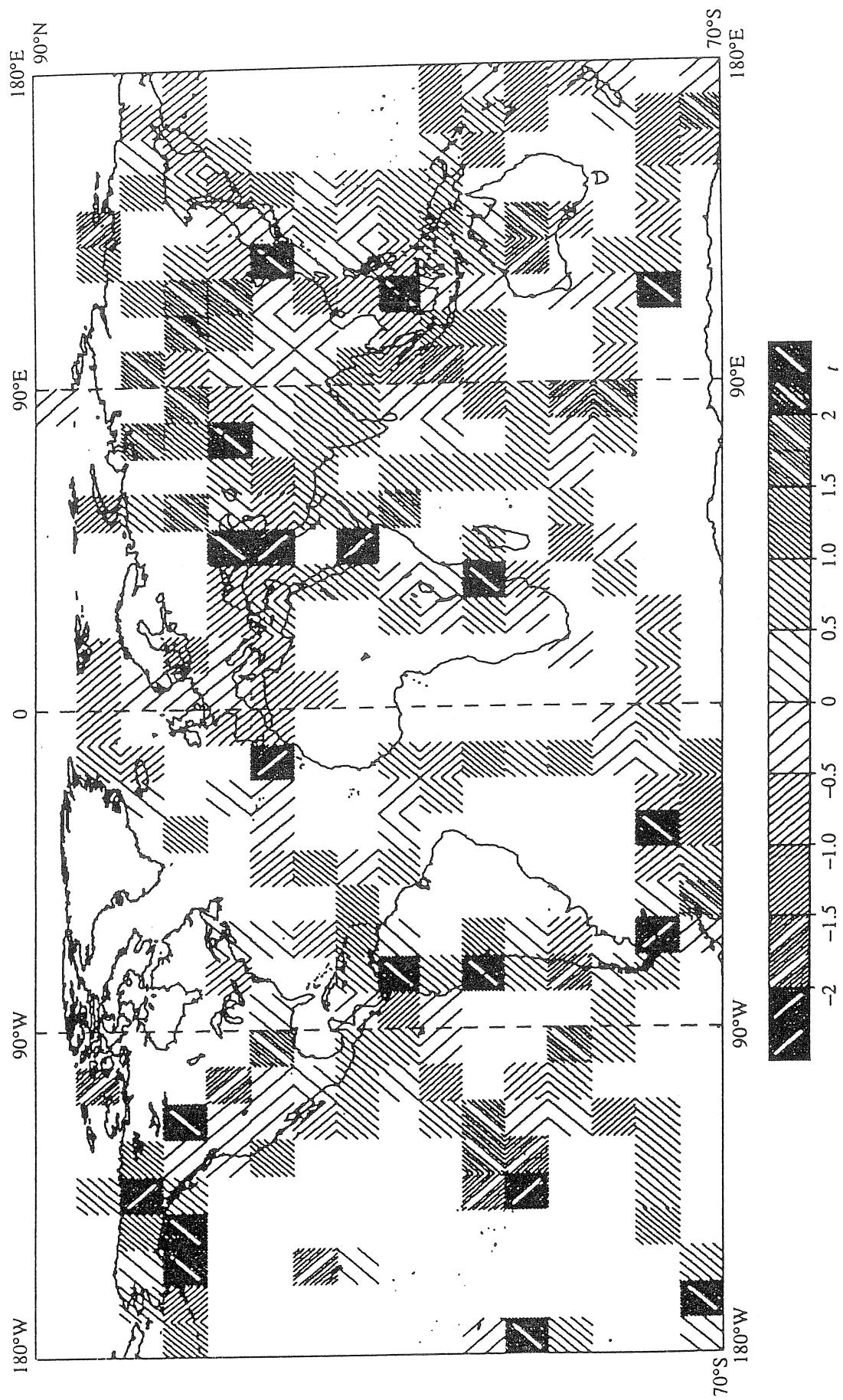


Figure 3 Répartition spatiale des régions avec prédominance des séismes dans les phases de dilatation ($t < 0$) et dans les phases de contraction ($t > 0$) de la composante Nord-Sud du vecteur de marée.

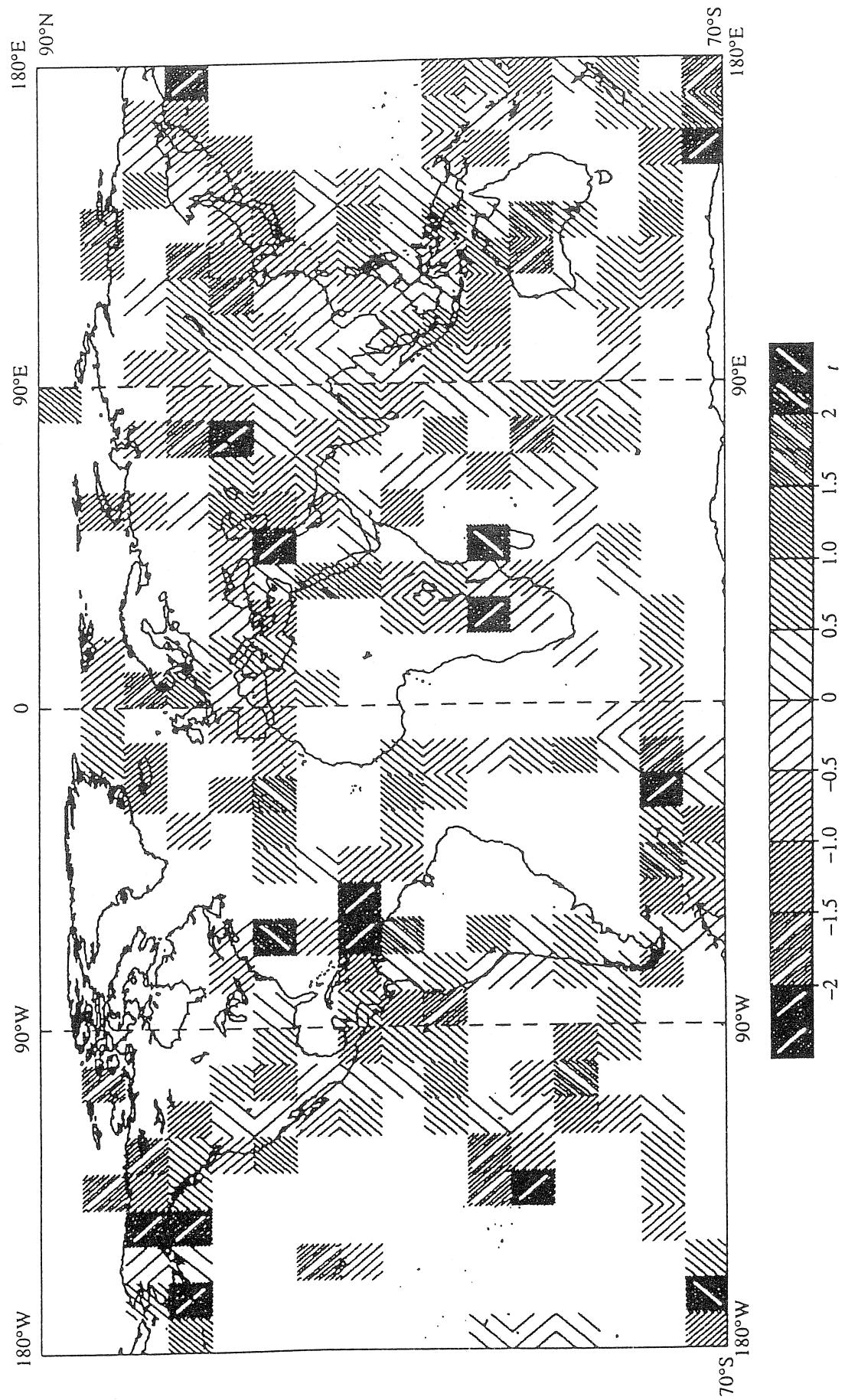


Figure 4 Répartition spatiale des régions avec prédominance des séismes dans les phases de dilatation ($t < 0$) et dans les phases de contraction ($t > 0$) de la composante Est-Ouest du vecteur de marée.

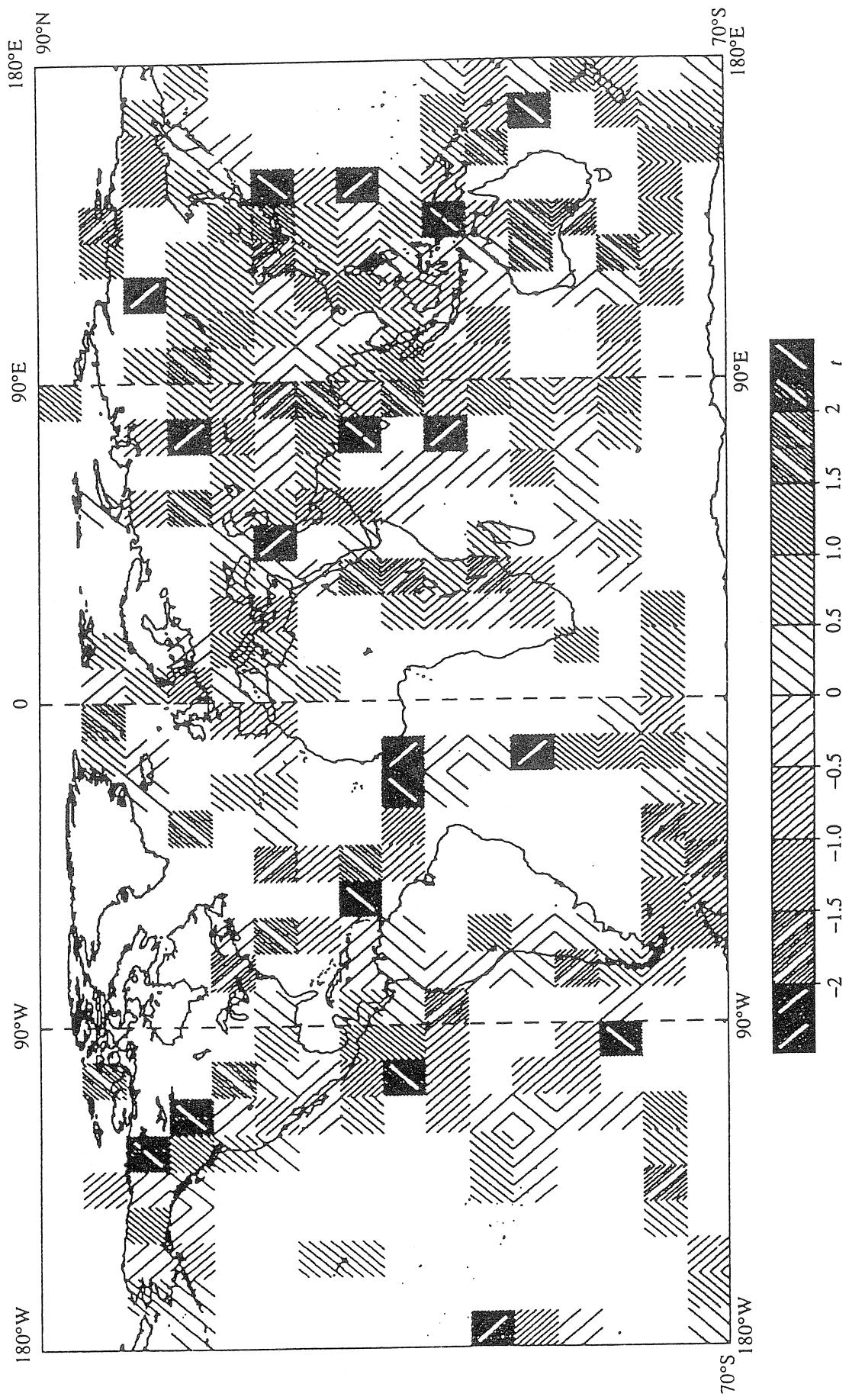


Figure 5 Répartition spatiale des régions avec prédominance des séismes dans les phases de dilatation ($t < 0$) et dans les phases de contraction ($t > 0$) du module du vecteur de marée.

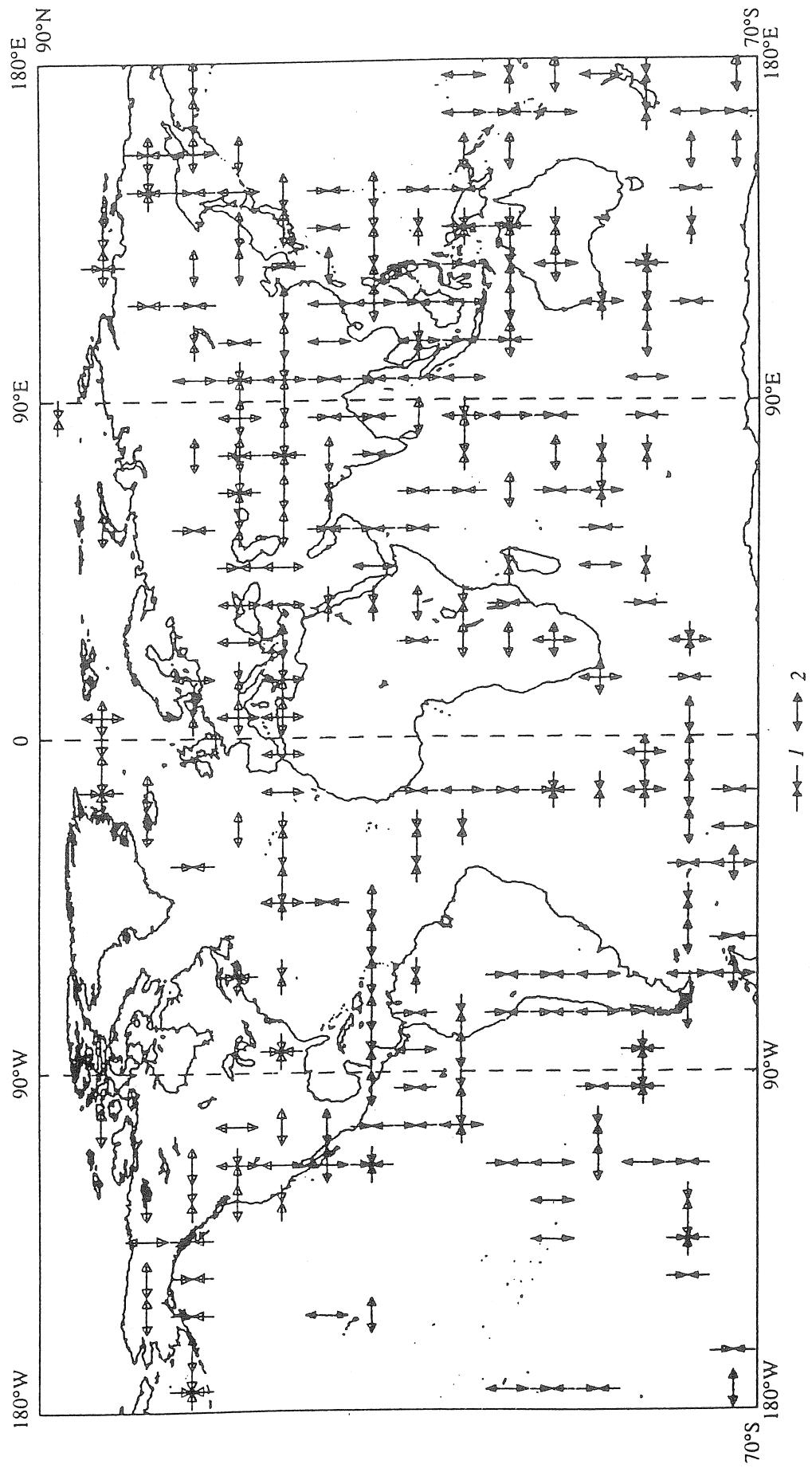


Figure 6 Orientation des tensions prédominantes horizontales
1 - contraction; 2 - dilatation.

BIBLIOGRAPHIE

1. *Yaramanci U., Zschau J.*, Anhang zum DFG Gesamtbericht über das Vorhaben Geogatisch-Geophysikalische Untersuchungen im seismoaktiven Bereich der nord-anatolischen Störzone // Kiel. 1988. T. 1. S. 26.
2. *Рыкунов Л.Н. и др.* Временные вариации высокочастотных шумов // Изв. АН СССР. Сер. Физика Земли. 1979. № 11. С. 72 - 77.
3. *Николаев А.В., Береснев И.А.* Изучение прохождения длительных вибросигналов через геофизическую среду // Изв. АН СССР. Сер. Физика Земли. 1990. № 1. С. 86 - 89.
4. *Николаев А.В., Верещагина Г.М.* Об инициировании землетрясений землетрясениями // ДАН СССР. 1991. Т. 318. № 2. С. 320 - 324.
5. *Челидзе Т.Л., Чарголейшвили Т.Т.* Теория неупорядоченных сред и некоторые проблемы нелинейной сейсмики // Проблемы нелинейной сейсмики. М.: Наука, 1987. С. 41 - 50.
6. *Джалаал ад-Дин.* Трактат о землетрясении. Баку: Элм, 1983. 77 с.
7. *Heaton T.H.*, Tidal Triggering of earthquakes // Geoph. J. Astr. Soc. 1975. V. 43. P. 307 - 326.
8. *Allen M.W.*, The lunar triggering of earthquakes in Southern California // Bull. seism. Soc. Am. 1936. V. 26. P. 147 - 157.
9. *Hoffman R.B.*, Aftershock-energy release versus tidal effects, Hebgen Lake earthquake, Montana // US. Geol. Survey Prof. Paper 1961. 426-C. P. 267 - 270.
10. *Tanriazyan G.P.*, Principal regularities in the distribution of major earthquakes relative to solar and lunar tides and other cosmic sources // Icarus 9. 1968. P. 574 - 592.
11. *Ryall A., Van Wormer J.D., Jones A.E.*, Triggering of microearthquakes by earth tides, and other features of the Truckee, California earthquake sequences of September, 1966 // Bull. seism. Soc. Am. 1968. V. 58. P. 215 - 248.
12. *Weems R.E., Perry W.H.*, Strong correlation of major earthquakes with solid-earth tides in part of the eastern United States // Geology. 1989. V. 17. P. 661 - 664.
13. *Polumbo A.* Lunar and solar tidal components in the occurrence of earthquakes in Italy // Geophys. J. Roy. Astron. Soc. 1986. V. 84. № 1. P. 93 - 99.
14. *Knopoff L.*, Earth tides as triggering mechanism for earthquakes // Bull. seism. Soc. Am., 1964. V. 54. P. 1865 - 1870.
15. *Willis D.E., Taylor R.W.*, Investigation of multiple seismic events and first zone discriminants // ARPA Semi-Annual Technical Report, April 1974. Caltech. P 101 - 156.
16. *Shlien S.*, Earthquake-tide correlation // Geoph. J. Roy. Astr. Soc. 1972. V. 28. P. 27 - 34.
17. *Simpson J.F.*, Earth tides as a triggering mechanism for earthquakes // Earth. Planet. Sci. Lett. 1967. V. 2. P. 473.
18. *Николаев А.В., Николаев В.А.*, Связь афтершоков сильных землетрясений с приливными фазами как индикатор напряженного состояния среды // ДАН. 1993. Т. 330. № 2. С. 261 - 266.
19. *Николаев В.А.* Связь сейсмичности с фазами отдельных приливных волн // Там же. 1994. Т. 336. № 3. С. 98 - 101.
20. *Longman I.M.* // J. Geoph. Res. 1959. V. 64. № 12. P. 2351 - 2355.
21. *Парийский Н.Н., Артамасова З.П., Крамер М.В.* К вопросу о роли приливных напряжений как спускового механизма при землетрясениях // Физические основания поисков методов прогноза землетрясений. М.: Наука, 1979. С. 62 - 63.
22. *Nikolaev V.A.* Tidal triggering of earthquakes as indicator of lithosphere stress state in Mediterranean // European Geoph. Soc. Annales Geophysicae. 1993. Pt 1. P. 66.

Traduction

SUR LES OSCILLATIONS PROPRES D'UN LIQUIDE VISQUEUX
EN ROTATION DANS LE NOYAU EXTERNE DE LA TERRE

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Dans l'espace des champs indivergenciers nous construisons dans une couche sphérique la base des champs propres de l'opérateur de Laplace, nuls sur la limite de la couche. La couche du globe correspond ici au noyau liquide dans le modèle de la Terre. On a montré que si on ordonne les champs propres de l'opérateur de Laplace suivant la croissance des nombres propres, alors la base obtenue apparaît également ordonnée suivant le degré d'oscillation des champs de base. Cette base est utilisée en outre pour une démonstration très simple (s'appuyant sur le théorème de Karleman) de la structure complète du système des fonctions propres et adjointes du problème sur les oscillations libres d'un liquide visqueux dans une couche sphérique lors de faibles écarts à une rotation stable.

Introduction

Le contenu de ce travail est lié au spectre des oscillations libres d'un liquide visqueux dans une couche sphérique lors de faibles écarts à une rotation stable. La couche sphérique correspond au noyau liquide dans le modèle le plus simple de la Terre. On décrit dans ce travail le processus d'établissement des champs propres de l'opérateur de Laplace et du calcul des nombres propres correspondants. Ces champs répartis suivant l'ordre de croissance des nombres propres forment la base dans laquelle les champs sont ordonnés d'après le degré d'oscillation dans l'espace. Cette base peut être utile lors de l'établissement des approximations de Galerkine aux champs propres du problème donné, d'après l'avis des auteurs, et dans les autres problèmes d'hydrodynamique dans la couche sphérique. Le premier nombre propre de l'opérateur de Laplace et la valeur de la viscosité permettent (très grossièrement) d'évaluer ci-dessous le module de la partie effective de la valeur propre la plus proche de zéro du problème de départ.

Ce travail, en outre, renferme la démonstration simple de la structure complète du système des fonctions propres et adjointes de l'opérateur du travail donné. Ce fait est connu et découle de la théorie générale de la stabilité hydrodynamique linéaire (voir 7, 8). Notre démonstration s'appuyant sur le théorème de Karleman est destinée au lecteur ne possédant pas une formation mathématique professionnelle.

Pose du problème

Soit H l'espace de Hilbert des champs vectoriels indivergenciels dans la couche du globe Ω avec le produit scalaire

$$\langle p, q \rangle = \int_{\Omega} (p, q^*) dv$$

(* conjugué complexe).

Nous examinerons en H le problème spectral suivant :

$$Aq \equiv 2[1_z, q] - \nabla\Phi - \epsilon \operatorname{rot} \operatorname{rot} q = \lambda q, \quad \operatorname{div} q = 0, \quad (1.1)$$

$$q|_{\partial\Omega} = 0. \quad (1.2)$$

Ici Ω est une couche du globe ayant pour rayon intérieur r_0 et un rayon extérieur unitaire; 1_z est le vecteur unitaire suivant l'axe de rotation; $\epsilon > 0$ est la viscosité cinématique; Φ est la fonction inconnue scalaire (potentiel) apparaissant en relation avec la condition $\operatorname{div}(Aq) = 0$; le calcul de $\nabla\Phi$ est la projection du champ $2[1_z, q]$ dans l'espace H .

L'équation (1.1) et la condition aux limites (1.2) sont la formulation mathématique du problème des oscillations faibles propres du liquide visqueux dans la couche du globe (voir [1]). Dans ce travail on propose la construction de la base dans l'espace H qui peut être utile lors de l'établissement des approximations de Galerkin des champs propres du problème (1.1) (1.2). Certaines propriétés spéciales de cette base sont indiquées au § 2.4. En outre est donnée la démonstration bien admise (de l'avis des auteurs) de la structure complète du système des fonctions propres et adjointes de l'opérateur examiné.

Domaine de la détermination de notre opérateur $D(A)$: une grande quantité des champs plans en H , satisfaisant à la condition (1.2). Il est évident que $D(A)$ est dense en H .

L'opérateur $P : q \rightarrow 2[1_z, q] - \nabla\Phi\phi$ est appelé opérateur de Poincaré. On sait (voir par exemple, [1]) que l'opérateur P est obliquement symétrique et de normes $|P| = 2$.

A l'aide de la formule $\operatorname{div}[a, b] = b \operatorname{rot} a - a \operatorname{rot} b$ il est simple de démontrer que l'opérateur $\operatorname{rot} \operatorname{rot}$ est symétrique sur les champs de $D(A)$, c'est à dire

$$\langle \operatorname{rot} \operatorname{rot} q, p \rangle = \langle q, \operatorname{rot} \operatorname{rot} p \rangle, \quad p, q \in D(A).$$

Sur les champs indivergenciels, l'égalité $\operatorname{rot} \operatorname{rot} q = -\Delta q$ est exacte, où Δ est l'opérateur Laplacien. C'est pourquoi $A = P - \epsilon B$, où $B = -\Delta = \operatorname{rot} \operatorname{rot}$ est l'opérateur symétrique sur $D(A)$.

2. Etablissement de la base des champs propres de l'opérateur de Laplace dans la couche du globe.

On décrit dans ce chapitre le processus de recherche des champs propres et des nombres propres de l'opérateur B . Ces résultats seront utilisés dans le chapitre 3.

2.1. Indications préliminaires

Nous énumérerons quelques faits connus utilisés dans ce chapitre. Ils sont contenus, par exemple en [2] et [3].

2.1.1. On sait que $2m + 1$ des fonctions sphériques d'ordre m

$$P_m^n(\theta, \varphi) = e^{in\varphi} P_m^{(n)}(\cos \theta), \quad n = 0, \pm 1, \dots, \pm m.$$

satisfont à la relation

$$\Delta_s P_m^n = -m(m+1)P_m^n,$$

où

$$\Delta_s = \frac{1}{\sin \theta} \left[\frac{\partial}{\partial \theta} \left(\sin \theta \frac{\partial}{\partial \theta} \right) - \frac{1}{\sin \theta} \frac{\partial^2}{\partial \varphi^2} \right]$$

est la partie angulaire de l'opérateur scalaire de Laplace (r, θ, ϕ sont les coordonnées sphériques dans la région Ω)

$$P_m^{(n)}(\cos \theta) = (1 - x^2)^{n/2} \left. \frac{d^n}{dx^n} P_m(x) \right|_{x=\cos \theta},$$

où $P_m(x)$ est le polynôme de Legendre du même degré. En coordonnées sphériques, l'opérateur de Laplace a sur les fonctions scalaires la forme

$$\Delta = \frac{1}{r^2} \left[\frac{\partial}{\partial r} \left(r^2 \frac{\partial}{\partial r} \right) + \Delta_s \right]. \quad (2.1)$$

2.1.2. Les fonctions sphériques satisfont aux relations d'orthogonalité

$$\int_{S_1} P_k^l P_m^n d\sigma = \delta_{km} \delta_{ln} Z_{kl}, \quad (2.2)$$

$$\int_{S_1} (\nabla_s P_k^l, \nabla_s P_m^n) d\sigma = k(k+1) \delta_{km} \delta_{ln} Z_{kl}, \quad (2.3)$$

où S_1 est la sphère unitaire;

$$Z_{kl} = \int_{S_1} (P_k^l)^2 d\sigma; \nabla_s = \frac{\partial}{\partial \theta} e_\theta + \frac{1}{\sin \theta} \frac{\partial}{\partial \varphi} e_\varphi$$

est le gradient angulaire; e_r, e_θ, e_φ sont les vecteurs de la base locale.

2.1.3. Les champs vectoriels satisfaisant aux relations

$$\mathbf{B}_t = \text{rot}(T(r)\mathbf{r}), \quad \mathbf{B}_p = \text{rot rot}(P(r)\mathbf{r})$$

où $T(r)$ et $P(r)$ sont des fonctions scalaires arbitraires, sont appelés respectivement champs toroïdaux et poloïdaux. On sait (voir par exemple [3, 4] que l'espace des champs indépendants de la divergence se compose de la somme orthogonale des espaces constitués par les champs toroïdaux et poloïdaux, en outre,

$$\Delta(P(r)\mathbf{r}) = \Delta P(r)\mathbf{r} \text{ et}$$

et

$$\text{rot } \mathbf{B}_p = -\text{rot}(\Delta P(r)\mathbf{r}). \quad (2.4)$$

De la formule (2.4) il s'ensuit que l'opérateur $B = \text{rot rot}$ conserve le type du champ : les champs poloïdaux donnent des poloïdaux, les toroïdaux des toroïdaux. Cela signifie que l'espace H est la somme orthogonale des sous espaces, invariants pour l'opérateur B , et on peut chercher séparément les champs propres du type toroïdal et poloïdal.

2.2. Champs propres toroïdaux et poloïdaux

2.2.1. Soit la fonction $Q(r)$ satisfaisant à l'équation

$$-\Delta Q(r) = \lambda Q(r), \quad (2.5)$$

où Δ est l'opérateur scalaire, λ un nombre arbitraire. Alors de (2.4) il résulte que le champ toroïdal $t = \text{rot}(Q(r)\mathbf{r})$ sera un champ propre pour l'opérateur B , puisque

$$Bt = -\text{rot}(\Delta Q(r)\mathbf{r}) = \lambda \text{rot}(Q(r)\mathbf{r}).$$

Et par contre : si $t = \text{rot}(Q(r)\mathbf{r})$ est un champ propre non nul pour l'opérateur B , alors

$$-\text{rot}(Q(r)) = \lambda \text{rot}(Q(r)\mathbf{r}),$$

où

$$(\Delta Q(r) + \lambda Q(r))\mathbf{r} = \nabla \varphi,$$

où φ est une fonction arbitraire. Si $\nabla \varphi \neq 0$, alors il est évident que $\nabla \varphi = d\varphi/dr e_r$ et Q doit être fonction uniquement de la seule variable radiale r . Mais alors

$$t = \text{rot}(Q(r)\mathbf{r}) = [\nabla Q(r), r] = 0.$$

Puisque le champ t a été proposé non nul, alors $\nabla \phi = 0$, et

$$-\Delta Q(r) = \lambda Q(r).$$

Il est juste également (en appliquant deux fois la formule (2.4)) qu'on peut s'assurer que le champ poloïdal $p = \text{rot rot}(F(r)r)$ sera champ propre pour l'opérateur B (avec un nombre propre $\tilde{\lambda}$) alors et seulement alors, quand

$$\Delta^2 F + \tilde{\lambda} \Delta F = 0. \quad (2.6)$$

2.2.2. Les champs propres de l'opérateur B doivent satisfaire à la condition aux limites (1.2). En résolvant l'équation (2.5) par la méthode de séparation de variables (et en utilisant la formule (2.1) pour l'opérateur Δ) nous obtiendrons que les fonction $Q(r)$ aient la forme

$$Q(r) = R(r)P_k^l(\theta, \varphi),$$

où $R(r)$ est la fonction propre de l'opérateur

$$T_k = \frac{1}{r^2}(k(k+1) - \frac{\partial}{\partial r}(r^2 \frac{\partial}{\partial r})). \quad (2.7)$$

Le champ toroïdal

$$t = \text{rot}(R(r)P_k^l r) = R(r)[\nabla P_k^l, r]$$

satisfait à la condition (1.2) si

$$R(r_0) = R(1) = 0. \quad (2.8)$$

Dans le chapitre 2.3 on indiquera comment trouver les nombres propres λ_{ik} et les fonctions propres correspondantes $R_{ik}(r)$ de l'opérateur T_k . Les champs propres toroïdaux de l'opérateur B ont ainsi la forme

$$t_{ik}^l = \text{rot}(R_{ik}(r)P_k^l(\theta, \varphi)), \quad l = 1, 2, \dots, \quad k = l, l+1, \dots, \quad i = 1, 2, \dots$$

En résolvant de la même façon l'équation (2.6) nous obtiendrons que

$$F = \tilde{R}(r)P_m^n(\theta, \varphi),$$

où la fonction $\tilde{R}(r)$ satisfait à l'équation

$$T_m^2 \tilde{R} = \tilde{\lambda} T_m \tilde{R} \quad (2.9)$$

(l'opérateur T_m est déterminé par la formule (2.7))

De la relation

$$p = \text{rot rot}(\tilde{R}(r)P_m^n r) = \tilde{R}[\nabla, [\nabla P_m^n, r]] + [\nabla \tilde{R}, [\nabla P_m^n, r]]$$

il en résulte que le champ p satisfera à la condition (1.2), si

$$\begin{cases} \tilde{R}(r_0) = \tilde{R}(1) = 0 \\ \tilde{R}'(r_0) = \tilde{R}'(1) = 0 \end{cases} \quad (2.10)$$

L'équation (2.9) admet des solutions $(\tilde{R}_{jm}(r))$ satisfaisant aux conditions (2.10) et l'admission correspondante des nombres $(\tilde{\lambda}_{jm})$ (comment les calculer est indiqué dans le chapitre 2.3). Ainsi les champs propres poloïdaux pour l'opérateur B ont la forme

$$P_{jm}^n = \text{rot rot}(\tilde{R}_{jm}(r)P_m^n(\theta, \varphi)r) \quad n = 1, 2, \dots, m = n, n+1, \dots, j = 1, 2, \dots$$

2.2.3. Dans l'espace des champs indivergenciers dans la région limite Ω les champs propres de l'opérateur $B = -\Delta$ satisfaisant à la condition (1.2) forment un système orthogonal complet mais l'opérateur même B est déterminé positif (voir par exemple [5] chapitre 6). Par conséquent les champs propres toroïdaux et poloïdaux construits (avec des coefficients convenables) forment une base orthonormale en H . Cette base peut être matérielle, en prenant des combinaisons linéaires des exposants entrant en P_k^l . Partout dans la suite nous considérerons que la composition des champs (t_{ik}^l, p_{jm}^n) est une base orthonormale matérielle dans l'espace H .

2.2.4. L'étude plus détaillée de la matrice de l'opérateur P dans la base construite [10] (en tenant compte des formules (2.2) et (2.3) et des relations de récurrence pour les polynômes de Legendre, voir par exemple [2]) montre que l'espace H se décompose en la somme directe des sous espaces invariants par rapport à l'opérateur A , H^l , $l = 1, 2, \dots$ correspondant à l'indice supérieur dans la notation de la fonction sphérique. La base dans chaque H^l est formée par les champs (t_{ik}^l, p_{jm}^n) où l est fixé, k et m parcourront indépendamment les valeurs $l, l+1, l+2, \dots, i = 1, 2, \dots, j = 1, 2, \dots$. Il est évident qu'on peut faire les examens ultérieurs dans H^l arbitraire que, par simplicité, nous désignerons par le symbole précédent H .

2.3. Choix des fonctions, dépendant du rayon, pour les champs propres de l'opérateur de Laplace

Dans ce chapitre nous montrons comment calculer les formes (λ_{ik}, R_{ik}) et $(\tilde{\lambda}_{jm}, \tilde{R}_{jm})$ pour réaliser la construction de la base en H .

2.3.1. Nous examinerons l'équation

$$T_k R = \lambda R, \quad (2.11)$$

où T_k est déterminé par la formule (2.7), $\lambda = \xi^2 > 0$. Il y a une solution générale

$$R = C_1 j_k(\xi r) + C_2 y_k(\xi r),$$

où j_k et y_k sont les fonctions sphériques de Bessel d'ordre k [9]. Les conditions (2.8) donnent le système d'équations pour la détermination de C_1 et C_2 :

$$\begin{cases} C_1 j_k(\xi r_0) + C_2 y_k(\xi r_0) = 0 \\ C_1 j_k(\xi) + C_2 y_k(\xi) = 0 \end{cases} \quad (2.12)$$

Les zéros de la fonction

$$\det_1(\xi) = \begin{vmatrix} j_k(\xi r_0) & y_k(\xi r_0) \\ j_k(\xi) & y_k(\xi) \end{vmatrix}$$

donnent la forme (ξ_{ik}) et la forme correspondante (R_{ik}) :

$$R_{ik}(r) = C_1(\xi_{ik})j_k(\xi_{ik}r) + C_2(\xi_{ik})y_k(\xi_{ik}r),$$

où $C_1(\xi_{ik})$, $C_2(\xi_{ik})$ sont les solutions du système (2.12) pour $\xi = \xi_{ik}$.

2.3.2. L'équation.

$$T_m^2 \tilde{R} = \tilde{\lambda} T_m \tilde{R} \quad (\tilde{\lambda} = \nu^2 > 0)$$

a pour solution générale

$$\tilde{R} = \tilde{C}_1 j_m(\nu r) + \tilde{C}_2 y_m(\nu r) + \tilde{C}_3 r^m + \tilde{C}_4 r^{-(m+1)},$$

où les deux premiers termes sont la solution générale de l'équation (2.11) pour $\lambda = \nu^2$ et les deux suivants appartiennent au noyau de l'opérateur T_m . Les conditions aux limites (2.10) donnent le système d'équations pour la détermination de \tilde{C}_n , $n = 1, \dots, 4$.

En utilisant la relation

$$f'_n(z) = f_{n-1}(z) - \frac{n+1}{z} f_n(z)$$

pour les fonctions sphériques de Bessel [9] on peut écrire le déterminant de ce système

$$\begin{aligned} \det_2(\nu) &= \\ &= \begin{vmatrix} j_m(\nu r_0) & y_m(\nu r_0) & r_0^m & r_0^{-(m+1)} \\ j_m(\nu) & y_m(\nu) & 1 & 1 \\ \nu j_{m-1}(\nu r_0) & \nu y_{m-1}(\nu r_0) & r_0^{m-1}(m\nu^m + m + 1) & (m+1)r_0^{-(m+2)}(1 - \nu^{-(m+1)}) \\ \nu j_{m-1}(\nu) & \nu y_{m-1}(\nu) & m\nu^m + m + 1 & (m+1)(1 - \nu^{-(m+1)}) \end{vmatrix}. \end{aligned}$$

Connaissant les zéros $\{\nu_{jm}\}$ de la fonction $\det_2(\nu)$ on peut trouver $\tilde{C}_n(\nu_{jm})$, $n = 1, \dots, 4$ et par cela même les fonctions $\tilde{R}_{jm}(r)$. Pour calculer les zéros on peut se servir du théorème de Laplace sur la décomposition du déterminant et par le rapport [9] :

$$\begin{vmatrix} j_m(\nu) & y_m(\nu) \\ j_{m-1}(\nu) & y_{m-1}(\nu) \end{vmatrix} = \frac{1}{\nu^2}.$$

Remarque

Nous rappellerons qu'ici H désigne n'importe quel H^l , $\ell = 1, 2, \dots$ (voir §2.2.4). Les calculs décrits dans les alinéas 2.3.1 et 2.3.2 sont les mêmes pour tous les H^l , il n'y a que la limite inférieure des indices k et m ($m, k \geq \ell$) qui change.

2.4. Propriété supplémentaire de la base construite

2.4.1. Au § 2.3 on a obtenu les compositions (λ_{ik}, t_{ik}) et $(\tilde{\lambda}_{jm}, p_{jm})$. Nous grouperons les nombres λ_{ik} et $\tilde{\lambda}_{jm}$ dans une seule composition, nous ordonnerons ces nombres selon leur croissance et nous numérotterons dans l'ordre les champs correspondants. Nous obtiendrons les nombres $0 < \alpha_1 < \alpha_2 < \dots$ et la succession (e_1, e_2, \dots) dans laquelle les champs toroidal et poloïdal sont mélangés. Nous montrerons plus loin que la base obtenue (e_i) est réglée dans un seul sens (ce qui peut n'être pas tout à fait évident).

2.4.2. Dans le travail [10], pour construire la base dans l'espace des champs dans la couche sphérique avec la condition $(q, n)|_{\partial G} = 0$, on utilise la fonctionnelle

$$\mathcal{R}^2(q) = \frac{\int_G B(q) dv}{\int_G |q|^2 dv},$$

où $\mathcal{R}(q) = \text{Tr}(D^* D)$, $D = G^{1/2} Q G^{-1/2}$, G est le tenseur métrique, Q est le tenseur covariant du champ q produit. La valeur $\mathcal{R}(q)$ a été proposée par les auteurs à titre d'estimation du degré d'oscillation du champ dans l'espace (plus grand est $\mathcal{R}(q)$ plus est forte l'oscillation du champ). Si $\{q_n\}$ est la base en H ,

$$q = \sum c_n q_n, \sum |c_n|^2 = 1$$

alors

$$\mathcal{R}^2(q) = \sum_{n,m} c_n^* c_m M_{nm},$$

où

$$M_{nm} = \int_{\Omega} \text{Tr}(D^*(q_n) D(q_m)) dv$$

matrice symétrique.

Le problème a été ramené ainsi à la construction de la base $\{q_n\}$ dans laquelle la matrice (M_{nm}) a une forme diagonale. Les éléments $\{q_n\}$ ont été ensuite ordonnés dans l'ordre de croissance de $\mathcal{R}(q_n)$.

2.4.3. Nous montrerons que pour les champs satisfaisant à la condition (1.2) à la limite, la relation

$$\int_{\Omega} \text{Tr}(D^*(q_n) D(q_m)) dv = - \int_{\Omega} (\Delta q_n, q_m) dv$$

est correcte.

Cette relation a une forme invariante et suffisamment vérifiée en coordonnées cartésiennes dans lesquelles G est la matrice unitaire et

$$D(q) = Q(q) = \left\{ \frac{\partial q^i}{\partial x_j} \right\}, \quad i, j = 1, 2, 3$$

nous avons

$$\int_{\Omega} \text{Tr}(D^*(q_n)D(q_m))dv = \sum_{i,j=1}^3 \int_{\Omega} \frac{\partial q_n^i}{\partial x_j} \frac{\partial q_m^i}{\partial x_j} dv = \\ = - \sum_{i,j=1}^3 \int_{\Omega} q_n^i \frac{\partial^2 q_m^i}{\partial x_j^2} dv = - \int_{\Omega} (q_m, \Delta q_m) dv$$

Nous verrons que dans notre cas les champs propres de l'opérateur de Laplace forment une base dans laquelle la matrice (M_{nm}) est diagonale. Cela signifie que la construction de la base $\{e_1, e_2, e_3, \dots\}$ des champs propres de l'opérateur de Laplace correspondant aux nombres $0 < \alpha_1 < \alpha_2$ est équivalente à la procédure décrite en 2.4.1. Nous obtenons ainsi que la base $\{e_1, e_2, \dots\}$ est ordonnée suivant la croissance du degré d'oscillation des champs e_i et $\alpha_i = \beta_i^2 (e_i)$.

3. Fermeture du système des fonctions propres et associées de l'opérateur du problème considéré.

Nous voudrions montrer dans ce chapitre que la base, construite dans le chapitre 2, peut être utilisée non seulement dans des buts de calculs mais aussi pour obtenir certaines déductions théoriques.

Nous rappellerons que si λ est la valeur propre de l'opérateur A alors ce vecteur s'appelle vecteur propre associé d'ordre m, soit

$$(A - \lambda I)^m v = 0, \quad (A - \lambda I)^{m-1} v \neq 0$$

Soit $R_\lambda(A) = (A - \lambda I)^{-1}$ la solution de l'opérateur A, S(A) est le système des fonctions propres et associées de cet opérateur. Nous montrons dans ce chapitre que le système S(A), où A est l'opérateur du problème (1.1) - (1.2) est complet en H. La démonstration s'appuie sur le théorème de Karleman dont la forme simplifiée est donnée dans l'ouvrage [5] comme suit :

Théorème (Karleman). Si A est l'opérateur résultant de la résolution de Hilbert-Schmidt et le long de chaque rayon $\lambda = re^{i\theta}$ (θ étant fixé) ne correspondent pas avec l'axe matériel $|R_\lambda(A)| = 0$ ($|\lambda|^{-1}$) pour de grands λ , alors le système S(A) est complet en H.

Le fait que nous voulons établir est la conséquence de résultats plus généraux (théorème de Keldich, voir par exemple [7]) applicable non seulement à l'opérateur donné. La formulation elle-même du théorème de Keldich exige cependant des déterminations et des explications complémentaires. L'approche proposée est élémentaire et donne comme il nous semble une certaine représentation de ces propriétés des opérateurs P et B dont il résulte que le système S(A) est complet.

Toutes les déterminations et les propriétés des opérateurs qui sont employées dans ce chapitre sans démonstration se trouvent dans l'ouvrage [5]. Au 3.1. on donne les références sur le théorème de Fredholm. Elles sont proposées dans beaucoup de livres (voir par exemple [6]).

3.1. Discrétilisation du spectre du problème examiné

3.1.1. Nous décrivons en détail la situation à laquelle nous avons affaire. Dans l'espace de Hilbert H avec une base matérielle orthonormalisée $\{e_n\}$ qui est décrit dans le § 2.4.1 on examine le problème spectral:

$$Aq \equiv Pq - \varepsilon Bq = \lambda q, \quad \varepsilon > 0. \quad (3.1)$$

Ici P est l'opérateur symétrique oblique $|P| = 2$. L'opérateur B est déterminé positif, dans la base $\{e_n\}$ sa matrice est diagonale avec les nombres $0 < \alpha_1 < \alpha_2 < \dots$ sur les diagonales. La région de la détermination de l'opérateur B

$$D(B) = \{x : x \in H, \sum |x_i \alpha_i|^2 < \infty\},$$

est évidemment étroite en H et $D(A) = D(B)$

L'opérateur B a B^{-1} pour inverse : $B^{-1}x = y$, où $y_i = x_i / \alpha_i$ (x_i, y_i sont les composantes de x et y dans la base $\{e_n\}$), $D(B^{-1}) = H$. Dans la base $\{e_n\}$ l'opérateur B^{-1} a une matrice diagonale avec les nombres $\alpha^{-1}_1 > \alpha^{-1}_2 > \dots$ sur les diagonales et est déterminé positif et compact.

3.1.2. Soit (λ, q) la paire propre pour (3.1). Alors $q = B^{-1}p$ pour un certain $p \in H$ et la paire (λ, p) satisfait à l'équation

$$T(\lambda)p = \varepsilon p, \quad (3.2)$$

où $T(\lambda) = (P - \lambda I)B^{-1}$ est l'opérateur compact pour n'importe quel λ (ici λ joue le rôle de paramètre, ε est la variable spectrale). Mais, par contre : si la paire (λ, p) satisfait à (3.2) alors $(\lambda, B^{-1}p)$ est la paire propre pour (3.1).

3.1.3. Ecrivons l'équation (3.2) sous la forme

$$(T(\lambda) - \varepsilon I)p = 0.$$

Maintenant à partir de la compacité de l'opérateur $T(\lambda)$ nous obtenons :

si (λ, q) est la paire propre pour l'équation (3.1) alors pour ce λ l'équation $(T(\lambda) - (\varepsilon I))p = 0$ a un nombre fini de solutions. Sa dimension est multiple géométrique de λ nombre propre de l'opérateur A .

Si pour une certaine équation en λ (3.1) il n'y a qu'une seule solution nulle, alors $\text{Ker}(T(\lambda) - \varepsilon I) = \{0\}$ et selon le deuxième théorème de Fredholm $(T(\lambda) - \varepsilon I)^{-1}$ est déterminé et borné sur tout H .

L'équation

$$(T(\lambda) - \varepsilon I)y = z \quad (3.3)$$

a une solution unique $\forall z \in H$.

3.1.4. Supposons que λ n'est pas une valeur propre de l'opérateur A . Alors pour n'importe quel $z \in H$ l'équation $(\lambda I - P - \varepsilon B)x = z$ a la solution $x = B^{-1}y$ où $y = (T(\lambda) - \varepsilon I)^{-1}z$ est la solution de l'équation (3.3). Cela signifie que l'opérateur $(\lambda I - P - \varepsilon B)^{-1}$ est borné et déterminé sur tout H , λ est un point régulier pour l'opérateur A dont le spectre, ainsi, n'a pas de partie continue.

Nous noterons que la discréttisation du spectre de l'opérateur A est démontrée sans utilisation de la symétrie oblique P et de la convertibilité

lité de A.

3.1.5. Si (λ, q) est la paire propre pour l'opérateur A, $|q| = 1$, alors $\lambda = (Pq, q) - \epsilon (Bq, q)$. En vertu des propriétés des opérateurs P et B il faut que $(Bq, q) \geq \alpha_1$, et (Pq, q) est un nombre purement imaginaire. Tout le spectre est ainsi réparti en demi bande
 $\Pi = \{\lambda : \operatorname{Re} \lambda \leq -\epsilon \alpha_1, |\operatorname{Im} \lambda| \leq |P|\}$. La valeur $\epsilon \alpha_1$ est une estimation inférieure grossière du module de la partie réelle voisine de zéro du nombre propre de l'opérateur A.

3.2. Vérification des conditions du théorème de Karleman

Nous montrons d'abord que la résolvante $R_\lambda(A)$ constitue l'opérateur de Hilbert-Schmidt.

3.2.1. Nous présenterons l'opérateur A sous la forme suivante (équivalente)

$$A = (PB^{-1} - \epsilon I) B.$$

L'opérateur PB^{-1} est compact, son spectre non nul est constitué des nombres propres. Si $PB^{-1}y = ty$ alors $(PB^{-1}y, B^{-1}y) = t(y, B^{-1}y)$. L'opérateur P est symétrique oblique et B^{-1} est déterminé positif. Il en résulte qu'il faut que tous les nombres propres $t \neq 0$ soient purement imaginaires. Cela signifie que $\epsilon > 0$ est point régulier pour l'opérateur PB^{-1} , $(PB^{-1} - \epsilon I)^{-1}$ est déterminé et borné dans tout H. Puisque B^{-1} est compact, $A^{-1} = B^{-1}(PB^{-1} - \epsilon I)^{-1}$ est aussi compact

3.2.2. L'opérateur compact T sera représenté sous la forme

$$Tx = \sum \gamma_n \psi_n(\varphi, x),$$

où γ_n sont les valeurs propres positives de l'opérateur $(T * T)^{1/2}$, ϕ_n sont les vecteurs propres correspondants et $\psi_n = T\phi_n/\gamma_n$. Les deux systèmes $\{\phi_n\}$ et $\{\psi_n\}$ sont orthonormés. Si $\sum (\gamma_n)^2 < \infty$, alors l'opérateur T s'appelle opérateur de Hilbert-Schmidt. Si T est auto-conjugué alors il est évident que γ_n sont ses valeurs propres, $\phi_n = \psi_n$ sont les vecteurs propres correspondants.

Pour la facilité nous désignerons par la lettre Γ la classe des opérateurs Hilbert-Schmidt.

3.2.3. L'opérateur A^{-1} est fermé par conséquent, et inversement, l'opérateur A est également fermé. Pour ces opérateurs "une équation résoluble" est évidemment:

$$R_\lambda(A) - R_\mu(A) = (\lambda - \mu)R_\lambda(A)R_\mu(A) \quad (3.4)$$

pour tous les λ, μ issus de la quantité résolue $\rho(A)$. De cette relation il s'ensuit que si $R_\mu(A) \notin \Gamma$ pour $\mu = 0$ alors $R_\lambda(A) \notin \Gamma$ pour tous les $\lambda \in \rho(A)$ (le terme de droite en (3.4) est le produit de $R_0(A) \in \Gamma$ par l'opérateur limité et ce produit entre également en Γ).

L'opérateur $R_0(A) = A^{-1} = B^{-1}(PB^{-1} - \epsilon I)^{-1}$ appartient à Γ si $B^{-1} \in \Gamma$. L'opérateur B^{-1} est symétrique et de 3.2.2 il s'ensuit que $B^{-1} \in \Gamma$ si $\sum (1/\alpha_n)^2 < \infty$. Mais les nombres propres de l'opérateur B sont les nombres (λ_{ik}) et $(\tilde{\lambda}_{jm})$ correspondant aux champs propres toroïdaux et poloïdaux (voir 2.2.1).

et 2.2.2) c'est pourquoi $B^{-1} \notin \Gamma$ si les séries $\Sigma_{i,k} (1/\lambda_{ik})^2$ et $\Sigma_{j,m} (1/\tilde{\lambda}_{jm})^2$ se combinent.

3.2.4 On sait (voir par exemple [5] chapitre 11) que dans une région limitée (dans un espace à trois dimensions) l'opérateur de Laplace sur les fonctions scalaires égales à zéro à la limite est en sens inverse l'opérateur de Hilbert-Schmidt (opérateur intégrale avec une faible singularité). D'où (voir 2.2.2) il résulte que $\Sigma_{ik} (1/\lambda_{ik})^2 < \infty$.

La convergence de la série $\Sigma_{jm} (1/\tilde{\lambda}_{jm})^2$ découle des réflexions suivantes. L'opérateur $B^{-1} = (\text{rot rot})^{-1} = (\text{rot})^{-1} (\text{rot})^{-1}$. L'opérateur $(\text{rot})^{-1}$ traduit les champs toroïdaux en poloïdaux, poloïdaux et toroïdaux (cela résulte de la formule (2.4) pour l'opérateur rot). Dans la base des champs propres des types toroïdal et poloïdal (dans cet ordre !) la matrice de l'opérateur $(\text{rot})^{-1}$ a la forme $\begin{pmatrix} O & U \\ W & O \end{pmatrix}$ et B^{-1} a la matrice $\begin{pmatrix} UW & O \\ O & WU \end{pmatrix}$.

Les matrices UW et WU sont diagonales, UW a sur les diagonales les nombres (λ^{-1}_{ik}) , WU - sont les nombres (λ^{-1}_{jm}) . Par la diagonalité de ces matrices et des propriétés de l'opération Trace nous obtenons

$$\sum_{i,k} (1/\lambda_{ik})^2 = \text{Tr}(UWUW) = \text{Tr}(WUWU) = \sum_{j,m} (1/\tilde{\lambda}_{jm})^2.$$

La série $\Sigma_{jm} (1/\tilde{\lambda}_{jm})^2$ se combine ainsi, $B^{-1} \in \Gamma$. D'où il résulte (voir 3.2.3.) que $R_\lambda(A)$ est l'opérateur de Hilbert-Schmidt $\forall \lambda \in \rho(A)$

3.2.5. Nous vérifierons la condition du théorème de Karleman.
Soit $\lambda \in \rho(A)$ et $|Im \lambda| > |P|$. Pour n'importe quel $y \in H$ l'équation

$$(A - \lambda I)x = (P - \varepsilon B - \lambda I)x = y$$

a pour solution $x = R_\lambda(A)y$. En prenant la partie imaginaire de l'équation

$$(Px, x) - \varepsilon(Bx, x) - \lambda(x, x) = (y, x),$$

nous obtiendrons

$$(Px, x) - Im \lambda |x|^2 = Im(y, x),$$

$$|Im \lambda| |x|^2 \leq |P| |x|^2 + |x| |y|,$$

c'est à dire

$$|x| = |R_\lambda(A)y| \leq |y| \frac{1}{(|Im \lambda| - |P|)}.$$

Cela signifie que pour $|Im \lambda| > |P|$

$$|R_\lambda(A)| \leq \frac{1}{|Im \lambda| - |P|}.$$

nous savons (voir 3.1.5) que tout le spectre de l'opérateur A est compris dans la demi bande Π de largeur $2|P|$. C'est pourquoi n'importe quel rayon $\lambda = re^{i\theta}$ ne correspond pas avec l'axe matériel (θ est fixé) soit qu'il se trouve en entier dans la quantité résolue, soit qu'il sort de celle-ci pour

$$|Im\lambda| = r|\sin\theta| > |P| \text{ et}$$

et

$$|R_\lambda(A)| \leq \frac{1}{r|\sin\theta| - |P|} = O\left(\frac{1}{|\lambda|}\right)$$

pour $\lambda \rightarrow \infty$ le long du rayon.

La vérification de la réalisation des conditions du théorème de Karleman est terminée. Le système des fonctions $S(A)$ est ainsi complet dans l'espace H .

Conclusion

Nous avons ainsi, dans l'espace des champs indivergenciels, dans la couche sphérique de deux bases ordonnées d'après le degré d'oscillation: l'une est constituée des champs avec la condition $(q, n) = 0$ à la limite établie dans le travail [10]; la seconde est constituée des champs tendant vers zéro à la limite, dans ce travail. Nous considérons que les deux constructions peuvent être utilisées pour les solutions approchées des problèmes d'hydrodynamique dans une couche sphérique (pas nécessairement reliés avec le liquide en rotation). Les auteurs espèrent que la démonstration du système des fonctions $S(A)$ est complet peut également présenter un intérêt pour les lecteurs.

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BIBLIOGRAPHIE

1. Гринспен Х.П. Теория вращающихся жидкостей. Л.: Гидрометеоиздат, 1975. 303 с.
2. Владимицов В.С. Уравнения математической физики. М.: Наука, 1981. 512 с.
3. Chandrasekhar S. Hydrodynamic and hydromagnetic stability. Oxford University press, 1961. 654 p.
4. Морфат Г. Возбуждение магнитного поля в проводящей среде. М.: Мир, 1980. 339 с.
5. Рихтмайер Р.. Принципы современной математической физики. т.1, М.: Мир, 1982. 486 с.
6. Колмогоров А.Н., Фомин С.В. Элементы теории функций и функционального анализа. М.: Наука, 1981. 542 с.
7. Копачевский Н.Д., Крейн С.Г., Нго Зуй Кан. Операторные методы в линейной гидродинамике. М.: Наука, 1989. 416 с.
8. Юдович В.И. Метод линеаризации в гидродинамической теории устойчивости. Изд-во Ростовского Гос. Университета, 1984. 192 с.
9. Справочник по специальным функциям / Под ред. М.Абрамовича и И.Стигана. М.: Наука, 1979. 830 с.
10. Резников Е.Л., Розенкрапп Л.М. О гладких приближенных собственных мод оператора Пуанкаре в шаровом слое // Теоретические проблемы геодинамики и сейсмология. М.: Наука, 1994. С.70-85. (Вычисл. Сейсмология, Вып.27).

A NEW PROGRAM DEVELOPED IN MADRID FOR TIDAL DATA PROCESSING

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Abstract

This paper is a description of a program named NSV for tidal data processing. Despite the considerable volume of the paper, it deals only with the most important elements of NSV, including its practical use. It is not possible to describe everything, even all algorithms, used in a program of several thousands statements, having a lot of different options and applications.

In the acronym NSV, SV is simply the name of an old program, while N means "New". The "New" itself means that we use an old algorithm, which has proven its validity, but through a new program, which has a lot of new elements. It is also related with the necessity a scientific program to be steadily improved, which we are namely doing with NSV.

The paper has two parts. In Part I more attention is given to the theory of NSV. People, more interested in the applications, may go directly to Part II, which deals with the practical use of NSV.

I. Basic algorithm of NSV.

I.1. General remarks about the method of analysis used by NSV.

The main function of NSV is tidal analysis by a method which will be referred as MV66. Its original version can be found in (Venedikov, 1966a,b, Melchior & Venedikov, 1968, Melchior, 1978, 1981). In NSV are also used ideas and algorithms or various kinds of processing of Venedikov (1979, 1981, 1984, 1986), Venedikov & Ducarme (1979), Simon et al. (1989), Toro et al. (1990, 1991, 1993a), Fernández et al. (1993) as well as suggestions made by Baker (1978a,b), Meyer (1980) and others.

MV66 is applied since a lot of time, successfully surviving a tremendous development of the computers, a considerable increase of the observation precision, as well as the appearance of more and more sophisticated tasks. This makes necessary to refresh the information about this method. Therefore we shall give the main principles of MV66 in some of the next Sections. We shall try to do this in a way which can be better understood than before.

However, NSV is not an application of a standard MV66. For example, in MV66 are given numerically fixed filters, while NSV is building up the filters at every application, and this can be done in many different ways. Actually, NSV is a sophisticated computational device through which, we hope, the processing can become an interesting research work in two directions:

- (i) the user can apply MV66 in many variants, actually creating and testing different methods of analysis and
- (ii) he can use various options which allow to study the data, looking for perturbations and particular phenomena.

Of course, despite the statement (i) above, the user cannot escape from the fundamental scheme of MV66, whose elements are:

- (a) model of the tidal signal which involves the grouping of the tides and the use of new (at the time) unknowns (Venedikov, 1961, 1966a) directly related with the unknown tidal parameters,
- (b) filtration of intervals of length n , e.g. $n = 48$, without overlapping,
- (c) processing of the filtered numbers by MLS (the Method of the Least Squares).

The model mentioned in (a) seems to be completely justified. It has been employed in most of the Earth tide methods of analysis, e.g. in the methods of Chojnicki (1973), Schüller (1977) and ETERNA of Wenzel (1994a,b), as well as in Tamura et al. (1991), Ishiguro et al. (1983).

It can easily be shown that our model is equivalent to the model of the ocean tidal phenomena of Munk and Cartwright (1966), discussed also by Meyer (1982). Hence, any good program for Earth tide data analysis, using our model, including NSV, can be applied on ocean tidal data.

The computational scheme (b) and (c) is more controversial. The matter is that it can be conceived as a sampling of the data over a step of $n = 48$ hours. Such a sampling is of course illegal. It would be absolutely funny to expect to find waves of frequencies concentrated around 1 and 2 cpd (cycle per day) using data with a step 2 days.

The elementary proof that (b) and (c) are not a simple sampling is that MV66 has provided a huge amount of useful results and we do not know about any failure. Another one is that all other reasonable methods provide results very close to those of MV66. In this sense, although these methods have been created as negations of MV66, all of them are in favour of MV66.

In relation with this problem, the flexibility of NSV, about which is said in (i) and (ii) can be useful. Unlike the original version of MV66, NSV allows to use various filters. Thus the length of the filters $n = 48$ can be replaced by any even number of the same order, e.g. $n = 36, 40, 42, 50, \dots$. It is also possible to use a shift $s \neq n$. Although $s < n$ is theoretically unacceptable, it is experimentally interesting to use such values. Even $s = 1$ is available, which is totally transforming our way of filtration into a moving filtration, similar to Chojnicki and ETERNA.

Other examples of flexibility is that one can change and test the grouping of the tides, the model of the drift, to determine the LP tides in many variants, to look for relations with non-tidal data and so on.

It is also interesting that through NSV can be computed residuals, drift, amplitude and phase variations and search for new tidal and non-tidal frequencies in different variants.

1.2. General model of the tidal data and model of the tidal signal.

Let $y(t)$, $t = 1, 2, 3, \dots$ are hourly tidal data or observations, t being the time in hours. Almost all large records have gaps. Therefore the sequence $t = 1, 2, 3, \dots$ should be understood as having jumps at the time of the gaps.

We would like to recall that one of the merit of MV66 is that it takes the data such as they are, with gaps and arbitrary length, rejecting any production of artificial data through interpolation. In such a way we have followed the principle to adjust the algorithm to the data and not the data to the algorithm.

In principle, an analysis of $y(t)$ through the Method of the Least Squares (here and further MLS) consists in the following.

We have first to create a model $M(t; x_1, x_2, \dots, x_m)$ of $y(t)$, which is a function of t and some unknown parameters x_1, x_2, \dots, x_m . Then writing

$$(1) \quad y(t) = M(t; x_1, x_2, \dots, x_m), t = 1, 2, \dots$$

we get a system of equations of number equal to the number of the data, in unknowns x_1, \dots, x_m . Finally, the system (1) is solved by the MLS. The use of MLS allows to get results with known optimum qualities and, which is very important, though sometimes neglected, with a well defined estimation of the precision.

Generally, (1) has the expression

$$(2) \quad y(t) = w_t + d_t + a_t + \varepsilon_t$$

where w_t is the tidal signal, d_t is the drift, a_t , are perturbations due to some phenomena and ε_t is the noise.

In expression (2), w_t is the most interesting component and can be called the useful signal in the data $y(t)$.

The remaining terms, with respect to w_t , can be considered as noise, having both random and deterministic components. However, for some purposes, The second component d can also be considered as useful signal, which has to be well estimated and studied.

The third component a can be due to different meteorological phenomena, among which the air-pressure seems to be the most important. Within this Part I it will be ignored. However, in Part II we shall briefly consider the model of a used by NSV.

The last term ε_t is random component. We know that it is correlated, i.e. ε_t is not a WN (white noise). One of the reason for the correlation is that we are not able to create good enough models for d_t and a_t , thus charging ε_t by the errors of these models. In the same time we do not dispose by any good estimation of the autocorrelation function or the covariance matrix of ε_t . This in principle does not allow us to apply rigorously MLS directly on the hourly data $y(t)$.

It is very natural to begin here with the model of w_t .

In any case w_t is generated by a theoretical signal like

$$(3) \quad W_t = \sum_{\omega} H_{\omega} \cos(\phi_{\omega} + \omega t)$$

where ω takes a set of *known* discrete values

$$\omega = \omega_1, \omega_2, \omega_3, \dots$$

H_{ω} is *known* theoretical amplitude and

ϕ_{ω} is *known* theoretical phase at $t = 0$

In Figure 1.a and 1.c is represented how looks the amplitude spectrum (H_{ω} as a function of ω) of W_t in the main D and SD domains of ω . Each line of this spectrum is corresponding to a term in (3), i.e. to a tidal wave or tide of frequency ω (here and further under frequency we understand angular frequency or velocity).

If W_t is represented by (3), the generated or observed signal w_t is

$$(4) \quad w_t = \sum_{\omega} \delta_{\omega} H_{\omega} \cos(\phi_{\omega} + \omega t + \kappa_{\omega}) = \sum_{\omega} h_{\omega} \cos(\varphi_{\omega} + \omega t)$$

where:

δ_{ω} is *unknown* observed amplitude factor,

κ_{ω} is *unknown* observed phase shift with respect to ϕ_{ω} ,

$h_\omega = \delta_\omega H_\omega$ is *unknown* observed amplitude and
 $\varphi_\omega = \phi_\omega + \kappa_\omega$ is *unknown* observed phase.

For the Earth tide data this expression has theoretical and empirical basis, supported by an abundant experience. More generally, including the case of ocean data, W_t can be considered (Munk et al., 1966), as an input to a system (Earth and ocean) with output w_t . Provided the system is linear, the output can be represented through (4). In this sense (4) as well as the model of w_t here considered can be used for both Earth and ocean tidal data. We have already a lot of experience in the application of our programs on ocean data (Carvajal, 1993).

The tidal part in (2) becomes very simple if (4) is represented as

$$(5) \quad w_t = \sum_{\omega} (u_{\omega} \cos \omega t + v_{\omega} \sin \omega t)$$

$$u_{\omega} = h_{\omega} \cos \varphi_{\omega}, \quad v_{\omega} = -h_{\omega} \sin \varphi_{\omega}$$

which is linear with respect to the unknowns u_{ω} and v_{ω} . However, a serious complication is that we have too many ω , some of which differ very little, i.e. the components of (5) are linearly dependent. E.g., in the development of Tamura are 1200 waves, i.e. 2400 unknowns and still more in (Xi Quinwen) and other new developments.

A solution of the problem is suggested through our model (Venedikov, 1961, 1966a).

The idea is to realise that the actual physical unknowns are not the amplitudes and the phases, h_{ω} and φ_{ω} , but the parameters δ_{ω} and κ_{ω} . These parameters define completely the output (4), i.e. they carry the total information about the phenomena.

There are theoretical and empirical evidences that δ_{ω} and κ_{ω} do not vary very fast with ω . Hence, the actual number of different unknown quantities is not the huge number of h_{ω} and φ_{ω} , but much lower.

This idea is applied in the following way.

As shown in Figures 1.a,c, there is a concentration of the energy in some bands or grouping of the tides around a small number of frequencies of some more important tides. These bands or groups can be defined through frequency intervals, say $\Omega_1, \Omega_2, \dots, \Omega_{\mu}$. In Figures 1.a,c a variant of such intervals are demonstrated. Each Ω_j there has a name which is the symbol of Darwin of the most important tide in the corresponding interval.

As far as Ω_j are narrow, we may suppose $\delta_{\omega} = \text{const.}$ and $\kappa_{\omega} = \text{const.}$ for $\omega \in \Omega_j$. Then, in order to get linear expressions, we define the unknowns

$$(6) \quad \left. \begin{array}{l} \xi_j = \delta_{\omega} \cos \kappa_{\omega} = \text{const.} \\ \text{and} \\ \eta_j = -\delta_{\omega} \sin \kappa_{\omega} = \text{const.} \end{array} \right] \quad \text{for } \omega \in \Omega_j, j = 1, 2, \dots, \mu$$

Using this, (5) is transformed in the following simple expression

$$(7) \quad w_t = \sum_{j=1}^{\mu} (c_j \xi_j + s_j \eta_j)$$

with coefficients at the unknowns

$$(8) \quad c_{ij} = \sum_{\omega \in \Omega_j} H_\omega \cos(\phi_\omega + \omega t), \quad s_{ij} = \sum_{\omega \in \Omega_j} H_\omega \sin(\phi_\omega + \omega t)$$

NSV is using a generalisation of this model. Namely, given Ω_j is not obligatory an interval, but, more generally, it is a set of frequencies ω . This means that we can extract some tides from a group corresponding to an interval and let them shape separate groups.

An example are the waves which we shall call $P3$ whose origin is the tidal potential of 3d degree. These are small waves whose spectrum is shown in Figures 1.b,d. The inconvenience of their existence is that if these waves remain in one and the same Ω_j with other waves, the condition (6) is theoretically false.

An option of NSV is to take apart the $P3$ and let them create groups like those in Figures 1.b,d. Somewhat surprisingly, it appeared possible to separate such groups, i.e. to determine the corresponding parameters δ and κ , although they are overlapping other D and SD groups.

The practical use of NSV will be discussed in Part II. Nevertheless, it may be interesting how simply this option of NSV is used.

At the first stage of work of NSV it sets a series of questions. One of them is

2.6. (SEPP3) SEPARATE THE TIDES OF DEGREE 3 ? y

If the answer is "y" (yes) as above, the $P3$ are determined separately. If the answer is negative, the $P3$ are left in their natural groups. Then NSV is applying a theoretical modulation of the amplitudes of $P3$. When 2.6. (SEPP3) is accepted this modulation is excluded.

Following this definition of Ω_j , NSV allows to pick up some other tides, e.g. given meteorological tide like S1, S2,...

1.3. Model of the drift.

Due to the model discussed in Section I.2, we dare say that the most interesting component, the useful tidal signal w_t , is the least problem of the tidal data processing. Much more serious problems create the remaining components in (2), especially the drift d_t .

In the following $I_i = I(T_i)$ will denote an interval of the data with central epoch T_i and I -s will mean a set of I_i .

A natural solution of the drift problem, which seems (but only seems) to be good, is that of (Horn, 1959): d_t is represented, over a single I_i of length 732^h, as a polynomial of t . Wenzel (1976a) has suggested the drift to be approximated in I -s, named blocks, which may be defined by the gaps in the data, i.e. also in large I -s. However, there are not evidences that such approximations can be successful.

Our attempts to approximate d_t in large I -s, have shown that such approximations are not successful, even using polynomials of very high powers. The matter is that the approximation of d_t in an I_i by a polynomial of power k implies the supposition that d_t has not continuous derivatives of the same order. Since d_t is very often changing its behaviour, this supposition is not justified for large I -s.

In Figure 2 is shown an attempt to approximate the drift in a large interval by a polynomial, using the analysis method (Toro et al., 1993b). The power $k = 17$ is not arbitrarily selected. It is chosen using the AIC criterion of Akaike (Sakamoto et al., 1986). We get residuals which are obviously not satisfying.

In Figure 3 is given the approximation of the drift in the same data, but in separate I -s of length 48^h, through the option RESAN of NSV. The residuals now are incomparably better than in Figure 2.

In this relation NSV is applying an approximation of d_t in short I -s in the following way.

The whole record is partitioned in I -s of equal length n , say $n = 48$. In principle, the I -s are without overlapping, i.e. the shift between I_i and I_{i+1} is $s = n$ or, if there is a gap, $s > n$. Optionally, for experimental purposes, one can use $s > n$ (without gaps) or $s < n$, i.e. intervals with overlapping.

The default value of n is $n = 48$, which has been most frequently used in the applications of MV66. As said above, NSV can use other values of n . Nevertheless, we shall sometime use the number 48, as if it a fixed one, but it should be understood that optionally we can use other values.

Within every I_i is used a time t measured from T_i , i.e.

$$(9) \quad \begin{aligned} \tau &= t - T_i = -v, -v + 1, \dots, v, & v &= (n - 1)/2 \\ &\text{e.g.} \\ \tau &= -23.5, -22.5, \dots, +23.5 & \text{for } n &= 48 \end{aligned}$$

The model of d_t in I_i is accepted to be

$$(10) \quad d_t = \zeta_{i0} + \zeta_{il}(\tau/v) + \dots + \zeta_{ik}(\tau/v)^k, \quad \tau = t - T_i, \quad T_i - v \leq t \leq T_i + v$$

where, it is very essential, ζ_{il} , $l = 0, 1, \dots, k$ are unknowns specific for given I_i . This is indicated by the subscript i of ζ_{il} .

Between the models of d_t in the neighbouring I_i and I_{i+1} are not imposed any conditions, although theoretically we should have equal values at time $T_i + n/2$ and $T_{i+1} - n/2 = T_i + n/2$ (if there is not a gap). Our explanations are (i) we do not use d_t at just $T_i + n/2$, because $v = n - 0.5$ and (ii) the lack of conditions allows arbitrary discontinuities between I_i and I_{i+1} . If there are important discontinuities within given I_i , they do not affect the global approximation of d_t in the other I -s. If the situation in given I_i is too bad, NSV allows to distinguish this and eliminate such an I_i .

In connection with this model and our partition into I -s it is convenient to use some vectors of size or dimension n .

Let $c(\tau)$ is a function of τ , e.g. $c(\tau) = \cos \omega\tau$. We shall denote by \mathbf{c} the vector column (calling it only vector) of the values of $c(\tau)$ in any I_i

$$(11) \quad \mathbf{c} = [c(\tau)] = [c(-v) \ c(-v + 1) \ \dots \ c(v)]^*$$

where * denotes the transposition of a matrix.

In the case of $y(t) = y(T_i + \tau)$ and $d(t) = d(T_i + \tau)$ we shall use the vectors

$$(12) \quad \begin{aligned} \mathbf{y}_i &= \mathbf{y}(T_i) = [y(T_i + \tau)] \\ \mathbf{d}_i &= \mathbf{d}(T_i) = [d(T_i + \tau)] \end{aligned}$$

whose elements are the values of $y(t)$ and $d(t)$ in given I_i . Here τ is taking the values (9) while T_i and the subscript i remain fixed, indicating the relation with given $I_i = I(T_i)$.

Let, according to (11),

$$(13) \quad \mathbf{p}_i = [(\tau/v)^l]$$

Then the model (10) of the drift can be represented in the vector form

$$(14) \quad \mathbf{d}_i = \mathbf{d}(T_i) = \sum_{l=0}^k \zeta_{il} \mathbf{p}_l$$

In NSV for intervals of $n \geq 48$ the default value of k is $k = 2$ and, if $n \leq 48$, $k = 1$. If LP (long period) waves are determined, using large I -s, e.g. $n = 360$, $k = 1$. If LP are determined in parallel with the D or SD waves, k is set to be $k = 0$.

Of course, k is optional and can be modified if the question:

3.1. (NFINT)NEW POWER k OF THE DRIFT POLYNOMIALS ?

gets an answer "y". Then the user is asked to enter the value of k .

At the end of this discussion of d , we would like to point out that a very interesting model is suggested in (Tamura et al., 1991, Ishiguro et al., 1983).

I.4. The stage of filtration of NSV.

NSV is proceeding the determination of the main species D, SD,... together, at one and the same time, but in parallel, independently from each other. Thus in the first stage mentioned in (b), Section I.1, we apply different filters for every species. In this relation in the following we have in mind filtration and processing related with one of the species.

More concretely, the first stage consists in the application of one even and one odd filter. The filters can be considered as vectors, according to (11), namely

$$(15) \quad \begin{aligned} \mathbf{f} &= [f_\tau], & f_\tau &= f_{-\tau} \text{ (even filter)} \\ \mathbf{g} &= [g_\tau], & g_\tau &= -g_{-\tau} \text{ (odd filter), } \tau = -v, -v+1, \dots, v \end{aligned}$$

where f_τ and g_τ are the coefficients of the filters.

The filtration consists in the computation of the filtered numbers

$$(16) \quad \begin{aligned} u(T_i) &= \mathbf{f}^* \mathbf{y}_i = \sum_{\tau=-v}^v f_\tau y(T_i + \tau) \\ v(T_i) &= \mathbf{g}^* \mathbf{y}_i = \sum_{\tau=-v}^v g_\tau y(T_i + \tau) \end{aligned}$$

related with I_i . We have an application of \mathbf{f}, \mathbf{g} on the I -s. Thus the filters are applied with a shift $s \geq n$, unlike the usual moving filtration when a filter is applied with $s = 1$.

In the second stage of the analysis, the set of $u(T)$ and $v(T)$ are processed by MLS, instead of the hourly data $y(t)$. This is namely seeming to be a sampling with a step $n = T_{i+1} - T_i$. Therefore this procedure needs our explanations.

Let $y(t)$, $t = 1, 2, \dots, N$ are N data without gaps. If $y(t)$ can be submitted to a Fourier analysis, we have to compute, for given Fourier frequency $0^\circ < \omega < 180^\circ$,

$$(17) \quad x(\omega) = (2 / N) \sum_{t=1}^N \text{Exp}(i\omega t) y(t) \quad (i^2 = -1)$$

Let $N = qn$ and the data are partitioned in our I -s I_i of length n . Just the same $x(\omega)$ can be computed in the following 2 stages.

(i) We compute the filtered numbers (16) using as filters

$$(18) \quad \mathbf{f} = [(2/n)\cos\omega\tau] \quad \text{and} \quad \mathbf{g} = [(2/n) \sin \omega\tau]$$

which will be called FFIL (Fourier Filters).

(ii) Then, using the filtered numbers (16), we compute

$$(19) \quad x(\omega) = (1/q) \sum_{i=1}^q \text{Exp}(i\omega T_i) [u(T_i) + iv(T_i)]$$

which is identical to $x(\omega)$ in (17).

Thus we have, rather similarly to NSV: stage (i) filtration of intervals I_i without overlapping, and stage (ii) processing of the filtered numbers which is a Fourier analysis of data with a step n hours.

If only (19) is considered, this would seem as an illegal analysis of a sampled data over a step n hours. On the contrary, if the nature of the filters is considered, we have a legal direct processing of hourly data.

By the way, the computations after (19) is much faster than (17), i.e. more computing time is not always necessary for getting better results.

Let us denote

$$(20) \quad \mathbf{c}(\omega) = [\cos \omega\tau] \quad \text{and} \quad \mathbf{s}(\omega) = [\sin \omega\tau]$$

The response of the filters \mathbf{f}, \mathbf{g} to a frequency ω is

$$(21) \quad \rho_f(\omega) = \mathbf{f}^* \mathbf{c} \quad \text{and} \quad \rho_g(\omega) = \mathbf{g}^* \mathbf{s}$$

If $\rho(\omega) = 0$, the corresponding ω (wave of frequency ω) is eliminated and, if $\rho(\omega) \neq 0$, ω is retained or reduced by a factor $\rho(\omega)$. Otherwise, ω is amplified by a factor $\rho(\omega)$.

If $\mathbf{f}^* \mathbf{p}_1 = \mathbf{g}^* \mathbf{p}_1 = 0$ for the vectors in (14), we have an elimination of the d_t in all I_i , as far as (14) is a good model of d_t .

Let $\mathbf{f} = \mathbf{f}_D$, and $\mathbf{g} = \mathbf{g}_D$ are created through (18) for $\omega = \omega_D = 15^\circ$ and $n = 48$, ω_D being one of the main D frequencies. Then we have

$$(22) \quad \rho_f(0) = \mathbf{f}_D^* \mathbf{p}_0 = \rho_g(0) = \mathbf{g}_D^* \mathbf{p}_0 = 0,$$

i.e. \mathbf{f}_D and \mathbf{g}_D eliminate a drift represented by a constant which can arbitrarily changing its values.

Further

$$(23) \quad \begin{aligned} \rho_f(15) &= \rho_g(15) = 1 \\ \rho_f(30) &= \rho_g(30) = 0 \\ \rho_f(45) &= \rho_g(45) = 0 \end{aligned}$$

Since all D tides have $\omega \approx 15^\circ$, SD tides - $\omega \approx 30^\circ$ and all TD tides - $\omega \approx 45^\circ$, we can say that the filters \mathbf{f}_D and \mathbf{g}_D amplify the D tides and eliminate or retain the SD and the TD species.

If we build up filters using once $\omega_{SD} = 30^\circ$, then $\omega_{TD} = 45^\circ$, which are also Fourier frequencies for $n = 48$, we shall get, together with \mathbf{f}_D and \mathbf{g}_D , 3 pairs of filters which, actually, separate the main tidal species, eliminating a model of the drift of very low power $k = 0$.

It appears that in the Fourier analysis we have a first stage consisting in application of filters which are tending to eliminate the drift (however, only as a constant in every I_i) and separate the main tidal species.

If there is a drift and the sophisticated tidal spectrum, we have to apply MLS instead of Fourier. In (Venedikov 1964b) is shown that the application of MLS directly on $y(t)$ has also the stage filtration of I_i . We have also FFIL but they have to be orthogonalized with respect to the drift polynomials. I.e. we have filters which eliminate the drift with the least possible deviations from FFIL. From this stage we get $u(T)$, $v(T)$ which are further processed.

In (Venedikov, 1964a) and (Wenzel, 1976a,b, 1977, Chojnicki, 1978, Meyer, 1980) has been established that the hourly $y(t)$ are correlated (non-WN or coloured noise). On the contrary, for various reasons, one of them the distance n between our I_i , $u(T_i)$, $v(T_i)$ are not correlated or considerably less correlated than $y(t)$. Follows the natural idea the second stage to be made, as said in (c) in Section I.1, i.e. to process $u(T)$, $v(T)$ by MLS as observations instead of $y(t)$.

Again in (Venedikov, 1964b) is shown that in this way we shall get estimates of the unknowns very close to the direct processing of $y(t)$. In the same time we shall have a more correct application of MLS with a rigorous estimation of the precision. The latter will reflect the effect of the drift and meteorological perturbations, which are frequency dependent, i.e. different for the main tidal species.

I.5. Construction of the filters by NSV.

The main idea is to get the filters as operators, solving equations about the data $y(t)$ in one, whichever I_i , according to the MLS. This idea has been applied in (Venedikov, 1966b) and further developed in (Venedikov, 1984) and NSV.

It has been said that the expression (5) cannot be used because there are too many waves. However, if we remain within a given short I_i , with $n = 48$, (5) can approximate the tidal signal using a very small number of the most important waves. Taking into account this and the model (14), we may use the following equations in vector form

$$(24) \quad \sum_{\omega} [u_{\omega} c(\omega) + v_{\omega} s(\omega)] + \sum_{l=0}^k \zeta_{il} p_l = y_i$$

for the data in anyone fixed I_i , where ω takes a limited number of values, e.g. $\omega = 15^\circ$, 30° and 45° .

The solution of (24), according to MLS, can be represented as

$$(25) \quad u_{\gamma}(T_i) = f_{\gamma}^* y_i, \quad v_{\gamma}(T_i) = g_{\gamma}^* y_i$$

where γ is one of the frequencies ω in (24) and f_{γ} , g_{γ} are vectors obtained through the orthogonalization of $c(\gamma)$ and $s(\gamma)$ respectively, with respect to all other vectors in (24). This means that f_{γ} , g_{γ} are filters which remain as close as possible to the FFIL, the deviations being imposed by the presence of the drift. Due to the orthogonal properties, f_{γ} , g_{γ} will amplify the corresponding $\omega = \gamma$ and eliminate all $\omega \neq \gamma$, as well as the model representing the drift.

Let first $\gamma = \omega_D$, then $\gamma = \omega_{SD}$ and, finally, $\gamma = \omega_{TD}$. If the subscript γ in (25) is replaced by 1, 2 and 3 corresponding to D, SD and TD, as well as to 1 cpd, 2cpd and 3 cpd, in the same way as above we shall get 3 pairs of filters f_j , g_j through which we shall get 3 sets of filtered numbers which can be used

- (26) $u_1(T_\nu), v_1(T_\nu)$ - for the determination of the D tides,
 $u_2(T_\nu), v_2(T_\nu)$ - for the determination of the SD tides,
 $u_3(T_\nu), v_3(T_\nu)$ - for the determination of the TD tides.

Obviously, all this can be expanded for the determinations of tides of higher frequencies like 4, 5 and 6 cpd.

Since the construction of the filters is related with the equations (24), we have to select the components, taking part there and, of course the number n which is the length of the I -s, as well as of the filters.

In NSV the default option is

- (27) $n = 48$ as length of the filters and I -s,
 $k = 2$ for the model of the drift,
S1, O1, S2, M2, S3, S4, S5, S6 as tidal components.

The set of tides indicated here means that ω in (24) takes the values of the frequencies of these waves, i.e. $\omega = \omega(S1), \omega(O1), \dots, \omega(S6)$. Obviously to each tide or ω are corresponding two vectors $c(\omega), s(\omega)$. We shall denote the couple (c, s) corresponding to a tide by the symbol of the corresponding tide.

In order to get a good approximation and separation it may be necessary to deal with 2, even more than 2 tides of one and the same species, e.g. S1 and O1. Then (24) cannot be directly solved. Therefore NSV first makes an orthogonalization of such components, after which (24) is solved. E.g. if in (24) are included S1, O1, Q1 then O1 is made orthogonal to S1 and Q1 - orthogonal to both S1 and O1.

In NSV the filters are obtained in the following way, which is an ordinary application of MLS on the data of any I_i .

Let C is the matrix of all vectors taking part in (24). Then (24) can be represented as

$$(28) \quad Cx_i = y_i$$

where x_i is the vector of all unknowns in (24). It is essential, that C is one and the same for all I_i , while x_i is different, depending on y_i .

The MLS solution, under the assumption of WN, is given by the estimates

$$(29) \quad \tilde{x}_i = (C^*C)^{-1}C^*y_i$$

If (27) is used, C is created by the following vectors or components

$$(30) \quad \begin{aligned} C &= [p_0 \ p_1 \ p_2 \ S1 \ O1 \ S2 \ M2 \ S3 \ S4 \ S5 \ S6] \\ &\quad \downarrow \qquad \downarrow \qquad \downarrow \qquad \downarrow \qquad \downarrow \qquad \downarrow \\ F &= [f_1, g_1 \quad f_2, g_2 \quad f_3, g_3 \quad f_4, g_4 \quad f_5, g_5 \quad f_6, g_6] \end{aligned}$$

where each tide should be understood as the couple of the corresponding vectors $c(\omega)$ and $s(\omega)$. According to what was said above, here O1 and M2 are only components of O1 and M2, orthogonal to S1 and S2 respectively.

NSV, in relation with (29) creates the matrix $F = C(C^*C)^{-1}$. It is composed by vectors, corresponding to the vectors in C , some of them shown in (30).

Since $\mathbf{F}^* \mathbf{C} = \mathbf{I}$ (identity matrix), every couple f, g in \mathbf{F} are filters amplifying the corresponding tide and eliminating all other tides, together with the drift vectors p_i . Thus f, g will amplify S_1 , will eliminate all other tides (but only the orthogonal component of O_1) and the drift model.

We obtain, through the MLS solution (29) and the matrix \mathbf{F} the filters (f_1, g_1) , (f_2, g_2) and (f_3, g_3) which can be used to eliminate the drift and separate the tidal species. They can be used for the separate determination of the D, SD and TD tides respectively. In this relation we shall call these filters D, SD and TD filters.

Through (30) we also get filters $f_j, g_j, j > 3$, which can be used for the determination of the species of higher frequencies.

In the application of NSV all elements (24) of the filters can easily be changed. It is even possible to make them amplify and eliminate selected non-tidal frequencies.

In Figure 4 is shown that the default filters (27) of NSV do not considerably deviate from the FFIL.

Very important characteristic of a filter is the quantity which we shall call RSTN (signal-to-noise ratio), computed through

$$(31) \quad \text{RSTN} = (2/n) \text{Var}(y(t)) / \text{Var}(h)$$

where h is a filtered signal of power 1, amplified by factor 1. The variances are computed theoretically, under the assumption of a WN.

If NSV is reasonably applied, we have

$$(32) \quad \begin{aligned} \text{for the filters in NSV:} \quad & \text{RTSN} \approx 0.75, 0.80 \\ \text{for the optimum Fourier filters:} \quad & \text{RTSN} = 1 \end{aligned}$$

We loose about 20-25% of precision and information which is the price of the elimination of the drift.

A high RSTN is of crucial importance. It is a guarantee that we remain close to the theoretically motivated scheme in (Venedikov, 1964b) and the FFIL and that we can process the $u(T)$ and $v(T)$ instead of $y(t)$.

Therefore NSV cannot use high pass filters without a separation of the main species. Such filters have much lower RSTN than pass band filters like our filters. For example, the filter of Pertsev, applied as a high pass filter by Chojnicki and ETERNA, has a very low RSTN = 0.06. Due to this these methods are obliged to apply a moving filtration with a shift $s = n$.

It should be noted that this filter remains an excellent device for the estimation of the drift, as a low-pass filter. Then it is also a pass band filter with a very high RSTN = 0.81.

In Figure 5 is shown the response of some of the filters of NSV. Compared to FFIL, we have a better elimination of the low frequencies (the drift) and better separation of the main species. The FFIL eliminate the non-tidal frequencies 7.5, 22.5 and 37.5. Our filters do not eliminate them because there has never been evidences for particular power concentrated at these domains of the spectrum.

I.6. Processing of the filtered numbers.

The second stage of the analysis consists in applying the method of the least squares on the filtered numbers. This means that we have to create the observation equations for $u(T_i) = f^* y_i$ and $v(T_i) = g^* y_i$ where f, g is one of the pairs of filters in (30), i.e. in $\mathbf{F} = \mathbf{C}(\mathbf{C}^* \mathbf{C})^{-1}$.

These filters eliminate (more or less well) the drift. Therefore, in creating the equations we have to take into account only the tidal component. We have to use the model considered in Section I.2, namely (6), (7) and (8). Obviously, we have to take into account the effect of the filters which an amplification of the tides by the factor (21). Also g is changing all phases by $\pi/2$, while f keeps the phases the same.

Having in mind all this we get, the observation equations

$$(33) \quad \begin{aligned} \sum_{j=1}^{\mu} [c_{jf}(T_i)\xi_j + s_{jf}(T_i)\eta_j] &= u(T_i) \\ \sum_{j=1}^{\mu} [-s_{jg}(T_i)\xi_j + c_{jg}(T_i)\eta_j] &= v(T_i), i = 1, 2, \dots, N \end{aligned}$$

where N is the number of the I -s and the coefficients at the unknowns ξ and η are

$$(34) \quad \begin{aligned} c_{jl}(T) &= \sum_{\omega \in \Omega_j} \rho_l(\omega) h_\omega \cos(\varphi_\omega + \omega T), \\ s_{jl}(T) &= \sum_{\omega \in \Omega_j} \rho_l(\omega) h_\omega \sin(\varphi_\omega + \omega T), \end{aligned}$$

where $l = f$ or g corresponding to the filters f and g,
 $\rho_f(\omega)$ is response of the even filter f,
 $\rho_g(\omega)$ is response of the odd filter g.

Since ω is multiplied by T and the shift between the intervals is $s = n$ (if there are not gaps), it seems that ω can be considered as $s\omega$, e.g. 48ω , with heavy aliasing problems. Actually, the coefficients remain functions of the initial ω through $\rho(\omega)$ and they can easily be solved. All suspected aliases are either eliminated through low $\rho(\omega)$ or they can be taken into account in the equations, if it is necessary.

After the equations (33) are created, it is a well known MLS procedure to solve them, to obtain the estimates of the unknowns ξ and η , then δ and κ , accompanied, of course, by estimates of their variances, i.e. of the precision. For the experimentation of different variant of the analysis, e.g. using different grouping of the tides, NSV provides the criterion of AIC of Akaike (Sakamoto, et al., 1986). The use of AIC is very simple: among several variants, the best one is the variant with the lowest AIC.

We recall that (33) are equations for one of the tidal species. We have such equations created for all species we want to determine, which are solved in parallel, but independently. This allows to get estimates of the variances different for every species.

Generally, the estimation of the precision through the filtered numbers, which is thus frequency dependant, is one of the most important features of NSV.

II. Application of the program NSV.

NSV has been developed using the ALPHA computer of the Computing Center of the University Complutense of Madrid. Available are the source routines which have to be compiled and linked. If another computer is used, small modifications are necessary, at least the subroutines for the date and time determination should be replaced.

Also available is a PC version of the program which can be directly used on PC of type 486 and 586. Available is a package with the .EXE of NSV, accompanied by some permanent data files, necessary for the execution, as well as by a small data bank, prepared for processing, according to the formats used by NSV.

The execution is made using the command "NS". NSV needs a series of files accompanying the program and, of course, some tidal data files.

II.1 Data files.

A data file is composed by hourly ordinates or observations. It can be organised in the well known and widely used, in the Earth tidal domain, international format, like Example 1

Example 1. A data file.

<u>date DT</u>	<u>12 consecutive ordinates</u>											
YYMMDDH	Y1	Y2	Y3	Y4	Y5	Y6	Y7	Y8	Y9	Y10	Y11	Y12
[13] 8710162	2006	4134	5738	6625	6738	6173	5157	3984	2938	2216	1913	1970
[13] 8710170	2193	2294	2027	1243	-63	-728	-1213	-3780	-5435	-5146	-3899	-1846
[13] 8710172	592	3034	5003	6156	6353	5660	4333	2757	1341	432	210	666
[13] 0	(a blank line or line with YY = 0 means a gap)											
[13] 8710202	362	2642	5026	6854	7502	6834	4854	2028	-1033	-3640	-5206	-5346
.....												

Here [13] are 13 columns which are not used by NSV so that in [13] can be written any 13 symbols, e.g. 13 blanks.

In Example 1 is shown how a gap is indicated. Notice:

- (i) One zero or blank line is enough to indicate a gap of any length. Several blank lines have the same effect.
- (ii) If there is not a gap but a jump between two lines which is not corrected, it should be indicated as a gap between these lines, although there is not a gap.

In this format 5 columns or digits are reserved for one ordinate. NSV allows to use larger numbers of columns, namely 6, 7, 8 or 9 columns per ordinate. The number of the columns is defined by the input number NCOL, as shown in Section II.3.

The ordinates should be integer numbers. Whatever are the data, the unit of an ordinate should be 0.1 of the usual unit, e.g. 0.1 µgal. Whatever unit is used, the amplitudes provided by the analysis are in unit 10 times greater, e.g. data in 0.1 mbars - amplitudes in mbars. However, one should take care the δ factors to be of the order of 1.

The effect of the units can be controlled using a multiplying constant SCALE (see Section II.3) as well as a multiplying constant (Sections II.2,4). If the unit is 0.1, as said above, SCALE = 0.1. If, unit 0.001 is used, SCALE = 0.001. For unit 1, SCALE = 1, etc.

II.2. Control file (CFILE).

Every data file should be accompanied by a control file (CFILE). It is convenient but not necessary both files to have one and the same name, but different extensions, e.g. BRU3M.DAT with a CFILE BRU3M.TIT.

A CFILE can have the form exactly the same as the CFILEs used in ICET and the program of Ducarme (1975). It is, however recommendable, to use CFILE in the particular form of NSV which is more comprehensive and flexible, and is allowing to give more information. NSV is automatically able to distinguish which form is used.

Example 2. A control file (CFILE).

```
!NAME_STATION      3      (follow 3 lines with name of the station)
STATION 0201 BRUXELLES-UCCLE  COMPOSANTE VERTICALE  BELGIQUE
50 47 55 N        04 21 29 E          H 101 M
```

GRAVIMETRE A SUPRACONDUCTIVITE GWR T3
!COMPONENT Gravimeter
!LONGITUDE_WEST_DEGREES -4.3581
!LATITUDE_NORTH_DEGREES 50.7986
!ALTITUDE_KILOM. 0.100
!GRAVITY_GALS 981.0
*!AZIMUTH_FROM_N_CLOCKWISE 0.0
!MULTIPLYING_CONSTANT 0.9907
!TIME_CORRECTION_SECONDS -30 seconds, negative = retardation.
!NO_INERT_CORRECTION_PARIISKII
!END

A CFILE consists, as shown in Example 2, of some control words (CW) which have a "!" in the first column. In many cases, CW is followed by a corresponding information, distant from the CW by at least one blank. A CW can be abbreviated by the first letters, at least the first 3, e.g. !COMP instead of !COMPONENT.

The !NAME, !COMP, !LONG and !LAT are obligatory, while all other CW are optional. With the exception of !NAME, which is at the first place, the CW can be given in any order. Letters upper or lower case can be used. One can add any number of rows with comments, which should not begin by an ! in column 1. Thus *!AZIMUTH in Example 2 is like a comment. Comments can be written before or after the numerical data but separated by at least one blank, like (follow 3 lines with...). The last CW should be !END.

The CFILE in Example 3, with the exception of the shorter name, is equivalent to the CFILE in Example 2.

Example 3. CFILE equivalent to Example 1 (without !NAME).

!NAME 1 (follows only 1 line)
STATION 0201 BRUXELLES-UCCLE COMPOSANTE VERTICALE BELGIQUE
data, rearranged in comparison with Example 1:
!Lat 50.7986 degrees, north
!Long -4.3581 degrees, positive to the West
!altitude 0.100 kilometers
!Grav approximately 981.0 gals
!COMPONENT Grav here only the first letter G is important
*!AZIMUTH for gravity data is not necessary
!Mult 0.9907
!TIME corr. -30 seconds, negative = retardation.
!NO_IN means NO_INERT._CORRECTION_PARIISKII
!END

The information used after !COMP is only the first letter of the first word. It is possible to use (only the letters upper case are important):

Gravimeter or Vertical component;
Horizontal component, Tiltmeter or Clinometer
Extenzometer horizontal
Strainmeter vertical
Ocean data, Air-pressure or temperature or Meteo data.

The cases O, A or M are processed using the theoretical tides of the static ocean tides.

There are a few more CW, some of which are:

!SEPARATED_POTENTIAL_3	see 2.6	(SEPP3)
!ADD_LUNISOL_WAVES_TAMURA	see 1.1.1.	(LUNISOL)
!ADD_VENUS/JUPITER_WAVES_TAMURA	see 1.1.2.	(VEN/JUP)

<code>!THRESHOLD_AMPLITUDE</code>	0.1	see 1.1.4.	(MIN.AMPL)
<code>!NO_CORRECTION_POTENTIAL_3</code>		see 1.2.1.	(NOCP3)
<code>!LENGTH_INTERVALS</code>	36	see 3.1.	(NFINT)
<code>!POWER_DRIFT_POLYNOMIALS</code>	1	see 3.3.	(KELIM)
<code>!PRINT_ONLY_4_DECIMALS</code>		see 6.1.	(OUT4)
<code>!DELTA_KAPPA_CORRECTIONS</code>		see 6.3.	(DKCORR)

A number is added after the CW which need a numerical information.

The effect of all these CW can be obtained using corresponding options in a dialogue (Section II.4), indicated here through "see ... ". In the dialogue can be used a help, explaining the meaning of the options. E.g., if the question

6.3. (DKCORR) DELTA AND KAPPA CORRECTIONS ? h

gets the answer h, as above, the user will obtain the information how to create the input which should follow `!DELTA_KAPPA_CORRECTIONS` or `!DEL`.

II.3. Data bank.

It is supposed that the user disposes by several data files. It is also expected that these data files can be subject to multiple processing, i.e. they have to be at disposition for a while, if not for ever. Therefore NSV is organising a simple information system.

It is necessary to write the names of the data in one or more OPEN files. In Example 4 is given an OPEN file called OPENA.DAT, which is accompanying the PC version of NSV.

Example 4. File OPENA.DAT with names of the files at disposition

Data, prepared as example for the use of the program NSV:

BRU3MP.TIT
BRU3MP.DAT
5 0.1
AIR-PRESSURE, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92

BRU3M.TIT
BRU3M.DAT
5 0.1
SUPRA, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92

BRU3MT.TIT
BRU3MT.DAT
5 0.1
BRUSSELS, TEMPERATURE, 3.8 MONTHS, 01.01.92-30.04.92

DUMMY.TIT
DUMMY.DAT
5 0.1
Names of data which do not exist

In the first line should be written a common name of the data, like "Data prepared as ..." in Example 4. Then follow, separated by a blank line, the names of every file in 4 lines. Their content is:

Line 1: Name of the CFILE, e.g. BRU3M.TIT
Line 2: Name of the data file, e.g. BRU3M.DAT
Line 3: The numbers: NCOL and SCALE, e.g. 5 0.1 (Section II.1)
Line 4: A short name of the data, e.g. SUPRA, BRUSSELS, 3 MONTHS,...

Up to 20 data files can be described in OPENA.DAT. There can also be included dummy names.

It is possible to organise an arbitrary numbers of OPEN files, e.g. OPENA.DAT, OPENB.DAT, OPENC.DAT ...,containing different sets of data. However, if a cross-regression will be used between some files, all of them should be written in one and the same OPEN file.

The names of the OPEN files should be written in the existing file OPENOPEN.DAT, e.g.

Example 5. OPENOPEN.DAT with the names of the OPEN files.

*This is the content of OPENOPEN.DAT

*(lines beginning by * are comments)

OPENA.DAT

OPENB.DAT

OPENC.DAT

*This is the end of the content

II.4. Start of the program NSV and a simple variant of analysis.

The work of the PC version of NSV is initiated by the command NS. At the beginning we get a message like

START OF THE PROGRAM NSV/1996 ON 25.06.1996 10h 26m 11.4sec
WITH WISHES FOR GOOD RESULTS FROM
VENEDIKOV, VIEIRA, DE TORO & ARNOSO FROM MADRID

Afterwards NSV is displaying the information from the OPEN files. If OPENA.DAT in Example 4 is used, this information will be:

DATA AT DISPOSITION:

FILE 11 BRU3MP.DAT	AIR-PRESSURE, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92
FILE 12 BRU3M.DAT	SUPRA, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92
FILE 13 BRU3MT.DAT	BRUSSELS, TEMPERATURE, 3.8 MONTHS, 01.01.92-30.04.92
FILE 14 DUMMY.DAT	Names of data which do not exist

Follows a message to choose the number of the data file to be processed. In this case it can be an integer between 11 and 13.

ENTER NR OF THE DATA FILE TO BE PROCESSED :

12

where "12" is the answer of the user.

If there is no answer (return key), NSV will display the information from the next OPEN file, e.g. OPENB.DAT, then the next one, OPENC.DAT, as many OPEN files as many names are given in OPENOPEN.DAT.

After the answer "12" follows a series of messages which do not need the intervention of the user.

The dialogue continues by getting the question:

0.0. SOME OPTIONS ?

If the answer is a blank (return), NSV proceeds the analysis using default values and options.

II.5. Options and questions.

NSV has a set of options. They are selected in a dialogue, during which NSV set up questions which have to be answered. The initial question has been given in Section II.4. If the answer is H or h (help), all questions are displayed. In Table 1 are given most of the questions. A few questions related with details of some options are not shown.

The answer of every question, written in the same line as the question, can be:

Y or y - accepted option,

H or h - NSV will provide explanations about the option,

Any other single letter is equivalent to Y or y

2 letters, e.g. CC, - brakes the dialogue, start of the processing.

blank or RETURN key - rejected option. In the following this answer will be denoted RET.

The underlined questions in Table 1 are "main" questions. If the answer is RET, the control goes to the next main question. Also, if the answer is RET to 1.1. the next question is 1.2. If the answer is RET to 1.2. the next question is 2.0.

In all other cases RET makes appear the next question.

In any case after 4.0. follows 5.0. If the answer to 4.0. is "y", then 4.1, 4.2,... are displayed, but later, during the execution.

Under some conditions, some question are not displayed. If 5.1. or 5.2. are used, 4.1. is not displayed. If 4.1. is used, the execution is interrupted after terminating its function, thus 4.2., 4.3,... are not displayed.

Example 6. A dialogue with NSV:

```
0.0. SOME OPTIONS ? x
1.0. (CORR) SOME CORRECTIONS AND CHANGES ?
2.0. (ANAL) CHANGES IN THE ANALYSIS ?
3.0. (FIL) CHANGES IN THE FILTERS ? x
3.1. (NFINT) NEW LENGTH n OF THE FILTERED INTERVALS ? x
ENTER n (EVEN NUMBER.LE.800): 44
3.2. (SHIFT) NEW SHIFT s OF THE FILTERED INTERVALS ? x
ENTER s (ANY INTEGER): 60
3.3. (KELIM) NEW POWER k OF THE DRIFT POLYNOMIALS ? x
ENTER THE VALUE k 0
3.4. (TIDEF) NEW TIDAL CONSTITUENTS OF THE FILTERS ? ss
END OF THE CONTROL INPUT AND
START CONSTRUCTION OF THE FILTERS AT 12h 34m 18.8sec CPU TIME: 27.85
END CONSTRUCTION OF THE FILTERS AT 12h 34m 20.2sec CPU TIME: 29.22
START OF THE DATA PROCESSING, CPU TIME SET TO 0
```

In Example 6 "x" (meaning y or yes), "ss" (meaning start of the processing) and the numbers "44" (for n), "60" (for s) and "0" (for k) are input given by the user. The questions 1.0. and 2.0. have the negative answers RET. Therefore 1.1., 1.2.,..., 2.1., 2.2.,... are skipped.

The effect of this dialogue is that NSV will use filters of length $n = 44$, shifted by $s = 60$, i.e. omitting 16 ordinates between the intervals. The filter will eliminate power $k = 0$, i.e. only an arbitrary constant, different in every interval.

The processing has started after the answer "ss" with the message

END OF THE . . .

II.6. Short comments of some options.

We think that the best and easiest way to understand the options is to use NSV, the explanations which can be displayed after an H or h and, most efficient, by making experiments. Therefore in the following we shall briefly concern only few options, which may be of greater interest.

1.1.3. (ADDW) ADDITIONAL WAVES ?

ADDW makes possible to deal with some waves which are not included in the potential development. They can be added to one of the existing species, D, SD,... or they can form a new species called AW. The waves can be defined either by a frequency (angular) or by an argument number of Doodson.

The option may be interesting for detection of particular tidal or non-tidal phenomena, as well as for including some particular waves which may exist in the ocean tidal phenomena.

2.1. (LONG) DETERMINATION OF THE LONG PERIOD (LP) WAVES ?

It is possible to determine LP using large filters of default length $n = 360$ which, however, can be optionally modified.

A new option of SV is to determine LP in parallel with the D (option named LONG(D)) or SD (option named LONG(S)) species, using the useful filters of length of the order of $n = 48$. This option may be applied to obtain the drift free of LP. This will be shown later in an example.

2.2. (MAIN) MAIN TIDAL SPECIES TO BE DETERMINED, CHOOSE D, SD,... ?

The default option of NSV is to determine the D, SD and TD waves. However, it is possible to determine also the species QD (quarterdiurnal, 4 cpd), FD (fifthdiurnal, 5 cpd) and D6 (sixthdiurnal, 6 cpd), as well as species defined by 1.1.3. (ADDW)

2.3. (WGHT) WEIGHTS OF THE FILTERED NUMBERS ?

The $u(T)$, $v(T)$ can be weighed by the estimated variances through the solution of (28). The option can be efficient in dealing with records having strong perturbations. If the use of this option improve considerably the results, it is possible that in the data are gross errors.

2.5. (ADDGR) ADDITIONAL TIDAL GROUPS FOR ELIMINATION OF A LEAKAGE ?

Since nothing is perfect, the separation of the main species by the filters is also not perfect. This may have an effect only for data of very high precision, e.g. theoretical model data. This option can strongly reduce the leakage.

4.0. (TVAR) TIME VARIATIONS, RESIDUALS ?

This set of options can be very useful for the detection of various anomalies in the data. Since such anomalies may be potential volcano and earthquake precursors, (TVAR) can be helpful for these fundamental problems.

Here we shall concern only some of these options.

4.1. (RESIN) RESIDUALS AND DRIFT IN THE FILTERED INTERVALS ?

RESIN deal with the data separately in every interval I_i , using the expressions (28) and (29)

$$(35) \quad \mathbf{C}\mathbf{x}_i = \mathbf{y}_i \quad \text{and} \quad \tilde{\mathbf{x}}_i = (\mathbf{C}^* \mathbf{C})^{-1} \mathbf{C}^* \mathbf{y}_i = \mathbf{F}^* \mathbf{y}_i$$

where \mathbf{F} is the matrix of the filters. In (30) are indicated only the filters which are used for the tidal analysis. Here, in (35), take part all filters or vectors of the matrix \mathbf{F} .

The hourly residuals provided by RESIN are computed through

$$(36) \quad \mathbf{r}_i = \mathbf{y}_i - \tilde{\mathbf{x}}_i.$$

These residuals are usually too small. However, big anomalies in single ordinates, e.g. gross errors, can be determined.

Some of the elements of the vector $\tilde{\mathbf{x}}_i$ are the estimates $\tilde{\zeta}_{il}$ of the drift unknowns ζ_{il} in (10) and (14). Using $\tilde{\zeta}_{il}$ in (10), RESIN is computing the drift for all points in the intervals.

RESIN is not used, when TORO1 or TORO2 are applied.

If RESIN is used, the execution is terminated after its application.

4.2. (RESAN) HOURLY RESIDUALS AND DRIFT AFTER THE ANALYSIS ?

RESAN provides hourly residuals, as well as the drift d_t in two variants: filtered d_t and residual d_t , called also drift + noise. The residual d_t is obtained as the difference between the observations $y(t)$ and the computed or adjusted tidal signal.

The filtered d_t is obtained through a moving filtration. The filter is not indicated in (30) but it is there. This is the first filter in the matrix \mathbf{F} , corresponding to the vector \mathbf{p}_0 . The filtered drift is actually the estimate $\tilde{\zeta}_0$ of the unknown ζ_0 in (10) and (14), obtained hour by hour.

The filtered drift is a smoothed drift, its values d_t being strongly correlated. In our opinion, more information is provided by the residual drift.

The hourly residuals are obtained as the difference between $y(t)$ and (estimated tidal signal + filtered d_t).

About other options related with the question 4.0. the user can find the information using the help option as well as through experimentation.

5.0. (CROSS) EFFECT OTHER SERIES (INPUT CHANNELS) OF DATA ?

Very essential option through which the component a_i in (2) is modelled and eliminated (approximately, of course).

The use of this option depends on the questions:

5.1. (TORO1) REGRESSION MODEL1 WITHOUT PHASE SHIFT ?

5.1. (TORO2) REGRESSION MODEL2 WITH PHASE SHIFT ?

In principle the idea of De Meyer (1982) is used but through our model (Simon et al., 1989, Toro et al., 1991, 1993a). Now this is done in the following way.

Let FILE1 are some tidal data, while FILE2 are data of a phenomenon which is perturbing FILE1. E.g. FILE1 are the gravity data BRU3M.DAT and FILE2 are the air-pressure data BRU3MP.DAT.

Let $u(T)$ and $v(T)$ are filtered numbers obtained from FILE1, $p(T)$ and $q(T)$ are corresponding filtered numbers obtained from FILE2, \tilde{p} and \tilde{q} are the mean values of $p(T)$ and $q(T)$ and

$$(37) \quad \begin{aligned} \Delta p(T) &= p(T) - \tilde{p}, \\ \Delta q(T) &= q(T) - \tilde{q} \end{aligned}$$

If MODEL 2 (TORO2) is accepted, the equations (33) are modified into

$$(38) \quad \begin{aligned} \sum_{j=1}^{\mu} [c_{jf}(T)\xi_j + s_{jf}(T)\eta_j] + b_1\Delta p(T) - b_2\Delta q(T) &= u(T) \\ \sum_{j=1}^{\mu} [-s_{jg}(T)\xi_j + c_{jg}(T)\eta_j] + b_1\Delta q(T) - b_2\Delta p(T) &= v(T) \end{aligned}$$

Here b_1 and b_2 are regression coefficients which can be represented as

$$(39) \quad b_1 = b \cos \beta, \quad b_2 = b \sin \beta, \quad \text{where}$$

b is a coefficient of proportionality and

β is a phase lag of the effect of FILE2 on FILE1

We get b_1 and b_2 , b and β separately for the main species, i.e. NSV is looking for frequency dependent relations.

If MODEL 1 is applied, it is accepted that $\beta \equiv 0$ and we get only one regression coefficient $b_1 = b$.

It is possible to deal at once with 6 files like FILE2, i.e. in (38) to put 6 pairs of unknowns b_1 and b_2 . However, it is very unlikely to be successful if more than two files FILE2 are considered.

5.3. (CROSSC) CORRECTING CROSS REGRESSION COEFFICIENTS ?

If neither 5.1 nor 5.2 is used, CROSSC can correct the filtered number using given regression coefficient b , common for all frequencies. The corrections to $u(T)$ and $v(T)$ are $bp(T)$ and $bq(T)$ respectively. Notice that here are used $p(T)$ and $q(T)$, instead of $\Delta p(T)$ and $\Delta q(T)$.

II.7. A few examples.

7.1. Options SLOW and TORO1.

Table 2 is a copy of the output provided by NSV. B1 are values of the coefficients b_1 in (9). We have got B1 frequency and time depending.

We have an application of SLOW with intervals of length 30 days and shifted by 30 days. The last interval has 46 days because there are left 16 days of data which cannot shape an interval of the selected length.

The output using this option as well as the output of all options for time variations is in a form suitable to use a plotting program. Therefore the dates are written in a not very convenient form.

In addition to this output we get the values of δ and κ for selected tides for the same time intervals and similar format.

7.2. Options LONG, TORO1, TORO2 and VGR.

In Table 3 we have variants of the results of the analysis about the long period waves. In the title is given part of the output when only one of the variants is applied.

When VGR is not applied, the tidal group called MF unify all LP tides. However, the filters which amplifies MF, are retaining or eliminating the longer tides, e.g. SSA and SA.

When VGR is applied, we have the variants,

VGR(MTM): the LP tides are separated in groups MF and MTM

VGR(MTM,MSOM): the LP tides are separated in groups MF, MTM and MSOM.

However, in Table 3 are given only the results for MF.

This example has needed a preparation of the file NGROUPS.DAT, where the groups used MF, MTM and MSOM are defined through intervals of the frequency ω .

Table 3 is also an example how the criterion AIC can be used. We have an important reduction of AIC at the first step LONG \Rightarrow (LONG,TORO1). Further we have not any improvement, even AIC is slightly raising. We have to remain at the lowest value, i.e. to accept the result (LONG,TORO1).

7.3. Option LONG(D), TORO1 and RESAN.

Let a filtered or residual d_t is determined using RESAN. In both cases, if CROSS is used, taking into account the air-pressure (AP), the estimated effect is subtracted from d_t .

If CROSS is not used, d_t remains charged by the effect of AP.

Also, if LONG is used, but only the variants LONG(D) or LONG(S), the estimated LP are subtracted from d_t .

In Figures 6, 7 and 8 are given examples of the determination of the drift.

More evident are the effects at the data of the superconducting gravimeter in Figure 6. RESAN provides a drift charged by the effect of the air-pressure where the most important signal is the LP tides. One can very distinguish the monthly and halfmonthly period.

The application of TORO1 eliminates quite well the effect of the air-pressure and the manifestation of the LP waves becomes completely clear.

At the end, through LONG(D) the LP are determined and thus eliminated. We get a drift in which we can look for some geophysical phenomena.

In Figures 7 and 8 the procedures are repeated for ocean data. We have similar effects as in Figure 6, but not so strong.

In Figure 6 we use a filtered d . We get data looking less noisy but, due to the smoothing, there are some small waves.

7.4. Options RESFN and TORO1.

In Figure 9 are given the D residuals of the filtered numbers, obtained using RESFN. The same residuals are computed after TORO1 is proceeding a regression on air pressure data. Obviously, there is a considerable reduction of the magnitude of the residuals.

However, the effect is not so strong for the SD residuals.

7.5. Option SLOW.

In Figure 10 is shown an example of the study of the slow time variation of the tidal parameters. SLOW can represent the δ factors of one wave as a function of the δ factor of another wave. If there is a correlation, this can be explained by a variation of the sensibility. This effect is well demonstrated in Figure 10.

7.6. Estimation of the precision by NSV using model data of the noise.

In principle, the reasonable estimation of the noise made by NSV is one of its most important properties.

Let V is the covariance matrix of the data $y(t)$ and $V = T^*T$, where T is an upper triangular matrix, and let y denotes the vector whose elements are all $y(t)$. According to the theory, we would have a correct application of MLS, if we work with the transformed data $u = T^*y$, instead of y . The reason of the transformation is that the noise of u becomes WN.

Since V is not available, NSV applies the transformation of y in our filtered numbers u and v . This is done under the supposition that u and v are also charged by a WN.

In Figure 11 is represented a simulated white noise. Its spectrum is representing the effect of the noise in the frequency domain. About the spectrum is used such a scale, that the level of the noise should be of the same order as the standard (the square root of the variance) of the simulated noise.

The spectrum obtained is according to the theoretical level of the noise which is the constant $L_\omega = \sigma = 1$ over the whole frequency domain.

In Figure 12 are shown the D and SD filtered numbers u . They obviously behave as a WN. NSV is applying filters, normalised in such a way that in the case of WN, u and v should have the same variance as the data. Indeed, the level of the spectrum of u in Figure 12 is corresponding to $\sigma = 1$.

In Figure 13 is shown a simulated stationary noise obtained as an MA (moving average) of the white noise. Unlike the WN, the spectrum is showing an accumulation of the energy at the D and SD domains, corresponding to the theoretically determined level of the noise. We have, more or less, similar picture of the real noise of the tidal data, mainly due to the meteorological effects in which are some variable D and SD waves.

As shown in Figure 14, the filtered numbers behave as a WN because the spectrum is maintaining a constant level. However, we have a higher level for the D filtered numbers and lower level for the SD filtered numbers, which is realistically corresponding to the spectrum of Figure 13.

In Table 4 are given the results of the estimation of the noise through ETERNA and NSV, using different filters.

7.6. Analysis results.

Table 5 is a copy, with very few modifications, of the analysis of the data file 12, BRU3M.DAT. There is applied a cross-regression with the air-pressure data, BRU3MP.DAT.

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REFERENCES

- (BIM = Bulletin d'Information des Marees Terrestres, editeur P.Melchior)
- Baker, T.F., 1978a: A review of the objectives of tidal analysis. BIM 78, 4571-4578.
- Baker, T.F., 1978b: Non-equilibrium influences on the tidal signal. BIM 78 4 4596-4610.0.
- Carvajal, J.M., 1993: Modelización de las mareas oceánicas y estudio de las variaciones del nivel medio del mar en los puertos de Gijón y Santander, Tesis doctoral, Universidad de Oviedo.
- Chojnicki, T., 1973: Ein verfahren zur Erdzeitenanalyse in Amlehnung an das Prinzip der kleinsten Quadrate, Mitt. Inst. Theoretische Geodesie Univ. Bonn, Nr 15.
- Chojnicki, T., 1978: Supplementary precision estimation of results of tidal data adjustment, BIM 78, 4670-4675
- Ducarme, B., 1975: The computation procedures at the International Center for Earth tides (ICET), BIM 72, 4156-4181.
- Fernández, J., Vieira, R., Venedikov, A.P., Díez, J.L., 1993: Vigilancias del riesgo volcánico en Canarias. Isla de Lanzarote. Física de la Tierra, Num. 5, Editorial Complutense, Madrid, 77-88.
- Horn, W., 1959: The harmonic analysis, according to the least square rule, of tide observations upon which an unknown drift is superposed, Third Int. Symp. Earth Tides, Trieste, p. 218.
- Ishiguro, M., Akaike, H., Ooe, M., & Nakai, S., 1983: A Bayesian approach to the analysis of Earth tides, Proc. 9th Int. Symp. Earth Tides, New York 1981, Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, 283-292.
- Melchior, P., sec. edition 1981: The tides of the planet Earth, Pergamon press, Oxford, Chapter 7, 161-191.
- Melchior, P., Venedikov, A.P., 1968: Derivation of the wave M3 (8^h 279) from the periodic tidal deformations of the Earth. Physics Earth Plan. Interiors, vol. 1, 363-372.
- Meyer, F. De, 1980: A study of various harmonic analysis methods for earth tides observations. BIM 83, 5187- 5235.
- Meyer, F. De, 1982: A multi input - single output model for earth tide data. BIM 88, 5628- 5674.
- Munk, W., Cartwright, D., 1966: Tidal spectroscopy and prediction. Phil. Trans. Roy. Soc., Ser.A 259, 533-581.
- Sakamoto, Y., Ishiguro, M., Kitagawa, G., 1986: Akaike Information Criterion statistics, D. Reidel Publ. Company, KTK Scientific Publishers/Tokyo, Chapter 8, p. 172.

- Schüller, K., 1977: Tidal analysis by the Hybrid Least Squares frequency domain convolution method, 8th Int. Symp. Earth Tides, Bonn.
- Schüller, K., 1978: About the sensitivity of the Venedikov tidal parameter estimates to leakage effects. BIM 78, 4635-4648.
- Simon, Zd., Stanchev, Vl., Toro, C. de., Venedikov, A.P., Vieira, R., 1989: Relation between Earth tide observations and some other data. BIM 105, 7470-7478.
- Tamura, Y., 1987: A harmonic development of the tide-generating potential, BIM, Nr 99, pp. 6813-6855.
- Tamura, Y., Sato, T., Ooe, M., & Ishiguro, M., 1991: A procedure for tidal analysis with a Bayesian information criterion, Geophys. J. Int., 104, 507-516.
- Toro, C. de., Venedikov, A.P., Vieira, R., 1990: Determination of some particular waves in the Earth tide data, BIM 106, 7511-7521.
- Toro, C. de., Venedikov, A.P., Vieira, R., 1991: A particular study of the relation between Earth tide data and other time series. Bulgarian Geophysical Journal, v. XVII, No 1, 37-43.
- Toro C. de., Venedikov A.P., Vieira R., 1993a: A new method for Earth tide data analysis. Bulletin d'Information des Marees Terrestres, 116, 8557-8586.
- Toro, C. de., Venedikov, A. P., Vieira, R., 1993b: Study of the sensibility variations of the earth tide records. Física de la Tierra, Num. 5, Editorial Complutense, Madrid, 61-70.
- Venedikov, A.P., 1961: Application à l'analyse harmonique des observations des marées terrestres de la Méthode des moindres carrés. Comptes Rendues Académie Bulgare des Sciences, Vol.t14, No 7, 671-674, Sofia.
- Venedikov, A.P., 1964a: Sur l'estimation de la precision des observations des marées terrestres. BIM 36, 1223-1229.
- Venedikov, A.P., 1964b: Sur une possibilité d'application de la Méthode des moindres carrés à l'analyse des observations des marées terrestres. Cinquième Symp. Intern. Marées terrestres, Bruxelles, 1964, Comm. Obs. R. Belgique n° 236, S. Gwoph. n° 69, 412-420.
- Venedikov, A.P., 1966a: Une méthode d'analyse des Marées terrestres à partir d'enregistrements de longueur arbitraires, Acad. Royal de Belgique, Bull. Cl. Sci., 5e S., t. LIII, fasc. 3, Communications de l'Observatoire Royal de Belgique, Série géoph. No 71, 463-485.
- Venedikov, A.P., 1966b: Sur la constitution de filtres numériques pour le traitement des enregistrements des Marées terrestres. Acad. Royal de Belgique, Bull. Cl. Sci., 5e S., t. LIII, fasc. 6, Communications de l'Observatoire Royal de Belgique, série géoph. No 76, 827-845.
- Venedikov, A.P., 1979: Testing of the grouping of the waves in the analysis of the Earth tidal records, BIM 81, 4988-4998.
- Venedikov, A.P., 1981: Determination of the tidal parameters from short intervals in the analysis of Earth tidal records. BIM 85, 5435-5441.
- Venedikov, A.P., 1984: Analyse des enregistrements de marées terrestres, BIM 92, 6078-6120.
- Venedikov, A.P., 1986: Determination of the quarterdiurnal earth tidal waves (in Bulgarian). Bulgarian Geophysical Review, v. XIII, No 1, 69-78.
- Venedikov, A.P., Ducarme, B., 1979: Determination of the long period tidal waves. BIM 81, 5051-5059.
- Venedikov, A.P., Vieira, R., Toro, C. de, 1995: The computer program NSV used in Madrid for tidal data processing. BIM 121, 9108-9126.
- Venedikov, A.P., 1995: Remark about the MV66 and ETERNA 3.1 tidal analysis methods (a discussion). Bulletin d'Information des Marees Terrestres, 121, 9127-9138.

- Wenzel, H.-G., 1976a: Zur Genaugkeit von gravimetrischen Erdgezeiten-beobachtungen. Wissenschaftlichen Arbeiten der Lehrsttle für Geodasie, Photogravimetrie und Kartographie, Technischen Universität Hannover, 67.
- Wenzel, H.-G., 1976b: Some remarks to the analysis method of Chojnicki. BIMt73, 4187-4191.
- Wenzel, H.-G., 1977: Estimation of accuracy for the earth tide analysis results, BIM 76, 4427-4445.
- Wenzel, H.-G., 1994a: Earth tide analysis package ETERNA 3.0., BIM 118, 8719-8721
- Wenzel, H.-G., 1994b: ETERNA - an Earth tide analysis program package for the personal computer, preprint, submitted to Bulletin Géodesique, January 1994.
- Xi Quinwen, 1985: The algebraic deduction of harmonic development for the tide-generating potential with the IBM-PC, Proc. 10th Int. Symp. Earth Tides, 481-490.

Table 1. List of the questions/options of NSV.

- 1.0. (CORR) SOME CORRECTIONS AND CHANGES ?
- 1.1. (THEOW) THEORETICAL WAVES (TIDAL POT. DEVELOPMENT) ?
 - 1.1.1. (LUNISOL) ADDITIONAL LUNI-SOLAR TERMS OF TAMURA ?
 - 1.1.2. (VEN/JUP) TIDAL WAVES OF VENUS AND JUPITER ?
 - 1.1.3. (ADDW) ADDITIONAL WAVES ?
 - 1.1.4. (MINAMPL) IGNORE WAVES UNDER A THRESHOLD AMPLITUDE ?
 - 1.1.5. (PRDEV) PRINT THE TIDAL POTENTIAL DEVELOPMENT ?
 - 1.2. (DATA) DATA CORRECTIONS ?
 - 1.2.1. (NOCP3) NO CORRECTIONS FOR POTENTIAL OF ORDER 3 AND 4 ?
 - 1.2.2. (NOINC) NO CORRECTION PARIISKII (ONLY GRAVIMETERS) ?
 - 1.2.3. (SECCOR) TIME CORRECTION IN SECONDS ?
 - 1.2.4. (MULT) MULTIPLYING CORRECTION ?
 - 1.2.5. (CUT) RESTRICT (CUT OFF) THE INTERVAL OF THE DATA ?
 - 1.2.6. (ADJUST) ADJUST THE VOLUME OF THE DATA TO OTHER SERIES ?
- 2.0. (ANAL) CHANGES IN THE ANALYSIS ?
- 2.1. (LONG) DETERMINATION OF THE LONG PERIOD (LP) WAVES ?
 - 2.2. (MAIN) MAIN TIDAL SPECIES TO BE DETERMINED, CHOOSE D, SD,... ?
 - 2.3. (WGHT) WEIGHTS OF THE FILTERED NUMBERS ?
 - 2.4. (VGR) CHOOSE THE VARIANT OF GROUPING ?
 - 2.5. (ADDGR) ADDITIONAL TIDAL GROUPS FOR ELIMINATION OF A LEAKAGE ?
 - 2.6. (SEPP3) SEPARATE THE TIDES OF DEGREE 3 ?
- 3.0. (FIL) CHANGES IN THE FILTERS ?
- 3.1. (NFINT) NEW LENGTH n OF THE FILTERED INTERVALS ?
 - 3.2. (SHIFT) NEW SHIFT s OF THE FILTERED INTERVALS ?
 - 3.3. (KELIM) NEW POWER k OF THE DRIFT POLYNOMIALS ?
 - 3.4. (TIDEF) NEW TIDAL CONSTITUENTS OF THE FILTERS ?
 - 3.5. (RESPF) RESPONSE OF THE FILTERS TO SOME TIDES ?
 - 3.6. (SPECF) SPECTRUM OF THE FILTERS ?
- 4.0. (TVAR) TIME VARIATIONS, RESIDUALS ?
- 4.1. (RESIN) RESIDUALS AND DRIFT IN THE FILTERED INTERVALS ?
 - 4.2. (RESAN) HOURLY RESIDUALS AND DRIFT AFTER THE ANALYSIS ?
 - 4.3. (FAST) FAST AMPLITUDE AND PHASE VARIATIONS ?
 - 4.4. (RESFN) RESIDUALS OF THE FILTERED NUMBERS ?
 - 4.5. (FILN) PRINT THE FILTERED NUMBERS ?
 - 4.6. (SLOW) SLOW AMPLITUDE AND PHASE VARIATIONS ?

- 5.0. (CROSS) EFFECT OTHER SERIES (INPUT CHANNELS) OF DATA ?
- 5.1. (TORO1) REGRESSION MODEL 1 WITHOUT PHASE SHIFT ?
- 5.2. (TORO2) REGRESSION MODEL 2 WITH PHASE SHIFT ?
- 5.3. (CROSSC) CORRECTING CROSS REGRESSION COEFFICIENTS ?

- 6.0. (OTHER) OTHER OPTIONS ?
- 6.1. (OUT4) OUTPUT WITH ONLY 4 DECIMALS OF DELTA ?
- 6.2. (ARTW) ADD AN ARTIFICIAL WAVE ?
- 6.3. (DKCORR) DELTA AND KAPPA CORRECTIONS ?

Table 2. Output when the option SLOW and TORO1 are applied.

Data used: Station Brussels, Superconducting gravimeter.

TIME VARIATIONS OF THE REGRESSION COEFFICIENTS

NR INT	INTERVAL FROM	INTERVAL TILL	CENTRAL EPOCH DAYS	DATA USED DAYS	B1(D)	B1(SD)	B1(TD)	B1(QD)
1	86111500	86121423	15.0	30.0	-0.325	-0.421	-0.359	-0.075
2	86121500	87011323	45.0	30.0	-0.354	-0.351	-0.441	-0.237
3	87011400	87021223	75.0	30.0	-0.458	-0.214	-0.223	-0.075
4	87021300	87031423	105.0	30.0	-0.378	-0.463	-0.508	0.017
5	87031500	87041323	135.0	30.0	-0.377	-0.241	-0.428	-0.733
6	87041400	87051323	165.0	30.0	-0.351	-0.298	-0.692	0.048
7	87051400	87061223	195.0	30.0	-0.444	-0.395	-0.228	-0.310
8	87061300	87071223	225.0	30.0	-0.309	0.012	-0.099	-0.441
9	87071300	87081123	255.0	30.0	-0.344	-0.435	-0.228	-0.223
10	87081200	87091023	285.0	30.0	-0.313	-0.188	-0.183	-0.040
11	87091100	87101023	315.0	30.0	-0.342	-0.289	-0.298	-0.320
12	87101100	87112523	353.0	46.0	-0.350	-0.418	-0.330	-0.304

Table 3. Application of the options LONG, TORO1, TORO2 and VGR.

Data: Station Brussels, Superconducting gravimeter, 21.04.82-20.01.93 and air-pressure for the same interval.

FILTERED INTERVALS OF LENGTH 360 HOURS
 SHIFT (DISTANCE BETWEEN THE EPOCHS) OF THE INTERVALS 360 HOURS
 APPROXIMATION IN THE INTERVALS BY DRIFT POLYNOMIALS OF POWER 1
 AND TIDES: MF S1 O1 Q1 S2 M2 N2 S3
 DATA USED: 3927 DAYS, 92520 READINGS, 3 BLOCKS, 257 INTERVALS
 82.04.21.00/82.06.09.11 82.06.02.00/86.10.29.23 86.11.15.00/93.01.19.23

Options used	$\delta(MF)$	$\sigma(\delta)$	$\delta(Mf)$	$\sigma(\kappa)$	$\sigma(y)$	AIC
LONG	1.1463	0.0163	1.262	0.811	21.00	4594.
LONG, TORO1	1.1483	0.0046	0.007	0.231	5.98	3305.
LONG, TORO1, VGR(MTM)	1.1467	0.0047	0.016	0.234	5.96	3307.
LONG, TORO2	1.1482	0.0046	0.006	0.231	5.98	3308.
LONG, TORO2, VGR(MTM)	1.1466	0.0047	0.014	0.234	5.97	3309.
LONG, TORO1, VGR(MTM, MSOM)	1.1467	0.0047	0.009	0.234	5.97	3312.

Table 4. Estimation of the precision using models of the noise.

Estimation of the precision of the white noise in Figure 11

Tidal species		D	SD	TD	QD
Theoretical level of the noise		1.00	1.00	1.00	1.00
Programs	variants	Mean Square Deviations MSD			
ETERNA	ET49	1.20	1.07	0.99	1.20
	ET51	0.98	1.02	1.00	1.16
	ET239	1.03	1.07	0.96	1.18
NSV	V48	1.00	1.00	1.02	1.00
	V36	1.01	1.04	1.04	1.01
	V40	0.99	1.00	1.00	0.99
	V44	0.98	1.00	1.03	1.04
	V52	0.99	0.98	1.00	1.04

Estimation of the precision of the stationary noise in Figure 13

Tidal species		D	SD	TD	QD
Theoretical level of the noise		3.08-3.99	0.93-1.81	0.65-0.84	0.47-0.52,
Programs	variants	Mean Square Deviations MSD			
ETERNA	ET49	4.50	1.58	0.70	0.57
	ET51	3.68	1.49	0.71	0.55
	ET239	3.86	1.55	0.69	0.56
NSV	V48	3.30	1.45	0.77	0.51
	V36	2.90	1.10	0.64	0.47
	V40	3.13	1.45	0.73	0.47
	V44	3.25	1.43	0.79	0.52
	V52	3.35	1.52	0.73	0.47

ETn means filters of ETERNA of length n.

Vn means filters of NSV of length n.

Table 5. Example of the output when the gravity data BRU3M.DAT are processed. A cross-regression (TORO1) is applied, using the air-pressure data BRU3MP.DAT.

START OF THE PROGRAM NSV/1996 ON 28.06.1996 14h 16m 39.0sec
 WITH WISHES FOR GOOD RESULTS FROM
 VENEDIKOV, VIEIRA, DE TORO & ARNOSO FROM MADRID
 DATA AT DISPOSITION:
 FILE 11 BRU3MP.DAT AIR-PRESSURE, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92
 FILE 12 BRU3M.DAT SUPRA, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92
 FILE 13 BRU3MT.DAT BRUSSELS, TEMPERATURE, 3.8 MONTHS, 01.01.92-30.04.92
 FILE 14 DUMMY.DAT Names of data which do not exist
 THE DATA FILE TO BE PROCESSED IS
 FILE 12 BRU3M.DAT SUPRA, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92
 END READING DATA IN NDTF.FOR FROM FILE 12
 DATA INTERVAL: 92.01.01.00. - 92.04.01.12.
 NUMBER DATA 2208 NUMBER BLOCKS, 1
 START READING DATA FROM CONTROL INPUT FILE 72

INFORMATION FROM THE CONTROL DATA FILE:

```
!NAME 3
!Comp G
!Longitude_WEST_DEGREES -4.3581
!Latitude_North_Degrees 50.7986
!Altitude_Kilom. 0.100
!Gravity_Gals_Or_Zero 981.0
!Multiplying_Constant 0.9907
!Time -30
!No_Inert_Correction_Pariiskii
0.0. Some Options ? Y
5.0. (CROSS) EFFECT OTHER SERIES (INPUT CHANNELS) OF DATA ? Y
5.1. (TORO1) REGRESSION MODEL 1 WITHOUT PHASE SHIFT ? Y
ENTER NUMBER(s) OF FILE(s)=2nd, 3d...7th CHANNEL
IN ONE ROW : 11.
```

THE SECOND... DATA FILE TO BE PROCESSED IS
FILE 11 BRU3MP.DAT AIR-PRESSURE, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92

```
END READING DATA IN NDTF FOR FROM FILE 11
DATA INTERVAL: 92.01.01.00. - 92.04.01.12.
NUMBER DATA 2208 NUMBER BLOCKS, 1
END OF THE CONTROL INPUT AND
START CONSTRUCTION OF THE FILTERS AT 14h 17m 10.1sec CPU TIME: 31.08
END CONSTRUCTION OF THE FILTERS AT 14h 17m 11.6sec CPU TIME: 32.62
START OF THE DATA PROCESSING, CPU TIME SET TO 0
```

```
DATA FILE 12:
BRU3M.DAT
SUPRA, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92
PROCESSING THE DATA, BE SO KIND TO WAIT.....
END OF THE DIRECT PROCESSING, I.E. END FILTRATION,
CREATION OF MODEL EQUATIONS AND NORMAL EQUATIONS;
FOLLOWS: RESOLUTION OF THE NORMAL EQUATIONS AND OUTPUT
```

```
STATION 0201 BRUXELLES-UCCLE COMPOSANTE VERTICALE
50 47 55 N 04 21 29 E H 101 M P 4M D 90KM 981 117 301
GRAVIMETRE A SUPRACONDUCTIVITE GWR T3
LEAST SQUARE ANALYSIS OF INDEPENDENT FILTERED NUMBERS
(VENEDIKOV/MV66)
PROGRAM NSV, VERSION 20.06.1996 FOR PC 486/UNIVERSITY COMPLUTENSE
PROGRAMMING VENEDIKOV, VIEIRA, DE TORO, ARNOSO,
INSTITUTO DE ASTRONOMIA Y GEODESIA, C.S.I.C-U.C.M, MADRID
```

```
FILTERED INTERVALS OF LENGTH 48 HOURS
SHIFT (DISTANCE BETWEEN THE EPOCHS) OF THE INTERVALS 48 HOURS
APPROXIMATION IN THE INTERVALS BY DRIFT POLYNOMIALS OF POWER 2
AND TIDES: S1 O1 S2 M2 S3 S4 S5 S6
TIDAL POTENTIAL DEVELOPMENT OF TAMURA
THEORETICAL ELEMENTS OF THE TIDES COMPUTED AFTER TAMURA
TIME CORRECTION -30.0 SECONDS
ANALYSIS ON 28.06.1996, START: 14h 17m 11.6sec, END: 14h 17m 13.2sec
CPU TIME USED: 1.60 SEC. FOR DATA OF LENGTH .25 YEARS
DATA USED: 91 DAYS, 2208 READINGS, 1 BLOCKS, 46 INTERVALS
REGRESSION MODEL 1 OF DE TORO, VENEDIKOV, VIEIRA
CROSS-REGRESSION, ADDITIONAL CHANNELS (INDEPENDENT VARIABLES):
DATA FILE 11:
BRU3MP.DAT
```

AIR-PRESSURE, BRUSSELS, 3 MONTHS, 01.01.92-01.04.92
THEORETICAL MODULATION OF THE D AND SD WAVES OF ORDER 3
THE INERTIAL CORRECTION (PARIISKII) IS *NOT* INTRODUCED
THE DATA ARE MULTIPLIED BY .9907

92.01.01.00/92.04.01.23

WAVE GROUP ARGUM.	N	WAVE	ESTIMATED	AMPLIT.	PHASE		
			AMPL.	M.S.D.	FACTOR	M.S.D.	DIFF.
REGRESSION ON CHANNEL=FILE 11							
B1(D)			REGRESSION COEFFICIENT	= -.3336,	MS ERROR=	.0147	
1B9-13X	143	Q1	7.038	.017	1.14022	.00280	-.0972
142-149	58	O1	35.400	.019	1.14266	.00061	.0509
152-158	48	NO1	3.097	.017	1.13067	.00634	.6312
160-169	56	K1	51.497	.020	1.13213	.00044	.1838
171-177	40	J1	2.868	.016	1.14683	.00631	-.0848
181-1J3	105	OO1	1.918	.018	1.15473	.01076	-.2668
REGRESSION ON CHANNEL=FILE 11							
B1(SD)			REGRESSION COEFFICIENT	= -.2920,	MS ERROR=	.0378	
2A7-23X	99	2N2	1.395	.007	1.13470	.00534	4.1523
242-249	50	N2	6.764	.007	1.15039	.00125	3.0903
252-259	56	M2	35.267	.008	1.17388	.00027	2.7392
261-268	39	L2	1.175	.012	1.13667	.01201	5.4301
270-2J3	133	S2	17.302	.007	1.18950	.00048	1.3055
REGRESSION ON CHANNEL=FILE 11							
B1(TD)			REGRESSION COEFFICIENT	= -.3846,	MS ERROR=	.0499	
315-375	68	M3	.427	.005	1.08662	.01299	-.1729
382-3X5	14	S3	.010	.004	2.29255	.94942	-88.0543
STANDARD DEVIATIONS:							
	D		.549	SD	.225	TD	.147
AIC (AKAIKE):							
	D		179.	SD	11.	TD	-79.

END OF THE PROCESSING 28.06.1996, 14h 17m 13.3sec CPU TIME 1.71 SEC.

